GEOLOGINEN TUTKIMUSLAITOS

BULLETIN
DE LA
COMMISSION GÉOLOGIQUE
DE FINLANDE

N:o 168

SUOMEN GEOLOGISEN SEURAN JULKAISUJA
MEDDELANDEN FRÅN GEOLOGISKA SÄLLSKAPET I FINLAND
COMPTES RENDUS DE LA SOCIÉTÉ GÉOLOGIQUE DE FINLANDE

XXVIII

HELSINKI
MAI 1955

GEOLOGINEN TUTKIMUSLAITOS
BULLETIN DE LA COMMISSION GÉOLOGIQUE DE FINLANDE N:o 168

SUOMEN GEOLOGISEN SEURAN JULKAISUJA
MEDDELANDEN FRÅN GEOLOGISKA SÄLLSKAPET I FINLAND
COMPTES RENDUS DE LA SOCIÉTÉ GÉOLOGIQUE DE FINLANDE

XXVIII

HELSINKI
MAI 1955

Helsinki 1955. Valtioneuvoston kirjapaino

SISÄLLYSLUETTELO — CONTENTS

Sivu—Page

KIRJOITUKSIA — PAPERS

SISÄLLYSLUETTELO — CONTENTS

SUOMEN GEOLOGINEN SEURA
THE GEOLOGICAL SOCIETY OF FINLAND

JÄSENLUETTELO — 31. 12. 1954 — MEMBERSHIP LIST

Kunniapuheenjohtaja — Honorary President

Eskola, Pentti, Prof.	Kauppiaankatu 8—10 B, Helsinki

Kirjeenvaihtajajäsenet — Corresponding members

Bowen, Norman L., Dr.	Geophysical Laboratory, 2801 Upton Street, N.W., Washington 8, D.C., U.S.A.
Cloos, Ernst, Dr.	The Johns Hopkins University, Baltimore 18, Md., U.S.A.
von Eckermann, Harry, Prof.	Linnégatan 75, Stockholm, Sverige
Flint, Richard F., Prof.	Dept. of Geology, Yale University, New Haven, Conn., U.S.A.
Lougee, Richard J., Prof.	Clark University, 936 Main Street, Worcester 10, Mass., U.S.A.
Magnusson, Nils, Prof.	Sveriges Geologiska Undersökning, Stockholm 50, Sverige
Sander, Bruno, Prof. Dr.	Alte Universität, Innsbruck, Österreich
Wegmann, C. Eugen, Prof.	Neuchâtel, Suisse
Vendl, A., Prof.	Rezeda utca 7, Budapest, Magyarország
Wojno, Tadeusz, Prof.	Politechnica, Warszava, Polska
Öpik, Ar., Dr.	Bureau of Mineral Resources, 485 Bowrke Str., Melbourne, C 1, Victoria, Australia

Ulkomaiset jäsenet — Foreign members

Adamson, Olge J., Fil. dr.	Postboks 42, Smestad, Oslo, Norge
Ahrens, L. H., D.Sc.	Dept. of Geology, Mass. Inst. Technology, Cambridge, Mass., U.S.A.
Backlund, H. G., Prof.	Geol. Instit., Uppsala, Sverige
Barbour, George B., Prof.	University of Cincinnati, Cincinnati, Ohio, U.S.A.
Brognon, Georges, Ingenieur Civil des Mines	7 Avenue des Chênes, Rhode St. Genese, Bruxelles, Belgique
Brotzen, Otto, Fil.lic.	Bolidens Gruv A.—B., Boliden, Sverige
Bütler, H., Dr.	Rheinhaldenstr. 105, Schaffhausen, Schweiz
Chen, Kwan Yan, Fil.lic.	Dept. of Geology, National University, Peking, China

Collet, Leon W., Prof.	Laboratoire de géologie de l'Université, Quai de l'Ecole de Medicine, Geneve, Suisse
Collini, Bengt, Fil.lic., Laborator	Lantbrukshögskolan, Uppsala 7, Sverige
Dahlström, Elis, Fil.lic., Chefgeolog	Promenaden 17—19, Falun, Sverige
Disler, Jürg, Dr.	909—85 Richmond Street W. Toronto, Ont., Canada
Du Rietz, Torsten, Fil.dr,	Bolidens Gruv A.—B., Boliden, Sverige
Dutt, A., Geologist	Geological Survey, 27 Chrowringhee, Calcutta, India
Ebert, Heinz, Dr.	Depart. Nacional du Producão Mineral, Avenida Pasteur 404, Rio de Janeiro, Brasil
Eriksson, Tryggve, Fil.lic., Statsgeolog	Sveriges Geologiska Undersökning, Stockholm 50, Sverige
Frederickson, A. F., Dr.	Dept. of Geology, Washington University, St. Louis, Missouri, U.S.A.
Fromm, Erik, Fil.lic., Statsgeolog	Sveriges Geologiska Undersökning, Stockholm 50, Sverige
Fuster, José M., Prof.	Laboratorio de Geologia, San Bernardo 51, Madrid, España
Föyn, Sven, Dr., Directör	Norges Geologiske Undersøkelse, Josefines gate 34, Oslo, Norge
von Gaertner, Hans Rudolf, Dr.	Reichsamt für Bodenforschung, Wiesenstrasse 74, Hannover, Deutschland
Gavelin, Sven, Prof.	Stockholms Högskola, Stockholm, Sverige
Gevers, T. W., Prof.	Geological Dept., University of Witwatersrand, Johannesburg, South Africa
Gjelsvik, Tore, Dr.	Norges Geologiske Undersøkelse, Josefines gate 34, Oslo, Norge
Graeter, Paul, Dr.	Mineral. Inst., Bernoullianum, Birmannsgasse 48, Basel, Schweiz
Grip, Erland, Fil.lic., Chefsgeolog	Bolidens Gruv A.—B., Boliden, Sverige
Hagner, Arthur F., Dr.	Dept. or Geology, University of Illinois, Urbana, Ill., U.S.A.
Hall, A. Jean, M.A.	P.O. Box 709, Kitwe, Northern Rhodesia
Hejtman, Bohuslaw, Dr.	Sekaninova 16, Praha XIV, Czechoslovakia
Heurtebize, Georges, Géologue	Laboratoire de Géologie de la Sorbonne, 1. rue Victor Cousin, Paris 5e, France
Hjelmqvist, Sven, Prof.	Geologiska Institutionen, Lund, Sverige
Homma, Fujio, Prof.	Inst. of Petrology & Geology, Imperial University, Kyoto, Japan
Jaanusson, Valdar, Fil.lic	Paleontol. Instit., Uppsala, Sverige
Joensuu, Oiva, Fil.maist.	Dept. of Geology, University of Chicago, Chicago 37, Ill. U.S.A.
Järnefors, Björn, Fil.lic.	Sveriges Geologiska Undersökning, Stockholm 50, Sverige
Kautsky, Gunnar, Fil.lic.	Sveriges Geologiska Undersökning, Stockholm 50, Sverige
von Knorring, Oleg, Fil.mag.	Geol. Survey, Batu Gajah, Perak, Federation of Malay

Koark, Hans Joachim, Dr.	Geol. Instit., Uppsala, Sverige
Kranck, E. H., Prof.	McGill University, Montreal, Canada
Kuenen, Ph. H., Prof.	Geol. Instituut, Melkweg 1, Groningen, Nederland
Kullerud, Gunnar, Dr.	Geophysical Laboratory, 2801 Upton St., N.W., Washington 8, D.C., U.S.A.
Kulling, Oscar, Fil.dr., Docent, Statsgeolog	Östermalmsgatan 44, Stockholm, Sverige
Landergren, Sture, Laborator, Docent	Sveriges Geologiska Undersökning, Stockholm 50, Sverige
Lelubre, Maurice, Dr.	Service de la carte géologique, 14, Brd. Baudin, Alger
Mac Leod, John, M.Sc.	23 Avalon Place, Hamilton, Ont., Canada
Makela, Anna H., Dr.	U.S. Geological Survey, Washington 25, D.C., U.S.A.
Marklund, Nils, Fil.lic.	Geol. Instit. Uppsala, Sverige
Martna, Jüri, Fil.kand.	Torsvikssvängen 16 ^{III}, Lidingö, Sverige
Meier, Otto, Fil.dr.	Svenska Diamantbergborrnings A.—B., Kungsgatan 44, Stockholm C, Sverige
Melcher, Geraldo, Géologue	Alameda Franca 1329, São Paulo, Brasil
Mellis, Otto, Prof.	Mineral. Instit., Stockholms Högskola, Stockholm, Sverige
Mineralogisch-petrologisches Institut	Bernoullianum, Basel, Schweiz
Noe-Nygaard, Arne, Prof.	Østervoldgade 7, København K, Danmark
Pilava-Podgurski, N., Fil.lic.	Lustigknoppsvägen 4 C, Falun, Sverige
Preston, John, Dr.	Dept. of Geology, Queen's University, Belfast, Ireland
Roubault, Marcel, Prof.	Université de Nancy, 94 Avenue de Strasbourg, Nancy, France
Ryan, J. Donald, Dr.	Dept. of Geology, Lehig University, Betlehem, Pa, U.S.A.
Sahlin, Anders, Fil.mag.	Geol. Instit., Uppsala, Sverige
Sandrea, André, Dr.	61 Rue de Buffon, Paris, France
San Miguel, Alfredo, Prof.	Laboratorio de Petrografia, Universidad de Barcelona, España
Schachner, Doris, Prof., Dr.	Technische Hochschule, Wüllnerstrasse 2, Aachen, Deutschland
Shaw, Denis. M., Ph.D.	Geology Dept., McMaster University, Hamilton, Ont., Canada
Shuiab, S. M., Dr.	Mineralogy Dept., University of Punjab, Lahore, W-Pakistan
van Straaten, L.M.J.U., Dr.	Geol. Instituut, Melkweg 1, Groningen, Nederland
Suisse, M., Géologue	Boîte postale 210 Bouake, Cote d'Ivoire, Afrique Occidentale Francaise
Tišler, Janko, Dr.	Novi Sad, Maksima Gorkog. 22, Jugoslavija
de Waard, Dirk, Prof. Dr.	Djalan Taman Sari 64, Bandung, Indonesia
Watanabe, Takeo, Prof.	Geol. Inst., Faculty of Science, Tokyo Univ., Tokyo, Japan
Welin, Eric, Fil.mag.	Fredrikslundsg, 14, Bromma, Sverige
Wells, M. K., M.Sc.	Geology Dept., University College, London, W.C. 1, England

Wenk, Eduard, Dr.	Mineralogisch-petrologisches Institut d. Universität, Basel, Schweiz
Wickman, Frans E., Prof.	Riksmuseet, Stockholm 50, Sverige
Yoder, Hatten S., Dr.	Geophysical Laboratory, 2801 Upton Street, N.W., Washington 8, D.C., U.S.A.
Zāns, Verners, State Geologist	7. Tucker Ave., Kingston, Jamaica, B.W.I.
Zeidler, Waldemar, Bergsingeniör	Närkes Kvarntorp, Sverige
Åhman, Erik, Fil.lic.	Sveriges Geologiska Undersökning, Stockholm 50, Sverige
Ödman, Olof H., Prof.	Kgl. Tekn. Högskolan, Stockholm Ö, Sverige

Kotimaiset jäsenet — Members

Aario, Leo, Prof.	Yliopiston Maantieteellinen Laitos, Siltasaarenk. 8, Helsinki
Ahlfors, Bruce, Dipl.ins.	Förby
Aho, Lea, Yliopp.	Bulevardi 29, Helsinki
Alanko, Risto, Dipl.ins.	Lohja
Alenius, P., Dipl.ins.	Lauttasaarentie 48, Lauttasaari
Arvela, A., Dipl.ins.	Koroistenti. 6—8 F 14, Helsinki
Auer, Väinö, Prof.	Fredrikink. 66 B, Helsinki
Aulanko, Heikki, Dipl.ins.	Outokumpu Oy, Outokumpu
Aurola, Erkki, Fil.toht.	Kauppiaank. 5 A 14, Helsinki
Backman, A. L., Fil.dr.	St. Robertsg. 25 A, Helsingfors
Borg, Torvald, Fil.kand.	Mannerheimv. 92 A 21, Helsingfors
Donner, Joakim, Fil.toht.	Pohjoisranta 12 A 3, Helsinki
Edelman, Nils, Fil.toht.	Haverin kaivos, Viljakkala
Enkovaara, Antti, Fil.kand.	Oulainen
Ervamaa, Pentti, Fil.kand.	Gyldenint. 12 C 14, Helsinki
Erämetsä, Olavi, Prof.	Otakallio, Otaniemi
Eskola, Salli, Apulaisprof.	Pihlajatie 49 B 27, Helsinki
Friberg, Sven, Fil.mag.	Varkaus 9, III
Granö, J. G., Prof.	P. Rautatienk. 11 A, Helsinki
Granö, Olavi, Fil.maist.	Iitintie 6, Helsinki
Grönros, Y., Ins.	Vänr. Stoolink. 3 A 14, Helsinki
Haapala, Olavi, Dipl.ins.	Outokumpu Oy, Tampere
Haapala, Paavo, Fil.toht.	Otakallio, Otaniemi
Halme, Erkki, Yliopp.	Mechelinink. 10 A 16, Helsinki
Halonen, Olli J., Yliopp.	P. Robertink. 4—6 N 78, Helsinki
Halonen, T. O., Yliopp.	Leppäsuonk. 7 A, Helsinki
Hausen, H., Prof.	Södra Strandv. 5, Brändö
Heikkinen, Aulis, Fil.maist.	Bulevardi 29, Helsinki
Heino, Eino, Dipl.ins.	Koroistent. 6—8. Helsinki
Heinonen Leo, Fil.maist.	Pihlajat. 15, Helsinki
Heiskanen, Erkki, Fil.kand.	Ruskealan Marmori Oy. Savonlinna
Heiskanen, V. A., Prof.	Messeniuksenk. 10 A, Helsinki
Helkavaara, Eero K., Fil.maist.	Hämeentie 3, Malmi
Heloma, Erkki, Yliopp.	Pengerk. 19 A 12, Helsinki
Helovuori, Olavi, Fil.kand.	Outokumpu Oy, Outokumpu

Himmi, Reino, Fil.maist.	Outokumpu Oy, Tampere
Holm, Caj, Vuori-ins.	Tytyrink. 3, Lohja
Huhma, Arto Olavi, Fil.kand.	Outokumpu Oy, Outokumpu
Huhma, Maija, Fil.maist.	Outokumpu Oy, Outokumpu
Huhta, Jussi, Fil.kand.	Aijala
Hukki, Risto, Prof.	Otakallio, Otaniemi
Huopaniemi, Pertti, Yliopp.	Yliopiston geologinen laitos, Snellmanink. 5, Helsinki
Hyppönen, Viljami, Fil.maist.	Tiaisenk. 44, Joensuu
Hytönen, Kai, Fil.maist.	5. linja 14 B 63, Helsinki
Hyvärinen, Lauri, Fil.kand.	Bulevardi 29, Helsinki
Hyyppä, Esa, Fil.toht.	Bulevardi 29, Helsinki
Hyyppä, Jussi, Fil.kand.	Bulevardi 29, Helsinki
Hämäläinen, Kaisa, Fil.kand.	Mechelinink. 40 A 16, Helsinki
Hämäläinen, Viljo, Fil.kand.	Santavuorentie 7 B 31, Helsinki
Härme, Maunu, Fil.toht.	Bulevardi 29, Helsinki
Ingnatius, Heikki, Fil.maist.	Savilankatu 1 A 11, Helsinki
Isokangas, Pauli, Fil.kand.	Outokumpu Oy, Alpua
Jalander, Holger, Dipl.ins.	Tukholmank. 7 A, Helsinki
Juurinen, Aarno, Fil.maist.	Pihlajatie 50 A 5, Helsinki
Jäntti, L. Olavi, Fil.maist.	Mechelinink. 51 B 20, Helsinki
Järvinen, Kauko, Prof.	Bulevardi 34a A, Helsinki
Järvinen, Matti, Fil.kand.	Eerikink. 40 A 17, Helsinki
Kahma, Aarno, Fil.toht.	Bulevardi 29, Helsinki
Kaitaro, Simo, Fil.toht.	Bulevardi 29, Helsinki
Kalla, Juhani, Fil.maist.	Aallonk. 2 B, Rovaniemi
Kanerva, Reino, Fil.maist.	Koskelantie 23 H, Helsinki
Kantele, Helvi, Fil.maist.	Vilhonk. 3 D 50, Mikkeli
Kauranne, Kalevi, Fil.kand.	Bulevardi 29, Helsinki
Keränen, Jaakko, Prof.	Topeliuksenk. 1, Helsinki
Kivinen, Erkki, Prof.	Mariank. 13b B, Helsinki
Konttinen, Lauri, Yliopp.	Kuninkaank. 15 H, Tampere
Koponen, Olavi, Dipl.ins.	Kupittaan Savi Oy, Turku
Korpela, Kauko, Fil.kand.	Pyhäkoski, Muhos
Koskela, Erkki, Dipl.ins.	Orijärvi, Kisko
Koskinen, Juhani, Yliopp.	Kauppiaank. 4 A 2, Helsinki
Kotilainen, Mauno J., Prof.	Korkeavuorenk. 8 C, Helsinki
Kouvo, Olavi, Fil.kand.	Outokumpu Oy, Outokumpu
Kulonpalo, Max, Fil.maist.	Bulevardi 29, Helsinki
Kurppa, Reino, Dipl.ins.	Outokumpu Oy, Tampere
Laatio, Gunnar, Vuori-ins.	Outokumpu Oy, Alpua
Laitakari, Aarne, Prof.	Bulevardi 29, Helsinki
Laitakari, Aatto J., Fil.kand.	Susitie 10 B 21, Herttoniemi
Laitakari, Ilkka, Fil.kand.	Erkki Melartinintie 11, Pukinmäki
Laitakari, Lauri, Kapt. evp.	Savio
Laitala, Matti, Fil. kand.	Leppäsuonk. 7 A 409, Helsinki
Laiti, Ilpo, Fil.kand.	Pohjoisranta 20 C 60, Helsinki
Lauerma, Raimo, Yliopp.	Leppäsuonk. 9 C 204, Helsinki
Laurila, Matti, Fil.kand.	Outokumpu Oy, Outokumpu
Lehijärvi, Mauno, Fil.maist.	Minna Canthink. 18 A 12, Helsinki
Lindberg, Eric, Yliopp.	Mannerheimintie 142 A 10, Helsinki

Lindholm, Ole, Fil.kand.	Otanmäki Oy, Otanmäki
Linna, Antti, Dipl.ins.	Messukylä
Lokka, Lauri, Prof.	Mannerheimintie 35 A, Helsinki
Lukkala, O. J., Prof.	Aurorank. 19 A, Helsinki
Lupander, Kurt, Fil.mag.	Grundvägen 24 A 18, Munksnäs
Lyytikäinen, Erkki, Yliopp.	Väinölänt. 20, Tapanila
Maijala, Paavo, Dipl.ins.	Mäntyt. 3, Helsinki
Makkonen, Väinö, Fil.maist.	Pateniemi
Marmo, Vladi, Fil.toht.	Geological Dptm., New England, Freetown, Sierra Leone, W.A.
Marttila, Erkki, Yliopp.	Yliopiston geologian laitos, Snellmanink. 5, Helsinki
Matisto, Arvo, Fil.maist.	Apollonk. 23 B 43, Helsinki
Mattila, Jorma, Fil.maist.	Menninkäisentie 6 H, Tapiola
Meriläinen, Kauko, Fil.maist.	Mechelinink. 10 A 16, Helsinki
Metzger, A. A. Th., Prof.	Pargas
Mikkola, Aimo, Fil.toht.	Outokumpu Oy, Alpua
Mikkola, Toini, Fil.maist.	Bulevardi 29, Helsinki
Mikkola, Toivo, Fil.kand.	Suomen Malmi Oy, Otakallio, Otaniemi
Mikkonen, Antti, Fil.kand.	Pihlajat. 37 as. 5, Helsinki
Mäntynen, Matti, Fil.maist.	Lielahdent. 2, Lauttasaari
Mölder, Karl, Fil.toht.	Bulevardi 29, Helsinki
Neuvonen, K. J., Fil.lis.	Bulevardi 29, Helsinki
Nieminen, Kalervo, Dipl.ins.	Jääkärink. 10 A 11, Helsinki
Niini, Risto, Prof.	Snellmanink. 23 E, Helsinki
Nisonen, Eino, Fil.maist.	Töölönk. 36 B 32, Helsinki
Nordenswan, Einar A. O., Fil.maist.	Runebergink. 17 A, Helsinki
Nortio, Jaakko, Fil.kand.	Leppäsuonk. 7 B 101, Helsinki
Nousiainen, Erkki, Isänn.	Ruskealan Marmori Oy, Louhi
Nuutilainen, Juhani, Fil.kand.	Sorsakoski
Nykänen, Osmo, Yliopp.	Huovitie 6, A 6, Pohjois-Haaga
Nystén, Henrik, Dipl.ing.	Karlav. 3 A 16, Drumsö
Näykki, Ossi, Fil.maist.	Pihlajatie 43 A 5, Helsinki
Ohlsson, Birger, Fil.mag.	Handelshögskolan vid Åbo Akademi, Åbo
Oivanen, Paunu, Yliopp.	Agricolank, 5 A 16, Helsinki
Ojanperä, Pentti, Fil.maist.	Bulevardi 29, Helsinki
Okko, Veikko, Fil.maist.	Bulevardi 29, Helsinki
Oksanen, E. V., Johtaja	Riihit. 23 B, Helsinki
Paarma, Heikki, Fil.maist.	Otanmäki Oy, Otanmäki
Palin, Urpo, Fil.kand.	Meritullink. 12 E 23, Helsinki
Parras, Kauko, Fil.maist.	Tytyrink. 3, Lohja
Pehkonen, Eero, Fil.kand.	Bulevardi 29, Helsinki
Pehrman, Gunnar, Prof.	V. Strandg, 17, Åbo
Penttilä, Aarne, K., Herra	Valtak. 60 A, Lappeenranta
Penttilä, Seppo, Fil.kand.	Maskunt. 8, Helsinki
Pesola, Pentti, Dipl.ins.	Paakkila, Tuusniemi
Pipping, Fredrik, Yliopp.	Kruunuvuorenk. 7 A 6, Helsinki
Pohjanlehto, V. O., Herra	Harjunpää, Pori
Puranen, Maunu, Fil.maist.	Bulevardi 29, Helsinki
Pääkkönen, Veikko, Fil.maist.	P. Kaari 6 A 8, Lauttasaari
Raja-Halli, Heikki, Dipl.ins.	Suomen Malmi Oy, Otakallio, Otaniemi

Rancken, Ragnar, Fil.lis.	Lönnrotsg. 7 B 13, Helsingfors
Rankama, Kalervo, Prof.	Yliopiston geologian laitos, Snellmanink. 5, Helsinki
Renvall, Åge, Fil.mag.	Haveri Gruva, Viljakkala
Repo, Reino, Fil.kand.	Bulevardi 29, Helsinki
Rosblom, Heikki, Yliopp.	Linnavuori
Rouhunkoski, Pentti, Yliopp.	Hietalahdenranta 15 A, Helsinki
Runolinna, Urmas, Dipl.ins.	Otanmäki Oy, Otamäki
Räsänen, Veikko, Fil.kand.	Korkeavuorenkatu 13 B 34, Helsinki
Saastamoinen, Jyry, Yliopp.	HYY:n asuntola, P. Kaari 25, Lauttasaari
Sahama, Th. G., Prof.	Yliopiston geologian laitos, Snellmanink. 5, Helsinki
Saksela, Martti, Prof.	Temppelik. 21, Helsinki
Salli, Ilmari, Fil.kand.	Bulevardi 29, Helsinki
Salmi, Martti, Fil.toht.	Bulevardi 29, Helsinki
Salminen, Kyllikki, Fil.maist.	Bulevardi 29, Helsinki
Saraste, Ahti, Herra	Topeliuksenk. 37 A 11, Helsinki
Sauramo, Matti, Prof.	Tunturik. 4, Helsinki
Savolahti, Antti, Fil.kand.	Jääkärink. 10 A 12, Helsinki
Savolainen, E., Fil.maist.	Bulevardi 29, Helsinki
Schröder, Henning, Fil.maist.	Nakkila
Seitsaari, Juhani, Fil.toht.	Runebergink. 49 A 19, Helsinki
Siikarla, Toivo Ilmari, Dipl.ins.	Bulevardi 29, Helsinki
Similä, Pentti, Dipl.ins.	Asemakatu 5—7, Lohja
Simola, Kaarlo, Kaivosmittaaja	Outokumpu, Raivionmäki
Simola, Torsti, Dipl.ins.	Ilmarink. 4 B, Helsinki
Simonen, Ahti, Fil.toht.	Bulevardi 29, Helsinki
Siren, Arne, Fil.dr.	Mecheling. 26 B 23, Helsingfors
Soveri, Urpu, Fil.toht.	Meritullink. 21 A, Helsinki
Stenberg, Aarne, Fil.kand.	Gyldenint. 8 A, Helsinki
Stigzelius, Herman, Tekn. dr.	Bulevarden 26 A 10, Helsingfors
Stjernvall, Gisela, Fil.maist.	Bulevardi 32, Helsinki
Stolpe, Tor Björn, Fil.kand.	Mannerheimintie 79 B 40, Helsinki
Strandström, Georg, Fil.kand.	Virkkala
Suila, Matti, Fil.kand.	Oy Vuoksenniska Ab, Imatrankoski
Sundell, I. G., Dr.	Museig. 17 A 9, Helsingfors
Suominen, Eero, Fil.maist.	Tehtaank. 1 A, Helsinki
Suominen, Paavo, Yliopp.	Savilank. 2 C, Helsinki
Suvenmaa, L., Herra	Somero
Syvänen, Marjatta, Fil.kand.	Bulevardi 29, Helsinki
Säynäjärvi, Klaus, Fil.maist.	Museokatu 44 A 12, Helsinki
Tammekann, August, Prof.	Kampink. 8 C, Helsinki
Tanner, Heikki, Vuori-ins.	Outokumpu Oy, Outokumpu
Tavela, Matti, Fil.kand.	Suurlohjank. 33, Lohja
von Timroth, Michael, Dipl.ins.	Ojamon kaivos, Lohja
Toivonen, A. V. P., Fil.maist.	Bulevardi 29, Helsinki
Toivonen, N. J., Prof.	Hallitusk. 3, Helsinki
Tuominen, Heikki, Fil.maist.	Suomen Malmi Oy, Otakallio, Otaniemi
Turunen, Eero, Dipl.ins.	Aijala
Tynni, Risto, Fil.kand.	Vakkola, Askola
Vaasjoki, Oke, Fil.lis.	Kammionk. 5 D 29, Helsinki

Vaasjoki, Pirjo, Rouva	Kammionk. 5 D 29, Helsinki
Wahl, W., Prof.	Bergmansg. 7 D, Helsingfors
Waldén, Olavi, Fil.kand.	Bulevardi 29, Helsinki
Valovirta, Veikko, Metsänhoit.	Bulevardi 29, Helsinki
Vanhala, Risto, Fil.maist.	Laivanvarustajank. 7 A 11, Helsinki
Vanninen, Mikko, Fil.kand.	Kauklahti
Varma, Arno, Fil.kand.	Aijala
Veijola, Erkki, Yliopp.	Snellmanink. 23 E 15, Helsinki
Veltheim, Valto, Fil.maist.	Bulevardi 29, Helsinki
Wennervirta, Heikki, Fil.kand.	Geofysisk Malmleting. Kong. Ingestgt. 5, Trondheim, Norge
Vesasalo, Arvo, Fil.kand.	Otavantie 8 A 2, Lauttasaari
Wessman, Curt, Fil.mag.	Toivolag. 27, Åbo I
Wessman, Tor H., Fil.mag.	Forsbyvägen 31 A 9, Helsingfors
Viento, Aimo, Joht.	Messeniuksenk. 8 A 33, Helsinki
Wiik, Hugo Birger, Fil.lic.	Fredsg. 11 C 42, Helsingfors
Viluksela, Erkki, Fil.kand.	Outokumpu Oy, Outokumpu
Virkkala, Kalevi, Fil.toht.	Bulevardi 29, Helsinki
Virkkunen, Marjatta, Rouva	Yliopiston geologian laitos, Snellmanink. 5, Helsinki
von Volborth, A., Fil.lis.	Miinalant. Leppävaara
Vorma, Atso, Yliopp.	Koivistontie 11, Leppävaara
Vormisto, Kauno, Yliopp.	P. Hesperiank. 5 A 10, Helsinki
Vuorjoki, Aarre, Fil.maist.	Kotipolku 16, Käpylä
Vähätalo, Veikko, Fil.toht.	Outokumpu Oy, Outokumpu
Väyrynen, Heikki, Prof.	Temppelik. 13, Helsinki
Yletyinen, V., Fil.maist.	Susitie 10 C, Herttoniemi
Öhman, Börje, Fil.mag.	Granbacka II A, Pargas

A SMALL NICKELIFEROUS SUBSILICIC CLUSTER AT PUUMALA IN SOUTHEAST FINLAND [1]

BY

VLADI MARMO

Geological Survey of Finland, Helsinki

ABSTRACT

A nickeliferous subsilicic cluster embedded in gneiss is described. The sulphide minerals are used as a means of interpreting the progress of metamorphism of this cluster.

CONTENTS

PETROLOGY OF THE CLUSTER

In southeastern Finland there are large areas occupied by gneissose rocks consisting of migmatized sediments. In the area surrounding Lake Saimaa and on its islands there occur many small subsilicic clusters embedded in gneiss and containing pyrrhotite in abundance. One of them, in Puumala on a narrow peninsula 4 km south of the church, contains pentlandite, chalcopyrite, and pyrrhotite.

The peninsula is from 1 km to 2 km in breadth and about 10 km in length. Geologically, it consists of gneisses of variable bulk composition. They are paragneisses with a basis of argillaceous composition, often garnet-bearing, and intimately mixed with microcline-rich aplitic granite which often occupies small independent areas also.

---

[1] Received May 15, 1953.

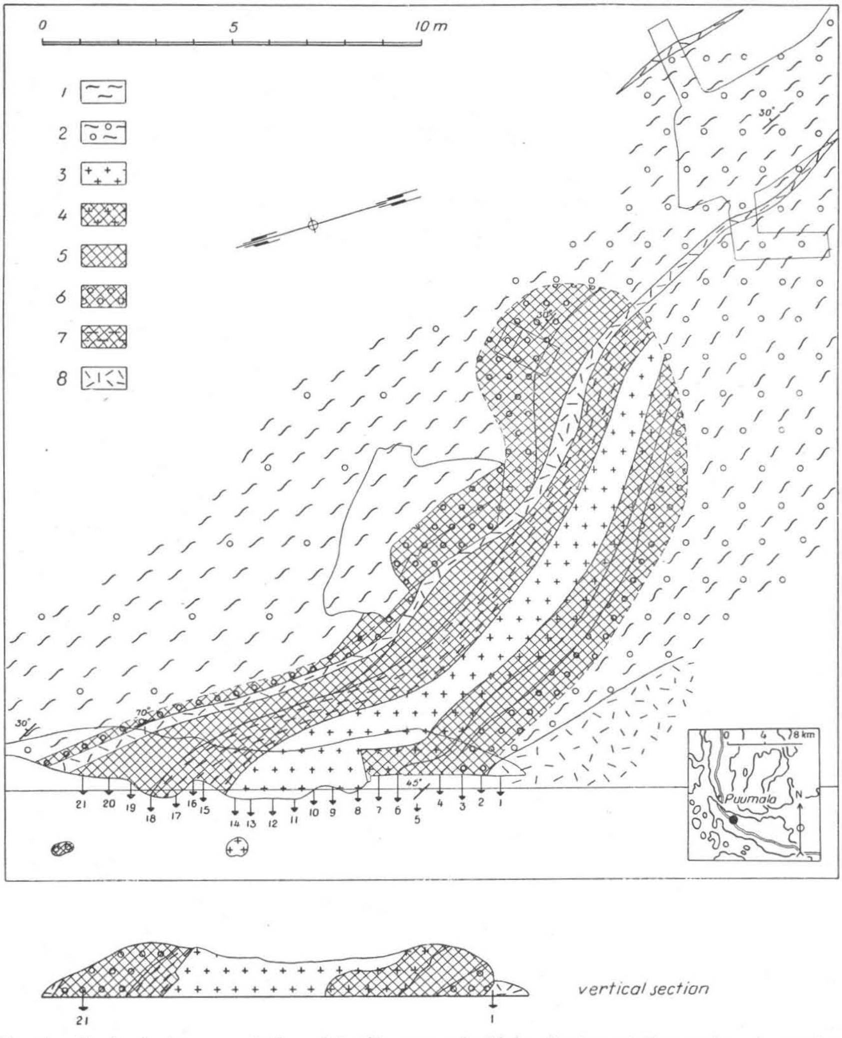


Fig. 1. Geological map of the nickeliferous subsilicic cluster at Puumala. 1, gneiss: 2, garnet gneiss; 3, diopside gabbro; 4, diopside amphibolite; 5, amphibolite; 6, garnet amphibolite; 7, strongly schistose amphibolite; 8, pegmatite. The figures indicate points of sampling.

The gneisses usually have a gentle dip of the schistosity, usually about 40°, but rather often from 10° to 20° only. The direction of strike and dip varies greatly within the area.

The subsilicic cluster occurs on the highway between Puumala and Imatra and is well exposed in a road cutting. The cluster seems to be lenticular in form and is conformable to the surrounding gneiss (Fig. 1). This cluster is up to 10 m thick and 30 m wide. Its length is not known

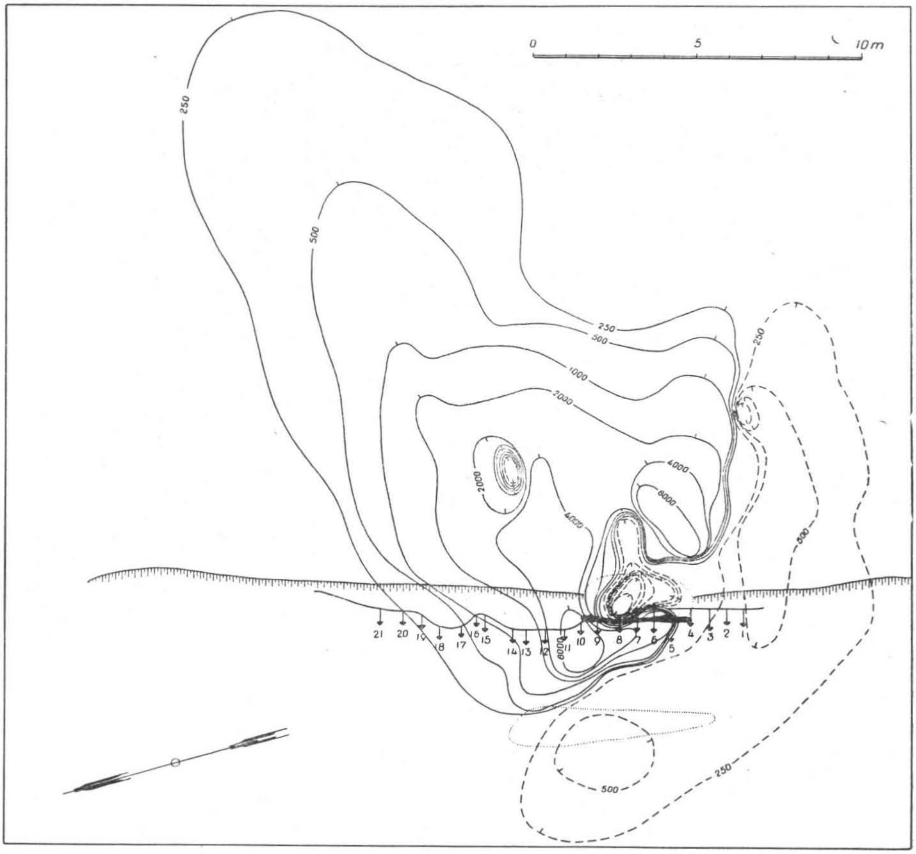


Fig. 2. Magnetic anomalies of the nickeliferous subsilicic cluster at Puumala. The marked outcrop lies between No. 4 and No. 10 of Fig. 1.

with certainty, but the magnetic anomalies indicate that it would not seem to be much more than 60 m (Fig. 2).

The rock of the cluster consists of diopside, tremolite, plagioclase, biotite, and some quartz. On its border zones garnet occurs as well. Furthermore, the cluster contains disseminated ore minerals. As a whole it forms a subsilicic body with a distinctly developed schistosity. In detail however, the rock is nonhomogeneous. Different parts in the direction perpendicular to the schistosity vary in their mineralogical composition. In Fig. 1 the figures mark the points at which the Ni and Cu content of the rock was determined. Using the same figures in a description of the rock itself, the following picture is obtained:

1—5. Amphibolite containing predominantly plagioclase, tremolite, quartz, and biotite. Ore minerals are present in moderate amount. The tremolite is resorbed, impregnated, and surrounded by strips of sulphides, and often contains remainders

Fig. 3. Diopside containing »drops» of quartz. In righthand lower corner, there are needles of tremolite (dark gray). Lamellae in the lefthand upper corner also consist of tremolite. Puumala. Magn. 40 ×, one Nicol.

of resorbed diopside in its centre Furthermore, this amphibolite encloses in its marginal part very small grains of garnet.

6—7. Diopside amphibolite with plagioclase, diopside, tremolite, and biotite as the chief constituents. The diopside is often altered into amphibole and this into biotite. Ore minerals are present in moderate amount.

8—13. Diopside gabbro containing no amphibole. The diopside is fresh and occurs together with biotite. The plagioclase is slightly saussuritized as is the case in the whole cluster except in No. 14. Ore minerals are abundant.

14. The plagioclase is fairly fresh, whereas the diopside is slightly chloritized along cracks. Ore minerals are abundant.

15—18. This part of the cluster is highly schistose. All minerals are strongly deformed, and the lamellae of plagioclase and biotite are bent. The diopside contains »drops» of quartz. The tremolite is sparse and is enclosed by the diopside (Fig. 3). The plagioclase is strongly chloritized and saussuritized. Ore minerals are sparse.

19—21. The rock is very similar to that in Nos. 1—5, but diopside and garnet are more abundant. The diopside is chloritized and serpentinized along cracks, and the biotite also is often surrounded by pale chlorite. The plagioclase forms large crystals, and the spaces between them are filled by dark minerals. Ore minerals are precent in moderate amount.

Consequently, it seems that the marginal parts of the cluster and the part between points 15 and 18 (Fig. 1) are more metamorphosed than are the other, inner parts of the cluster.

The plagioclase of the nickeliferous cluster contains about 65 % An. When it is saussuritized, the saussurite often forms a zone within the crystal with boundaries parallel to its borders.

The diopside is usually fresh, or only slightly chloritized (or serpentinized) along cracks. Amphibole, when present together with diopside, seems to have formed from diopside (uralitization). The drops of quartz, often occurring in the diopside crystals, may be ascribed to the uralitization. The diopside is very pale, and according to measurements made by Mrs. Toini Mikkola, M. A., at the Geological Survey of Finland, its $2V\gamma = 56°$, $c \wedge \gamma = 48°$, $\gamma = 1.718$ and $\alpha = 1.692$.

The tremolite is colourless and, according to the determinations of Mrs. Mikkola, its $2V\alpha = 78°$, $c\wedge\gamma = 15°$, $\alpha = 1.640$ and $\gamma = 1.662$.

The biotite is reddish brown with α pale brownish yellow; β equals γ and both are dark reddish brown. Small variations were observed in the optical properties of biotite. Some flakes of biotite were separated from the rock. Some of these were megascopically rather black and some were brownish. The optical properties of both varieties according to Mrs. Mikkola were as follows

Black biotite	$2V\gamma$	1.5—2°	$\beta = \gamma$ 1.647
Brown »	»	0°	» 1.645

According to a partial chemical analysis made by Mr. V. Nurmi at the Geological Survey of Finland, the black biotite contains 16.7 % FeO, 3.96 % TiO_2, 14.38 % MgO and 0.20 % MnO, corresponding to about 2.8 MgO · 2.2 FeO in the formula of biotite. Heinrich (1946) published a graph showing the relationship between the index of refraction and the chemical composition of biotite. If the indices of the biotite from Puumala are interpolated on the graph of Heinrich, the percentage of Fe_2O_3 appears rather low. Hence in the formula of biotite, 1.4 Al_2O_3 · 0.6 TiO_2 may be written instead of 2 Al_2O_3.

Ilmenite and magnetite are the earliest ore minerals of the cluster. The ilmenite is disseminated as small, rounded grains throughout the whole cluster. The magnetite, also in the form of rounded grains, is comparatively rare, and often occurs together with ilmenite, forming, however, separate grains.

The pyrrhotite and pentlandite are of about the same age, but distinctly younger than the ilmenite and magnetite. Pyrrhotite is rather abundant in most parts of the cluster and occurs as grains but also brecciates the silicate minerals. In some polished sections of the pyrrhotite, thin lamellae are observed (Fig. 4), which are somewhat paler in colour than the main pyrrhotite, and less anisotropic. These lamellae differ from the lenticular lamellae found in pyrrhotite (Ramdohr 1950, p. 408), in being regular and forming a network.

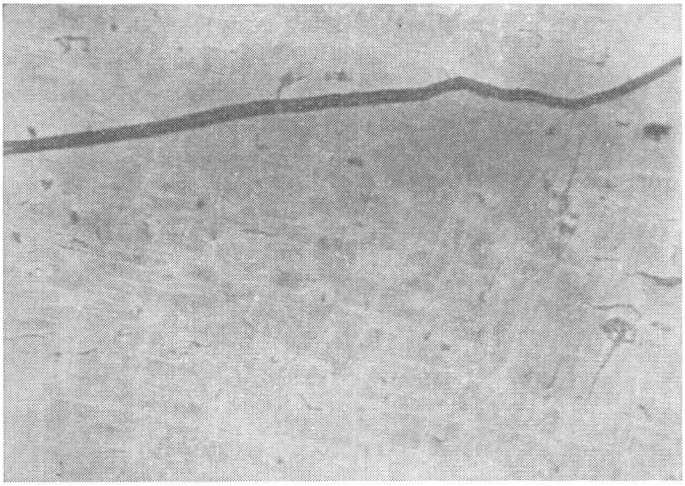


Fig. 4. Pyrrhotite containing a network of light-coloured lamellae. Puumala. Polished section, magn. 300 ×, unpolarized light.

Pentlandite grains always occur together with pyrrhotite. There is about twice as much pyrrhotite as pentlandite in the rock. The pentlandite also occurs in the pyrrhotite, forming flame-shaped bodies as the result of exsolution, like those described by Ramdohr (1950, p. 410).

Pyrite, forming small, strongly resorbed grains, is enclosed by pyrrhotite. The pyrite occurs also in fissures of pyrrhotite and chalcopyrite. This pyrite belongs to a very young generation (see below).

Chalcopyrite forms veins which penetrate and replace the pyrrhotite and brecciate the silicate minerals. It is of a distinctly younger age than any of the ore minerals previously mentioned.

The youngest ore minerals of the rock are those whose origin is due to secondary alteration of sulphides within the zone, at and just beneath the level of the circulating ground waters. The pyrrhotite has been altered into a lamellar mass containing marcasite-like sulphide and pyrite of later origin. Such products of alteration are always accompanied by abundant limonite. Pentlandite has been altered to a small extent into bravoite. In the fissures of the chalcopyrite a mineral was observed which is taken to be chalcocite. Furthermore, in the gossan of the cluster a greenish staining caused by some secondary copper mineral has been observed.

The nickel and copper content of different parts of the cluster was determined chemically in samples weighing from 1 $^1/_2$ kg to 2 kg (Table 1). The content varies in the different parts of the cluster. Platinum in traces is only present in that part of the basic rock containing very little amphibole.

Table 1. The Ni and Cu content of the different parts of the nickeliferous cluster from Puumala. The numbers of the samples correspond to those in Fig. 1, indicating the points from which the analyzed samples were taken. Analyst, A. Heikkinen, Geological Survey of Finland

Sample No	Ni, %	Cu, %
2	1.15	0.74
3	0.50	0.26
4	0.56	
5	0.40	0.27
6	0.57	
7	0.74	
8	trace	
9	0.46	
10	0.46	1.70
11	1.19	
12	0.53	
13	1.02	
14	0.96	0.38
15	0.15	
16	0.41	0.95
17	0.37	
18	0.89	
20	0.51	
21	0.62	0.11

Petrographically the rock is similar to some diopside gneisses, common in the Archaean of Finland, e. g. those described by Parras (1941) from southern Finland. Now, however, most of the diopside gneisses described by Parras are sedimentogenic diopside-bearing rocks. The nickeliferous cluster of Puumala cannot be interpreted as being of similar origin. Variations in its mineralogical composition are not a result of differently composed layers, but of schistosity. During the movements which caused this schistosity the metamorphosing agents obviously affected crushed parts of the rock (the lamellae of plagioclase are in some parts strongly bent) in a different degree. Some parts of the cluster have preserved the primary composition of the rock: plagioclase, diopside, and quartz; but other, weaker, parts have been altered into rock containing plagioclase, diopside, quartz, amphibole, and biotite.

Furthermore, if the assemblage of ore minerals is considered, it is observed that the paragenesis, ilmenite, magnetite, pentlandite, pyrrhotite, is characteristic of nickel ores of magmatic origin (Schneiderhöhn, 1949). The nickeliferous cluster of Puumala must therefore be considered magmatic, and diopside gabbro is the most appropriate name for the rock composing it.

Among many examples of well-investigated deposits of magmatic nickel ores, those of Chicagof Island, Alaska, may be mentioned. According to Kerr (1924, p. 370), the ore occurs in a gabbro which contains labradorite, pyroxene (isomorphic mixture of diopside and hedenbergite), hornblende ($c \wedge \gamma = 22°$), colourless edenite, and tremolite. The country rock of the Chicagof Island nickel ore is also very similar to the Puumala cluster: a quartz-schist. Overbeck (1917) believed that a definite relationship exists between the ore deposition and the contact of the schist with the intruding gabbro, because the nickel minerals are found either near the contact or in its vicinity.

The cluster of Puumala is, consequently, a »miniature» nickel ore deposit. If its dimensions (see p. 3) obtained from the magnetic anomalies are correct, the cluster is too small to be regarded as an independent magmatic body. Therefore it must be supposed to be a segregation derived from some bigger magmatic body, the situation of which, however, is unknown.

From the fact that the cluster lies conformably with the surrounding schists (now gneisses) and that it is small and, according to its petrological nature, of magmatic origin, the conclusion is drawn that during the folding and penetrative movements of the strata the cluster was removed from a bigger magmatic body and brought into a new environment — perhaps a considerable distance from its mother rock. Whether its primary environment was similar to the present one or not, cannot be inferred, but the geology of the Puumala area indicates that diopside gabbro intruded the argillaceous sediments. The gabbro contained much nickel that was perhaps concentrated in the marginal parts of the intrusive body (see Overbeck 1917), and a part of it was subsequently removed during the movements.

METAMORPHISM OF THE CLUSTER

The central part of the cluster (between points 8 and 14 in Fig. 1) contains diopside, biotite, and plagioclase (labradorite). Such a mineral assemblage, common in a diopside gneiss, was described by Parras (1941) who proposed the separation of this assemblage as a facies of its own in Eskola's system of mineral facies (1920).

Turner (1948, p. 70) considered 700°C—750°C not too low a value for the temperature of transition between the pyroxene hornfels facies and the amphibolite facies. Rosenqvist (1952, p. 61) suggested 750°C as an average for the expulsion of water from hornblende at atmospheric pressure. Schneiderhöhn (1941, p. 114) supposed 800°C—600°C the lowest temperature for the formation of liquid magmatic ores.

In the Puumala rock, ilmenite, magnetite, pyrrhotite, and pentlandite are present, all being typical ore minerals for high-temperature conditions. Judging from a microscopic study of polished sections of the sulphides from the Puumala cluster, the ilmenite and magnetite obviously represent an earlier stage of mineralization than do the pyrrhotite and pentlandite, hence corresponding to a temperature higher than the transition temperature between the pyroxene hornfels facies and the amphibolite facies. Consequently, the ilmenite and magnetite obviously originated in conditions of the pyroxene hornfels facies. It may be that this mineral combination is earlier than the combination diopside-biotite-labradorite, but the sulphides, pyrrhotite and pentlandite were probably formed in conditions of metamorphism characterized by the above-mentioned mineral assemblage. Hence it seems that the present association, diopside, biotite, labradorite, would correspond to a high temperature stage of the amphibolite facies.

Pentlandite occurs also as flame-like bodies resulting from unmixing in the pyrrhotite. Scholtz (1933) stated that consideration of the distributions of pyrrhotite and pentlandite in some African ores leads one to the conclusion that, although both minerals appear to have separated more or less simultaneously, pentlandite continued to crystallize after pyrrhotite had been completely precipitated. He consequently supposed that the ores consisted primarily of a solid solution of pentlandite in pyrrhotite. Hewitt (1938, pp. 317—318) examined the problem of the exsolution of pyrrhotite and pentlandite experimentally and found that pentlandite may be caused to unmix from pyrrhotite by slow cooling from 800°C, but that pentlandite usually goes into solid solution in pyrrhotite between the 425°C and 450°C. Furthermore, he said that the limited solubility of pentlandite in pyrrhotite makes it seem improbable that all pentlandite is a product of exsolution, but that probably large masses are produced by crystallization from a field of immiscible liquids. The temperature for exsolution of pentlandite from pyrrhotite was determined also by Hawley, Colgrove, and Zurbrigg (1943) who gave 425°C—450°C.

In the pyroxene-biotite-plagioclase rock of Puumala the flames of pentlandite in pyrrhotite are evidently products of an exsolution, which probably occurred at a temperature between 425°C and 800°C, but more probably between 425°C and 450°C (see above). If primary pyrrhotite and pentlandite (crystallized from immiscible liquids before the exsolution of pentlandite from pyrrhotite), are regarded as products of orthomagmatic origin (above 600°C) as is commonly believed, then the exsolution of flame-like pentlandite must correspond to a temperature much less than the transitional temperature between the pyroxene hornfels facies and the amphibolite facies.[1]

At this temperature alteration of the assemblage of minerals of the rock from the pyroxene hornfels facies into the assemblage corresponding to the amphibolite facies, and also the uralitization of diopside, would be expected, as has actually happened in most parts of the cluster. In the innermost part, however, the diopside is relatively fresh and amphibole is absent,

---

[1] Vähätalo (1951, 1953) found that the flame-like bodies in pyrrhotite from the Outokumpu mine are not pentlandite but linneaite. In Puumala, however, the flames are definitely pentlandite.

probably because of the insufficient amount of water present. The ordinary diaphthoretic metamorphism is always accompanied by incorporation of water. Therefore, in Puumala the metamorphism took place only within the ore minerals (exsolution of pentlandite from pyrrhotite). When discussing the system, $MgO—Al_2O_3—SiO_2—H_2O$, Yoder (1952) emphasized the importance of water in the development of mineral associations of the related metamorphic facies. Rosenqvist (1952, p. 91) called metamorphism characterized by an insufficient amount of water »dipsenic».

The mineral association, diopside-biotite-plagioclase, is considered to belong to a subfacies of the amphibolite facies and to correspond to a temperature distinctly below the transitional point between the pyroxene hornfels facies and the amphibolite facies. Because the above-mentioned mineral association seems to have been stable also at the temperature of exsolution of pentlandite from pyrrhotite (at 425°C—450°C) and has survived these conditions because of the presence of insufficient water, the corresponding subfacies may be considered to represent the »dipsenic amphibolite facies».

The metamorphism of ore minerals thus is incorporated in Eskola's mineral facies.

In the border parts of the nickel-bearing cluster of Puumala amphibole also occurs. Here a sufficient amount of water was present, perhaps derived from surrounding sediments, and during local crustal movements was able to react with pyroxene and plagioclase to make amphibole. During this reaction quartz is liberated. Fig. 3 illustrates this reaction very well by showing a diopside crystal containing amphibole and small grains of quartz. The main mineral assemblage in the border parts of the cluster corresponds to the amphibolite facies. The association of ore minerals in these parts is similar to the association in the inner parts of the cluster, already shown to correspond well with the conditions governing the amphibolite facies.

In the parts closest to the surrounding rocks garnet is also present (see p. 4), as it is in the surrounding rocks themselves. Because of the smallness of the grains of garnet in the subsilicic rock, its accurate determination was difficult. Most probably it is almandite.

Almandite is stable in the amphibolite facies (almandite-diopside-hornblende subfacies). In the Puumala rock the aluminium needed for the formation of garnet may have been incorporated into the system from the aluminous schists of the surrounding country rocks.

Marcasite occurs among the ore minerals of the Puumala cluster as an alteration product of pyrrhotite. Marcasite either as a hypogene replacement mineral or as a supergene alteration product of pyrrhotite in an acid environment is known to occur (Edwards 1947, p. 100 and 129; Ramdohr 1950, pp. 603—609). At a temperature above 350°C marcasite is unable to exist. The marcasite corresponding to the hydrothermal range and brought into ore deposits by ascending solutions is rare and has not been found at Puumala.

Here the marcasite is supergene. It always occurs together with limonite. Consequently, the marcasite of Puumala must represent a very low facies, corresponding to the weathering of rock-making minerals.

The epidote-amphibolite facies (the saussurite facies) is represented in the cluster of Puumala only by hysterogenic minerals, as a weak saussuritization of plagioclase (see p. 4). Marcasite of ascending nature is absent, and no evidence characteristic of the conditions of the epidote-amphibolite facies has been observed among the ore minerals. The weak chloritization of biotite and pyroxene is also a hysterogenic mark of the said facies.

CONCLUSIONS

As a result of the arguments given in the foregoing discussion, the metamorphism within the nickel-bearing cluster of Puumala may be supposed, to have proceeded, as follows:

1. The nickel-bearing rock of the cluster is magmatic, and its crystallization occurred under conditions corresponding to the dipsenic amphibolite facies, characterized by the mineral association, diopside-biotite-plagioclase.
2. During the movements, the marginal parts of the cluster obtained water from the surrounding schists, and the typical mineral assemblage of the amphibolite facies was formed.
3. Facies lower than the amphibolite facies are present only in the form of hysterogenic minerals.
4. The chief mineralization of the rock of the cluster took place under orthomagmatic and pneumatolytic conditions (formation of ilmenite, magnetite, pyrrhotite, and pentlandite). The ore minerals of this stage of mineralization may be considered to belong to the pyroxene-hornfels facies.
5. The exsolution of pyrrhotite and pentlandite (flames of pentlandite) and the deposition of chalcopyrite correspond to a transitional stage between pneumatolytic and hydrothermal conditions, overlapping with the amphibolite facies.
6. Marcasite and limonite are supergene alteration products formed under conditions corresponding to the weathering of the rock-making minerals.

ACKNOWLEDGMENTS

During the field work the present author was assisted by Mr. L. Hyvärinen. Professor P. Eskola read the manuscript and gave useful advice. For all these courtesies, the author wishes to express his gratitude.

REFERENCES

EDWARDS, A. B. (1947) Texture of the ore minerals and their significance. Proc. Australasian Inst. Mining and Met., Melbourne.

ESKOLA, P. (1920) The mineral facies of rocks. Norsk Geol. Tidsskr. 6, p. 143.

HAWLEY, J. E., COLGROVE, G. L., and ZURBRIGG, H. F. (1943) The Fe-Ni-S system. Econ. Geol. 38, p. 335.

HEINRICH, E. WM. (1946) Studies in the mica group: the biotite-phlogopite series. Am. J. Sci. 244, p. 836.

HEWITT, R. L. (1938) Experiments bearing on the relation of pyrrhotite to other sulphides. Econ. Geol. 33, p. 305.

KERR, PAUL F. (1924) A magmatic sulphide ore from Chicagof Island, Alaska. Econ. Geol. 19, p. 369.

OVERBECK, R. M. (1917) Mineral resources of Alaska. — U. S. Geol. Survey. Bull. 692, p. 125.

PARRAS, KAUKO (1941) Das Gebiet der Pyroxen führenden Gesteine im Uusimaa in Südfinnland. Geol. Rundschau 32, p. 484.

RAMDOHR, P. (1950) Die Erzmineralien und ihre Verwachsungen. Berlin.

ROSENQVIST, IVAN TH. (1952) The Metamorphic facies and the feldspar minerals. Univ. Bergen Årbok 1952, Naturvitenskap. Rekke No. 4.

SCHNEIDERHÖHN, HANS (1941) Lehrbuch der Erzlagerstättenkunde. Erster Band. Die Lagerstätten der magmatischen Abfolge. Jena.

SCHOLTZ, D. L. (1933) The magmatic nickeliferous ore deposits of East Griqualand and Pondoland. Univ. Pretoria Publ., part. 1.

TURNER, FRANCIS J. (1948) Mineralogical and structural evolution of the metamorphic rocks. Geol. Soc. Amer. Mem. 30.

VÄHÄTALO, VEIKKO (1951) Outokummun malmin hivenmineraaleista. Geologi (Finland) 3, p. 21.

—»— (1953) On the geology of the Outokumpu ore deposit in Finland. Bull. Comm. géol. Finlande 164.

WINCHELL, ALEXANDER N. and WINCHELL, HORACE (1951) Elements of optical mineralogy. Part II. Description of minerals. Fourth Edit. New York, London.

YODER, HATTEN S. JR. (1952) The $MgO—Al_2O_3—SiO_2—H_2O$ system and the related metamorphic facies. Am. J. Sci. Bowen Volume, p. 569.

2

ON THE COMPOSITION OF NATURAL MELILITES[1]

BY

K. J. NEUVONEN

Geological Survey of Finland, Helsinki

ABSTRACT

The refractive indices, specific gravity, and unit cell dimensions of six chemically analyzed natural melilites are determined and compared with those reported for artificial melilite preparations. The mode of occurence of sodium in natural melilites is discussed and a method is proposed for evaluating the composition of the natural melilites by means of refractive indices and powder diffraction analysis.

CONTENTS

INTRODUCTION

Natural melilite minerals usually contain considerable amounts of sodium. The composition of the sodium end member, however, has remained unsolved, although different kinds of formulas have been suggested. The most generally accepted formula has been that proposed by Winchell (1924) as $Na_2Si_3O_7$. While studying the rule of alkalis in melilites, Goldsmith (1948) was not able to synthesize any artificial melilite corresponding to Winchell's formula. Such a melilite was supposed to be stable, however, in solid solutions containing adequate amounts of gehlenite because Goldsmith had

---

[1] Received December 28, 1954

crystallized a melilite with 3.85 % Na_2O from a mixture of $Na_2Si_3O_7$ and $Ca_2Al_2Si\ O_7$ (gehlenite).

Quoting Goldsmith's experimental results, Winchell (1951) used the formula $CaNaAlSi_2O_7$ for the sodium end member of the melilite family. Recently, this composition was crystallized by Nurse and Midgley (1953) as a phase very similar to the common melilites. They were also able to show that this material formed crystalline solid solutions both with åkermanite and with gehlenite in any ratio.

As pointed out by Goldsmith (1948), the composition $CaNaAlSi_2O_7$ is nothing else but $Na_2Si_3O_7 + Ca_2Al_2SiO_7$. From the structural point of view these formulas differ, however. According to the crystal structure of melilite studied by Warren (1930) and by Smith (1953), silicon occupies sites inside the linked Si_2O_7 tetrahedron pairs while magnesium ($+ Fe^{\cdot\cdot}$ and Mn) occupies a four-fold position at the origin of the unit cell. In gehlenite, the aluminium replaces Si and Mg in the ratio one to one. Consequently, the arrangement of the atoms in the sodium-melilite, if assumed to be composed of $Na_2Si_3O_7 + Ca_2Al_2SiO_7$ is as follows:

$$\begin{array}{c|c|c|c}
Ca_2 & Al & Al\ Si & O_7 \\
Na_2 & Si & Si\ Si & O_7
\end{array}$$

This arrangement would mean substitution between Si and Mg in alkali-bearing melilites. Such a replacement is unlikely, as pointed out by several writers.

The end member formula $CaNaAlSi_2O_7$ does not have this disadvantage since the arrangement

$$\begin{array}{c|c|c|c}
CaNa & Al & Si\ Si & O_7
\end{array}$$

indicates substitutions which are common in natural melilites.

Compiling the cations in three different groups according to the structural sites, the general formula of melilites can be written as

$$(Ca,Na,K)_2\ (Mg,Mn,Fe,Al)_1\ (Si,Al)_2\ O_7$$

The distribution of Al is of interest. In pure gehlenite or in any mixture free of alkalis the Al substitutes Si and Mg in equal amounts. In an alkali-bearing melilite, however, all the Al atoms of the alkali melilite component are subsituting Mg. Consequently, if a chemical analysis of a natural melilite is recalculated to seven oxygen atoms, the mole fraction of the alkali melilite is determined by the number of aluminium atoms left over when the gehlenite molecule is formed. Since gehlenite is the only end member in the melilite family in which Al substitutes Si, the number of Al atoms needed to build gehlenite is equal to $= 2\ (2 - Si)$. The number of Al atoms left over for alkali melilite is therefore $= Al - 2\ (2 - Si)$. On the other hand, the mole fraction of the alkali melilite in the solid solution is given by the number of alkali atoms present.

Consequently, if the occurrence of alkalis in the melilite minerals is accompanied by the substitution of Mg with Al, two independent ways of calculating the alkali melilite content are available. If these two methods yield similar figures, it seems justified, on the other hand, to conclude that the alkali content of the melilites may be explained by means of the hypothetical end member $Ca(Na,K)AlSi_2O_7$.

CHEMICAL COMPOSITION

For calculations, six natural melilite samples were purified by centrifuging in Clerici's solution. Pure samples were chemically analyzed by Dr. H. B. Wiik of the Geological Survey of Finland. The analyses are reproduced in Table 1. This table lists also the number of atoms based on recalculation of the analyses when the number of oxygen atoms is scaled to seven. Water, TiO_2, and P_2O_5 were neglected. They were assumed to derive from impurities observed, i.e. from sphene, apatite, and alteration products.

Sample 1. Melilite from a melilite basalt, Honolulu, Hawaii. The original rock sample was received by the courtesy of Dr. C. D. Wentworth. Melilite occurs in the rock as irregular or lathy crystals in a matrix including olivine as phenocrysts. A weak zoning is observable as a higher birefringence at the margin of the crystals. The rock has been described by Cross (1915).

Sample 2. Melilite from a melilite basalt from Mt. Elgon, Uganda, East Africa. The rock sample was kindly placed at my disposal by Dr. N. L. Bowen. Melilite in this rock occurs as phenocrysts with olivine and pyroxene in a glassy amygdaloidal matrix. The crystals have invariably an outer zone of higher birefringence.

Sample 3. Melilite from the classical Uncompahgrite locality, Iron Hill, Colorado, described by Larsen and Hunter (1914). The specimen was kindly offered by D. W. F. Foshag from the U.S. National Museum collections.

Sample 4. This melilite derives from a basalt flow in Katunga, Ankole, Uganda, East Africa. The mineralogy of this rock has been described by Holmes (1937).

Sample 5. The sample was purified from a »gehlenite» specimen obtained through the kindness of Dr. C. Frondel. The sample originates from Fassathal in Tyrol. The melilite in the sample is very homogeneous and the only major impurities are calcite and alteration products along the cracks.

Sample 6. This melilite derives from a massive melilite (»gehlenite») specimen purchased from Ward's Natural Science Establishment, Inc., originating from Luna County, New Mexico.

Table 2 gives chemical formulas of the analyzed minerals based on the number of atoms listed in Table 1. In calculating the formulas, the sum of Si and Al within the third brackets is taken equal to 2.000. The sum of the atoms in the first and second parentheses, theoretically, should equal 2 and unity, respectively. The deviations observable are not serious. The maximum deficiency is found in sample 5, the sum of Mg,Mn,Fe, and Al being low, about 5 per cent.

Table 1. Chemical composition of the purified melilite samples. Analyzed by Dr. H. B. Wiik, Geological Survey of Finland

	1 Sugar Loaf Hawaii		2 Mt. Elgon Uganda		3 Iron Hill Colorado		4 Katunga Uganda		5 Fassathal Tyrol		6 Luna County New Mexico	
	wt %	atoms	wt %	atoms	wt %	atoms	wt %	atoms	wt %	atoms	wt %	atoms
SiO_2	42.81		42.86		43.27		43.12		30.09		32.60	
Si		1.964		1.972		1.971		1.987		1.409		1.496
Al_2O_3	6.59		5.02		6.44		4.59		21.67		18.10	
Al		0.356		0.272		0.346		0.249		1.196		0.979
Fe_2O_3	1.90		0.26		0.97		0.00		1.36		0.95	
Fe ···		0.065		0.009		0.033		0.000		0.048		0.033
FeO	3.06		2.99		2.32		2.13		2.14		0.92	
Fe ··		0.117		0.115		0.088		0.082		0.084		0.035
MnO	0.08		0.08		0.16		0.09		0.04		0.02	
Mn		0.003		0.003		0.006		0.004		0.002		0.001
MgO	7.48		9.20		7.90		10.18		3.87		6.91	
Mg		0.511		0.631		0.536		0.699		0.270		0.472
CaO	33.27		34.90		34.76		35.75		38.36		40.08	
Ca		1.633		1.719		1.695		1.765		1.923		1.969
Na_2O	3.75		3.68		3.59		2.16		0.75		0.18	
Na		0.333		0.310		0.317		0.166		0.068		0.016
K_2O	0.33		0.42		0.20		0.69		0.16		0.21	
K		0.019		0.024		0.011		0.039		0.010		0.012
TiO_2	0.12		0.06		tr.		0.11		tr.		0.00	
P_2O_5	n. d.		n. d.		0.00		0.14		0.00		0.00	
H_2O+	0.43		0.22		0.28		0.81		1.64		0.48	
H_2O-	0.00		0.00		0.00		0.06		0.00		0.00	
Total	99.82		99.69		99.89		99.83		100.08		100.45	

As mentioned many of the samples show zoning indicating variation in the chemical composition. As the chemical analyses represent average composition of such samples, the optical properties of the minerals were also measured to correspond to the »average» of the sample. This was done by measuring a large number of grains. The indices of refraction and the densities are listed in Table 3.

Table 2. The chemical formulas of the analyzed melilites

No.	Chemical formulas
1	$\underset{1.985}{(Ca_{1.633}Na_{.333}K_{.019})}\underset{1.016}{(Mg_{.511}Mn_{.003}Fe^{\cdot\cdot}_{.117}Fe^{\cdot\cdot\cdot}_{.065}Al_{.320})}\underset{2.000}{(Si_{1.964}Al_{.036})}O_7$
2	$\underset{2.053}{(Ca_{1.719}Na_{.310}K_{.024})}\underset{1.002}{(Mg_{.631}Mn_{.003}Fe^{\cdot\cdot}_{.115}Fe^{\cdot\cdot\cdot}_{.009}Al_{.244})}\underset{2.000}{(Si_{1.972}Al_{.028})}O_7$
3	$\underset{2.023}{(Ca_{1.695}Na_{.317}K_{.011})}\underset{0.981}{(Mg_{.536}Mn_{.006}Fe^{\cdot\cdot}_{.088}Fe^{\cdot\cdot\cdot}_{.033}Al_{.317})}\underset{2.000}{(Si_{1.971}Al_{.029})}O_7$
4	$\underset{1.960}{(Ca_{1.765}Na_{.166}K_{.039})}\underset{1.021}{(Mg_{.699}Mn_{.004}Fe^{\cdot\cdot}_{.082}Fe^{\cdot\cdot\cdot}_{.000}Al_{.236})}\underset{2.000}{(Si_{1.987}Al_{.013})}O_7$
5	$\underset{2.001}{(Ca_{1.923}Na_{.068}K_{.010})}\underset{0.949}{(Mg_{.270}Mn_{.002}Fe^{\cdot\cdot}_{.084}Fe^{\cdot\cdot\cdot}_{.048}Al_{.605})}\underset{2.000}{(Si_{1.409}Al_{.591})}O_7$
6	$\underset{1.997}{(Ca_{1.969}Na_{.016}K_{.012})}\underset{1.016}{(Mg_{.472}Mn_{.001}Fe^{\cdot\cdot}_{.035}Fe^{\cdot\cdot\cdot}_{.033}Al_{.475})}\underset{2.000}{(Si_{1.496}Al_{.504})}O_7$

Table 3. Optical properties and densities of the analysed melilite samples

Sample no.	Index of refractivity for Na-light			Density $d_{4^\circ}^{20^\circ}$
	ω	ε	$\omega-\varepsilon$	
1	1.638	1.630	0.0079	2.98
2	1.635	1.633	0.0025	2.99
3	1.630	1.627	0.0030	2.980
4	1.638	1.635	0.0028	2.981
5	1.659	1.656	0.0035	3.066
6	1.655	1.654	0.0015	3.018

In Fig. 1, the mole fraction of the alkali melilite, as determined by the number of sodium and potassium atoms, is plotted against the mole fraction of the alkali melilite given by the excess of aluminium. The straight line in the diagram represents the theoretical requirement that the mole fractions so determined are equal. Since the role of ferric iron in melilites is similar to that of aluminium, the ferric iron is added to aluminium in the calculations. The black circles represent the melilite analyses listed in Table 1. All these points lie close to the line of the theoretical requirement. Only the dot representing sample 2 deviates notably from the line since the sum of alkali atoms is greater than the excess of aluminium. As seen in Table 2, the sum of calcium, sodium, and potassium atoms in sample 2 is large by 2.5 %, viz. 2.053. This gives reason to suspect that the deviation observed is caused by an analytical error or by impurities present. It seems justified, consequently, to explain the alkalis in melilite minerals by means of the end members $CaNaAlSi_2O_7$ and $CaKAlSi_2O_7$.

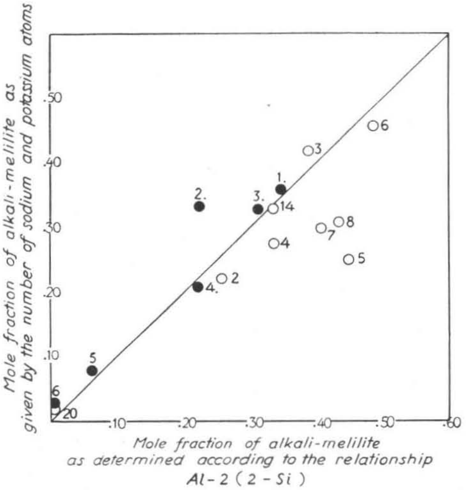


Fig. 1. Mole fraction of alkali melilite determined by the number of alkali atoms plotted against mole fraction of alkali melilite given by the number of aluminium atoms not bonded in the gehlenite component. The black cirkles are calculated from the analyses in Table 1. The open circles are calculated from the list of melilite analyses of Berman (1929). The numbers on the circles refer to the sample numbers in Table 1 and in Berman's list.

Additional melilite analyses are presented in the diagram of Fig. 1 (open circles), calculated from the list of melilite analyses given by Berman (1929). From Berman's analyses, only those were used which are complete, i. e. in which both FeO and Fe_2O_3 are given. Also these analyses have a clear tendency to become distributed about the straight line. Most of the open circles, however, are under the line, indicating deficiency of alkalis. As indicated by the investigation of the precision and accuracy in the analysis of silicates (Fairbairn et al, 1951), the precision of the chemical method in alkali determination is inferior to spectrochemical and flame photometer analysis, and yields results which tend to be low. The alkalis in the old analyses listed by Berman are naturally determined by the chemical method. This fact may explain the deficiency of alkalis in these analyses and the distribution of the circles below the straight line in the diagram in Fig. 1.

OPTICAL PROPERTIES

As far as the chemical composition of the analyzed melilite samples is concerned, it seems justified to explain the alkali content found by means of the end member $Ca(Na,K)AlSi_2O_7$. It will be interesting, however, to compare the optical properties measured with those reported for the synthetical preparations. Nurse and Midgley (1953) give data for the variation of the indices of refraction in the binary system sodium-melilite—åkermanite and in the binary system sodium-melilite—gehlenite. The optical properties in the system åkermanite—gehlenite are reported by Ferguson and Buddington (1920). No data, however, are available for the ternary system sodium-melilite—åkermanite—gehlenite. Based on the values given for the associated binary system, tentative diagrams of the variation of the ω and

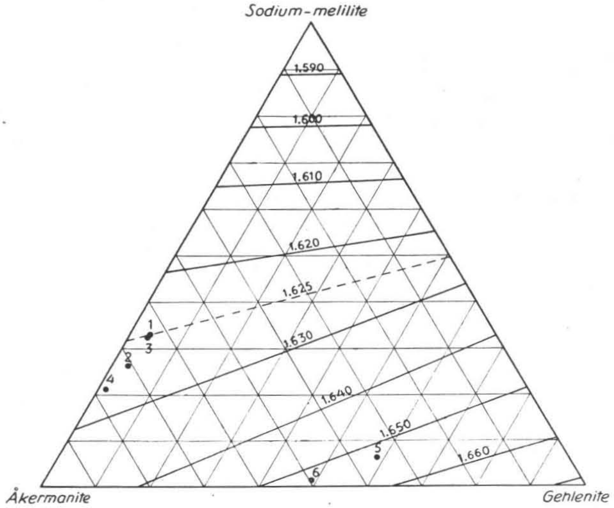


Fig. 2. Variation of the ω-index of the optical refractivity in the system åkermanite — gehlenite — sodium-melilite (approximately).

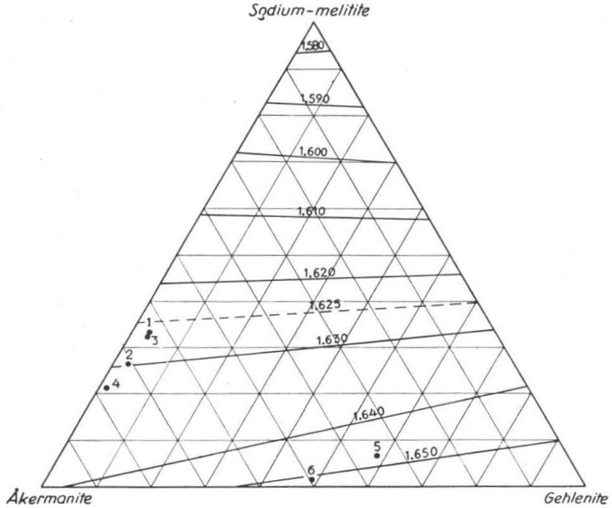


Fig. 3. Variation of the ε-index of the optical refractivity in the system åkermanite — gehlenite — sodium-melilite (approximately).

ε indices are given in Figs. 2 and 3, respectively. The compositions of the analyzed melilite samples of this paper are indicated as black dots in the diagrams. In calculating the composition of the samples for the diagrams, the mole fraction of åkermanite is obtained as the sum of the atoms Mg,Mn, and Fe˙˙, and the mole fraction of gehlenite is taken as (2 — Si). The mole fraction of sodium-melilite is read from the diagram in Fig. 1. The mole fractions so obtained are corrected to sum up to 1.00.

From the diagrams in Fig. 2 and in Fig. 3, the indices of refraction of samples 1 — 6 of Table 1 may be read with a result given in Table 4.

Table 4. Indices of refraction of the analyzed melilite samples read from the diagrams in Fig. 2 and Fig. 3

Sample No.	ω	ε	ε—ω
1	1.625	1.627	+0.002
2	1.627	1.630	+0.003
3	1.625	1.627	+0.002
4	1.628	1.632	+0.004
5	1.652	1.649	—0.003
6	1.652	1.650	—0.002

It is readily observed that the values in Table 4 differ greatly from those actually measured and given in Table 3. All the samples, however, contain remarkable amounts of iron and manganese, which increase the measured indices. By means of the data given by Nurse and Midgley (1953) for the system gehlenite—iron-gehlenite, and with the optical values reported by Bowen, Schairer, and Posnjak (1933) for the iron-åkermanite, it is possible to correct the measured indices for iron-free compositions. In Table 5, the corrected refractive indices and the corresponding recalculated compositions are listed.

Table 5. Hypothetical indices of refraction of the analyzed melilite samples recalculated free of iron (and manganese)

No	Composition in mole fractions			Refractive indices		
	åker.	gehl.	sod. mel.	ω	ε	ε—ω
1	0.64	0.03	0.33	1.628	1.622	—0.006
2	0.71	0.03	0.26	1.628	1.628	0.000
3	0.64	0.03	0.33	1.622	1.621	—0.001
4	0.78	0.01	0.21	1.633	1.633	0.000
5	0.35	0.58	0.07	1.652	1.650	—0.002
6	0.50	0.49	0.01	1.651	1.651	0.000

When making the corrections, the role of potassium found in the analyzed samples was left out of consideration. The effect of manganese was assumed to be the same as that of ferrous iron. The mutual agreement, observable between the values in Table 4 and in Table 5, is as good as can be expected. It may be stated, consequently, that the optical properties of the analyzed melilite samples are in harmony with the hypothesis that $CaNaAlSi_2O_7$ exists as an end member in the natural melilites.

THE UNIT CELL DIMENSIONS

Using Copper Kα-radiation, powder diffraction patterns were obtained for the analyzed melilite samples on the Norelco recording X-ray spectrometer. The d-values were corrected utilizing silicon as internal standard. As mentioned, some of the samples show zoning under the microscope indicating variation in the chemical composition. Consequently, the »peaks» on the corresponding charts are not as sharp as desired. For calculations, d-values measured as weighed averages are employed in such a case. The dimensions determined are given in Table 6.

Table 6. Unit cell dimensions of the analyzed melilites

Sample No.	Mole fraction of åkermanite	a_0	c_0	V_0
1	0.64	7.781	5.024	304.2
2	0.71	7.787	5.026	304.8
3	0.64	7.772	5.015	303.0
4	0.78	7.805	5.016	305.6
5	0.35	7.738	5.045	302.1
6	0.50	7.760	5.048	304.0

Ervin and Osborn (1949) and Neuvonen (1952) presented diagrams which give unit cell dimensions and other X-ray data as a function of the composition in the system åkermanite—gehlenite. The åkermanite content of the present natural melilite samples, as determined by the a_0-values listed and by the spread of the pairs (213) — (521) and (213) — (440), are given in Table 7.

Table 7. Åkermanite content of the melilite samples determined by means of the powder diffraction method

Method	Åkermanite content in mole fractions					
Sample No.	1	2	3	4	5	6
a_0	0.64	0.68	0.58	0.80	0.35	0.50
(213) — (521)	0.64	0.75	0.65	0.83	0.40	0.51
(213) — (440)	0.67	0.70	0.64	0.84	0.35	0.53
Chemically	0.64	0.71	0.64	0.78	0.35	0.50

As shown in Table 7, the åkermanite contents determined from the X-ray diffraction agree well with the values calculated from the chemical analyses. This was unexpected since the diagrams used are actually for pure åkermanite—gehlenite mixtures, and most of the natural samples used are rich in alkalis and contain considerable amounts of iron as well. It can be calculated from the X-ray diffraction values given by Muan and Osborn (1951) for the iron-åkermanite that the effect of ferrous iron on the unit cell dimensions of åkermanite is negligible. Consequently, the only possibility to explain the agreement observed is to assume that unit cell dimensions of sodium-melilite (and iron-gehlenite) are very similar to those of gehlenite.

The unit cell of sodium-melilite measured by Nurse and Midgley (1953), viz. $a_0=8.511$ and $c_0=4.809$, differ greatly from those of gehlenite ($a_0=7.685$ and $c_0=5.076$, Neuvonen, 1952). This fact is in a disagreement with the statement presented in the foregoing. A very great deviation from Vegard's law might explain this discordance, but only small deviations have been reported in stable solid solutions (Barth, 1930). The unit cell volume of the sodium-melilite, 348.3 cubic Å, based upon the cell dimensions of Nurse and Midgley, differs also from those of all other melilites. The calculated specific gravity based on these dimensions is about 2.45. This value is by far lower than what one could expect on the basis of the measured densities of the sodium-rich melilites. Re-examination of the unit cell dimensions of the synthetic sodium-melilite is therefore worthwhile.

COMPOSITION OF MELILITES BY OPTICAL AND X-RAY MEASUREMENTS

As was noted in the foregoing, it is possible to evaluate the åkermanite content of a natural melilite sample by means of powder diffraction measurements. This method might have a general application. On the other hand, it may be possible, by means of the optical properties, to determine the ratio of gehlenite to sodium-melilite using the diagrams in Figs. 2 and 3, if the åkermanite content is known. This procedure requires, however, that the samples are free of iron or that the refractive indices measured for natural iron-bearing melilites could be corrected to corresponding iron-free compositions.

Such a correction could be based on an hypothetical substitution of iron-åkermanite with åkermanite and iron-gehlenite with gehlenite. The change of the refractivity involved in these substitutions is proportional to the difference of the indices of the counterparts and to the mole fraction of the iron-bearing end members in the sample. Because the difference of the ω-indices of iron-åkermanite and åkermanite ($= 0.058$) is about equal to the difference between the ω-indices of the iron-gehlenite and gehlenite

(= 0.057), a correction of the ω-index can be performed according to the following equation:

$$\omega_{\mathrm{corr.}} = \omega_{\mathrm{measured}} - 0.057\ (\mathrm{Fe})$$

in which (Fe) denotes the sum of the mole fraction of iron-åkermanite and of mole fraction of iron-gehlenite in the sample. The mole fraction of the iron-melilites (Fe) may be obtained, for a close approximate, by dividing the percentage of total iron in the sample, expressed as FeO, by the figure 26.0. This empirical rule holds for iron contents found in the natural melilites. To determine the total iron, only chemical methods are available. Fortunately, iron is easily and rapidly determined in several ways. The spectrochemical (Dennen et al, 1951) and photometrical (Shapiro and Brannock, 1952) methods are most convenient.

The determination of the composition of an unknown melilite sample proceeds as follows:

1) ω-index of the mineral is measured.

2) X-ray powder diffraction pattern of the mineral is obtained using Copper Kα-radiation. The åkermanite content of the sample is evaluated using the diagrams given by Ervin and Osborn (1949) or those given by Neuvonen (1952). The åkermanite content may also be obtained from the following equations:

a) $(\text{åk}) = 4.11 - 0.51 \cdot \triangle_{(213)-(440)}$
b) $(\text{åk}) = 4.08 - 0.57 \cdot \triangle_{(521)-(213)}$
c) $(\text{åk}) = 1.31 - 0.82 \cdot \triangle_{(310)-(002)}$

in which (åk) means the mole fraction of åkermanite and $\triangle$ denotes the distance of the two lines indicated, expressed in degrees of 2Θ. The various peaks are located on the pattern in the following way:

(440)	$2\Theta =$ 67.5 — 69.0	weak
(521)	66.8 — 68.2	medium
(213)	61.1 — 61.4	medium
(310)	36.2 — 37.0	strong
(002)	35.3 — 35.9	weak

3) Total iron content of the sample is determined and expressed as FeO. The mole fraction of the iron-melilite is

$$(\mathrm{Fe}) = \frac{\mathrm{FeO}\ \%}{26.0}$$

The corrected ω-valve is calculated accordingly.

4) From the diagram in Fig. 2, a point is traced at which the åkermanite content equals that determined and at which the $\omega_{\mathrm{corr.}}$ matches that of the diagram.

The mole fraction of åkermanite so obtained includes iron-åkermanite, and the mole fraction of gehlenite includes the iron-gehlenite.

If the composition of the melilite samples described in the present paper is determined by means of the method described, it yields the results given in Table 8.

Table 8. Composition of the melilite samples determined by the method proposed

Sample No.	Composition in mole fractions		
	åker.	gehl.	sod. mel.
1	0.65	0.06	0.29
2	0.73	0.05	0.22
3	0.62	0.00	0.38
4	0.82	0.07	0.11
5	0.37	0.56	0.07
6	0.51	0.48	0.03

When the values in Table 8 are compared with those calculated from the chemical analyses and reproduced in Table 5, it is readily observed that the agreement is very good. The largest deviation is found with sample 4, in which case the mole fraction of sodium-melilite determined by the proposed method is low by 10 %. This sample, however, is rich in potassium, which might be the reason for the said deviation. The sample derives from a unusual type of rock very rich in potassium, giving reason to believe that melilites as rich in potassium as this particular sample are very scarce.

Melilite from Capo di Bove, Italy, might serve as an additional example. For this mineral, Smith (1953) gives $a_0 = 7.789$ and $\omega = 1.638$. According to the old analyses of this material (Berman, 1929) it contains ca. 10.5 % Fe_3O_2. The a_0-value measured indicates that the fraction of åkermanite in the sample is about 0.65. It can be seen from Fig. 2 that the lowest possible ω-index corresponding to such an åkermanite content (0.65 åker. + 0.35 sod. mel.) is 1.624. If the iron correction is made for the sample, however, the measured index should drop down to 1.617, indicating that the iron content reported is too high. This is in good agreement with the statement made by Smith on the basis of the structure amplitudes observed, that the Capo di Bove melilite can have only little if any Fe in the structure.

CONCLUDING REMARKS

The question of the end members in a mineral family is mostly academic. More exactly, the mode of occurrence of different ions in a mineral is explained by ionic substitutions in the structure. From this point of view, sodium-melilite, $CaNaAlSi_2O_7$, is but diadocic substitution of the cations Ca-Al-Al (in gehlenite) or of the cations Ca-Mg-Si (in åkermanite) with the cations Na-Al-Si in sodium-bearing melilites. The chemical composition and the optical properties of the analyzed natural melilite samples indicate that these substitutions take place in the natural melilite minerals. The unit cell dimensions of the melilites studied differ notably, however, from the values expected in the light of these substitutions. For further conclusions, additional measurements of cell dimensions are needed.

The method proposed for specification of composition is not meant to be exact. Before a precision method can be created, the variation of the optical properties in the ternary system åkermanite—gehlenite—sodium-melilite must be accurately known. The effect of iron, manganese, sodium, and potassium on the unit cell should also be studied before more dependable analyses are possible.

Acknowledgments— The writer wishes to thank all the persons who put melilite samples at his disposal, many of them not mentioned because of the difficulties in purifying the specimens. The writer is also grateful to Dr. H. B. Wiik for the chemical analyses of the samples and Miss Thyra Åberg for drawing the figures. Thanks are due, furthermore to Prof. Th. G. Sahama for constructive criticism of the manuscript.

REFERENCES

Barth, Tom F. W. (1930) Optical properties of mixed crystals. Am. J. Sci. 19, p. 135.

Berman, H. (1929) Composition of the melilite group. Am. Min. 14, p. 389.

Bowen, N. L., Schairer, J. F., and Posnjak, E. (1933) The system, CaO-FeO-SiO_2. Am. J. Sci. 26, p. 193.

Cross, W. (1915) Lavas of Hawaii and their relations. U.S. Geol. Survey, Profess. Paper 88.

Dennen, W. H., Ahrens, L. H., and Fairbairn, H. W. (1951) Spectrochemical analysis of major constituent elements in rocks and minerals. U.S. Geol. Survey Bull. 980, part 3.

Ervin, Guy and Osborn, E. F. (1949) X-ray data on synthetic melilites. Am. Min. 34, p. 717.

Fairbairn, H. W. et al (1951) A cooperative investigation of precision and accuracy in chemical, spectrochemical, and modal analysis of silicate rocks. U.S. Geol. Survey Bull. 980.

Ferguson, J. B. and Buddington, A. F. (1920) The binary system akermanite-gehlenite. Am. J. Sci. 50, p. 131.

Goldsmith, J. R. (1948) Some melilite solid solutions. J. Geol. 56, p. 437.

Holmes, A, (1937) The petrology of Katungite. Geol. Mag. 74, p. 200.

Larsen, E. S. and Hunter, J. F. (1914) Melilite and other minerals from Gunnison County, Colo. Journ. Wash. Acad. Sci. 4, p. 473,

Muan, A. and Osborn, E. F. (1951) Phase relations in the system $2CaO.SiO_2$-CaO. SiO_2-$2CaO.Al_2O_3.SiO_2$-FeO. Yearbook Am. Iron and Steel Institute. p. 325.

Neuvonen, K. J. (1952) Thermochemical investigation of the åkermanite-gehlenite series. Bull. Comm. géol. Finlande 158.

NURSE, R. W. and MIDGLEY, H. G. (1953) Studies on the melilite solid solutions. J. Iron and Steel Institute 174, p. 121.

SHAPIRO, L. and BRANNOCK, W. W. (1952) Rapid analysis of silicate rocks. U.S. Geol. Survey Circular 165.

SMITH, J. V. (1953) Reexamination of the crystal structure of melilite. Am. Min. 38, p. 643.

WARREN, B. E. (1930) The structure of melilite, $(Ca, Na)_2(Mg, Al) (Si,Al)_2O_7$. Zeitschr. Krist. 74, p. 131.

WINCHELL, A. N. (1924) The composition of melilite. Am. J. Sci. 8, p. 375.

WINCHELL, A. N. (1951) Elements of optical mineralogy. Part. II. Fourth edition. New York.

3

SEDIMENTARY TRANSPORTATION IN KARELIAN QUARTZITES[1]

BY

TOIVO MIKKOLA

Suomen Malmi Osakeyhtiö, Otaniemi, Finland

ABSTRACT

The paper deals with the direction of sedimentary transportation in the Finnish pre-Cambrian as determined from current bedding.

The direction of sedimentary transportation has been of current interest in Finland, expressly as regards Quaternary sediments. It would be interesting to know, however, if there are possibilities to study this problem in the pre-Cambrian. The present investigation is the first effort to determine the direction of transportation in the ancient sediments of Finland. The ridge of Karelian quartzite at Selkie, East Finland, was chosen for this purpose. The work was inspired by Dr. E. Cloos and Dr. F. J. Pettijohn.

In strongly deformed rocks the only applicable method seems to be that of current beds[2]. The current beds dip downstream. The average direction of stream is the horizontal line perpendicular to the intersection of current and horizontal beds. If the strata are folded, it is possible to »rotate them back» to the original position by the aid of Schmidt's net. If the strata are only gently folded the error is minimal. In strongly deformed rocks, especially if the fold axis is nearly vertical, as sometimes happens in the Finnish pre-Cambrian, the error can be considerable. The line of intersection has rotated around the vertical axis and it is difficult to ascertain its original position. The error is less if the axis is not vertical but dips, as in the Karelians, between horizontal and about 20°. If the folding is regular, it is possible to calculate the maximum error. If the dip of fold axis is a, the dip of bedding b, and the error v, then

$$\sin v = \frac{\sin a}{\sin b}\sqrt{1 - \frac{\mathrm{tg}^2 a}{\mathrm{tg}^2 b}} - \frac{\mathrm{tg}\, a}{\mathrm{tg}\, b}\sqrt{1 - \frac{\sin^2 a}{\sin^2 b}}$$

---

[1] Received December 31, 1954.

[2] An extensive list of references is to be found, for example, in Potter, P. and Olson, J. (1954) Variance components of cross-bedding direction. J. Geology, 62.

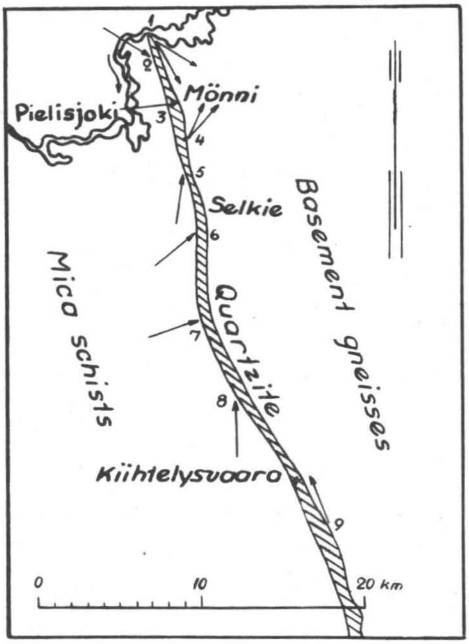


Fig. 1. Map of Selkie. The arrows show the direction of transportation in quartzite.

In general, however, the problem is not a mathematic but a tectonic one, and thus easiest to solve separately at each place. By using only outcrops where the axis is nearly horizontal, the results should be quite satisfactory. This method has been applied in the present study.

Current bedding from pre-Cambrian has been reported only occasionally. The idea was to look for it systematically in places where it has been found earlier. Starting from two occurrences reported by Messrs. Mikkonen and Laiti, I was able to find some more current-bedded outcrops. The current bedding is easiest to find in the conglomeratic part close to the base of quartzite. Measuring the strike and dip of current bedded rocks on ice-polished surfaces needs some practice. Because it is necessary to use the net, in any case, the apparent strike and dip are sufficient. Table 1 shows the results. In some cases the beds were distorted to a vertical position, but no overturned strata were found. The arrows on the map show the direction of the stream. As is to be seen on the map almost all the observations show an easterly direction of stream. No. 9 is approximately parallel to the strike of the fold axis. No directions deviating west from the strike were observed.

An average of ten measurements were made at each place. The deviation was less than 10°, except at two places where the results have been given separately. Only once did the angle between current-bedded and horizontal layer exceed the theoretical maximum value 36° in considerable amount. However, the difference does not exceed the possible maximum error in measurements, which certainly can be about 5° for each layer. It is evident, of course, that the angle has changed during the deformation, too, but no investigations in this respect were made.

The Karelian quartzite is a typical epicontinental or foreland formation. The geosynclinal core is believed to be somewhere in the west. Thus the average direction of stream, 60°, has been towards the foreland. However, the area investigated is too small and the observations too few to be of

Table 1

No. of observ. on map	Horizontal bed		Foreset bed		Angle of intersection	Direction of stream
	Strike	Dip	Strike	Dip		
1 a	180°	60° W	160°	40 ° SW	25°	120°
1 b	180	60 W	140	50 SW	35	150
2	170	80 W	130	50 SW	45	125
3	150	40 SW	120	20 SW	25	85
4 a	150	90	165	50 SW	40	45
4 b	150	90	165	70 SW	25	25
5	150	80 SW	165	70 SW	20	10
6	180	80 W	15	60 NW	25	55
7	10	75 NW	25	55 NW	25	70
8	30	50 NW	45	60 NW	15	360
9	160	75 SW	20	80 NW	40	340

general or statistical value. The topographic relations could have been exceptional in this area. In any event the observations made show that a more extensive study of this kind is possible and certainly would give important information concerning the evolution of pre-Cambrian relief and sedimentation.

4

ON THE CHEMICAL EQUILIBRIUM BETWEEN THE SULPHIDE MINERALS IN ORES [1]

BY

HEIKKI VÄYRYNEN

Technological Institute, Helsinki

ABSTRACT

New formulas were constructed for the dissociation pressures of pyrite and covellite and for the partial pressures of sulphur in the dissociated hydrogen sulphide at various temperatures. By means of these formulas, combined with the law of mass-action, the equilibrium points for the associations pyrite — pyrrhotite and covellite — chalcocite in the presence of hydrogen sulphide were calculated for various temperatures. Finally certain conclusions were drawn as to the genesis of the sulphide ores.

As early as 1912 Allen and his colleagues observed that the equilibrium point between pyrite and pyrrhotite in sulphur gas of 1 atm. pressure lay at the temperature of 665°C (Allen et al., 1912). In hydrogen sulphide gas of 1 atm. pressure the equilibrium point was then determined to be at 565°C. These experiments were performed in an open tube and were estimated to be of preliminary character only. In 1917 Allen and Lombard constructed an apparatus with a closed tube and succeeded in determing 10 points on the curve of dissociation of pyrite to pyrrhotite and sulphur (Allen and Lombard, 1917). These equilibrium points lying between the temperatures of 575° and 680°C were measured and the corresponding pressures of sulphur gas calculated.

For the purpose of controlling the consistency of the observations the authors have adopted the known theoretically constructed formula:

$$1) \qquad \log p = -\frac{A}{T} + B \log T + C.$$

Starting from three of the points determined at the temperatures 672, 645 and 595°C the three constants A, B, and C were calculated and the following figures obtained:

---

[1] Received December 29, 1954.

$$\begin{aligned} A &= 191\,942.61 \\ B &= -434.195 \\ C &= 1\,497.567 \end{aligned}$$

The rest of the equilibrium points determined likewise verify the formula arrived at.

It is a matter of fact that the association pyrite — pyrrhotite is quite common in the composition of the sulphide ores. Only a slight part of such occurrences, however, is estimated to be formed at temperatures as high as those mentioned, and in most cases even at a very much lower point. It is obvious that such high sulphur pressures must be rare in the processes of sulphide ore formation. In consequence the valuable researches cited have remained unimportant to the study of ore genesis. In the interest of their application to the common sulphide ores, the results of the experimental investigations referred to must be extrapolated down to a temperature of 300—400°C. An attempt at such an extrapolation has been made in the following (cf. Väyrynen, 1951).

In fact the formula 1) can also be called an interpolation formula between the temperature range of the determined points and the absolute zero because at this point ($T = 0$) the pressure $p = 0$ ($\log p = -\infty$). In the form obtained by the authors cited, however, both the terms dependent on temperature being negative, the formula cannot be applicable at higher temperatures. At a certain temperature the pressure, according to this formula, attains its maximum and then falls. This is certainly not what happens in the dissociation. The difference between this formula and the theoretically correct one, where the constant B is positive, decreases towards the absolute zero to 0, but there must be a difference also at the middle temperatures. Thus a new solution was sought.

The consistency of the determined values can be controlled also in the following way. The values of p and T of every determined point are placed in the equation 1) and the resulting 10 equations are combined two by two and C is eliminated. By so doing, 9 equations of the form $A + m\,B = n$ are obtained. If now n in these equations increases together with m, the constant B must be positive, while in the opposite case it is negative. It proved that the values of m plotted against those of n formed such scattered groups that the solution remained indefinite. Neither did a calculation of the constants A, B and C starting from three other points, than those used by the authors cited, give more acceptable results. The numbers arrived at were in every case too great for calculations. Therefore the value of the thermodynamic constant of oxygen, used sometimes in the literature also for sulphur, was accepted to the constant term C. Thus the following formula was arrived at:

$$\log p = -\frac{18\,850}{T} + 5.65 \log T + 2.8 \quad \text{(Fig. 1).} \tag{2}$$

This formula is in the following accepted to present the sulphur pressures, p, in atmospheres by the equilibrium of pyrite with pyrrhotite at various temperatures.

Among the volcanic emanations, however, hydrogen sulphide is commoner than sulphur. Being at higher temperatures partly dissociated in sulphur and hydrogen, it participates, of course, by its partial pressure of sulphur in the equilibrium between pyrite and pyrrhotite and in the following way:

$$FeS + H_2S \leftrightarrows FeS_2 + H_2.$$

Hydrogen is thereby liberated and this gas is in fact a usual constituent in the volcanic emanations. Owing to this fact the partial pressure of sulphur in the equilibrium between pyrite and pyrrhotite in the presence of hydrogen sulphide must on the one hand correspond to the dissociation pressure of pyrite and on the other hand to the partial pressure in the dissociated hydrogen sulphide (Väyrynen, 1935). This latter is ruled by the law of mass-action according to the formula:

$$[S_2] \cdot [H_2]^2 = K \cdot [H_2S]^2 \tag{3}$$

where the brackets denote respective concentrations or pressures and K is a constant, variable with the temperature. The concentration (partial pressure) of sulphur is thus directly proportional to the square root of the concentration (partial pressure) of hydrogen sulphide and inversely proportional to the square root of the concentration (partial pressure) of hydrogen. As this latter gas occurs abundantly among the volcanic gases, it is obvious that the concentration of sulphur must be very low also in the processes of ore formation in the presence of hydrogen sulphide. To be correct, the pressure of sulphur gas at a certain concentration depends also on its molecular composition, but at low pressures, as here, the molecule is S_2.

If now the dissociation constant, K, of the hydrogen sulphide, too, is known at all temperatures, it is in fact possible to transform all concentrations or partial pressures of sulphur by the pyrite—pyrrhotite equilibrium to corresponding concentrations or partial pressures of hydrogen sulphide or hydrogen according to formula 3). The temperature correlation of the dissociation constant, K, of hydrogen sulphide is, indeed, theoretically deducible, but the formula obtained in this way is not very accurately verified by the points determined experimentally, the deviations being 100 to 200°C (Breuner, 1909). Therefore new constants were calculated for this formula, too. The new formula attained is:

4) $$\log K = -\frac{7\,800}{T} + 2.84 \log T - 4.96.$$

The curve plotted (Fig. 1) discloses that hydrogen sulphide is, also at higher temperatures, a relatively resistent compound and its dissociation increases comparatively slowly with rising temperature. Consequently the partial pressure of sulphur, being at lower temperatures higher than the dissociation pressure of pyrite, is at higher temperatures much lower than this latter pressure. Thus the curves representing the dissociation pressures of pyrite and the partial pressures of sulphur in dissociated hydrogen sulphide cut each other at a point which is the equilibrium point of pyrite and pyrrhotite in the presence of hydrogen sulphide. According to the formulas 2) and 4) this point lies at the temperature ca. 610°C while the earlier determination cited had given a value of 565°. The difference is considerable but a notable correction is arrived at in the determination of the equilibrium point in sulphur gas, too. According to the earlier observation at the pressure of 1 atm., it was at 665°, but the equation 2) gives 685°. The correction is smaller in this latter case but the direction of deviations is in both cases the same.

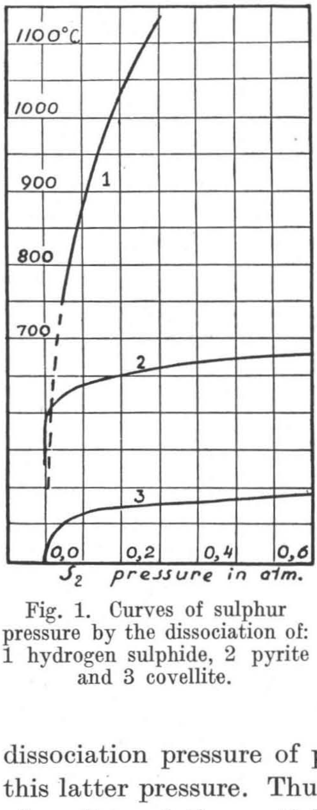


Fig. 1. Curves of sulphur pressure by the dissociation of: 1 hydrogen sulphide, 2 pyrite and 3 covellite.

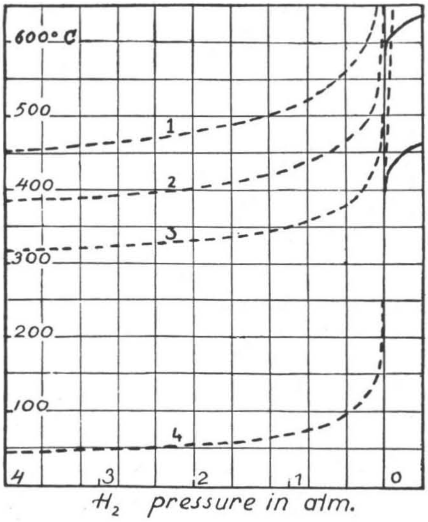


Fig. 2. Equilibrium temperatures of the association pyrite — pyrrhotite at hydrogen sulphide pressures of 10, 1 and 0,1 atm. (curves 1, 2 and 3 resp.) and covellite — chalcocite at the hydrogen sulphide pressure of 1 atm. (curve 4).

Since the dissociation constant, K, of hydrogen sulphide as well as the pressure of sulphur, p, at the equilibrium point of pyrite and pyrrhotite are known as a function of temperature, the formula 3) gives the partial pressure of hydrogen also in dissociated hydrogen sulphide at all temperatures and at all pressures of hydrogen sulphide. We may write the formula 3) in the following form:

$$[H_2] = \sqrt{K/[S_2]} \cdot [H_2S] \qquad \text{where } [S_2] = p \text{ in 2).}$$

For every temperature and pressure of hydrogen sulphide we can calculate the corresponding partial pressure of hydrogen, and thus Fig. 2 is obtained.

In addition Allen and Lombard have carried out similar investigations on the equilibrium between covellite and chalcocite, $CuS—Cu_2S$. The constants arrived at by the calculation of the interpolation formula have also in this case been large numbers but by calculations similar to the foregoing a formula more convenient for calculations was obtained:

$$5) \qquad \log p = -\frac{11\,650}{T} + 4.185 \log T + 2.8.$$

This equilibrium is similarly influenced by the presence of hydrogen sulphide, lowering the sulphur pressure and thus also the equilibrium temperature, while:

$$Cu_2S + H_2S \leftrightarrows 2\ CuS + H_2.$$

Some points on this equilibrium curve, too, were calculated and plotted in Fig. 2.

The accuracy of formulas 2) and 5) seems to be very reliable for the calculations worked out in the foregoing. More uncertain, however, is the interpolation of the dissociation of hydrogen sulphide according to formula 4). There are determinations only for points lying at 750°C and above this temperature. The distance to the equilibrium temperatures of covellite — chalcocite at 250—50°C is very great but the bending of the curve in this region is slight. On the other hand the curves cut each other at low angles and the inaccuracy as regards temperature will be considerable. As to the equilibrium point of pyrite — pyrrhotite, at the temperatures 600—400°C, the uncertainty, however, is probably not so significant.

As to the result of this investigation the curves 1, 2 and 3 in Fig. 2 give rise to the following conclusions:

1. The equilibrium of pyrite — pyrrhotite depends on three factors: the temperature and the partial pressures of hydrogen and hydrogen sulphide.

2. The formation of pyrrhotite at temperatures beneath 350°C is only possible at higher pressures of hydrogen and especially at very low pressures of hydrogen sulphide. Its content in ores formed under hydrothermal conditions, below 350°C, must accordingly be very low. The formation of ores consisting of compact pyrrhotite and of those with a higher content of pyrrhotite must have taken place at higher temperatures. The presence of pyrite in such ores presupposes a higher pressure of hydrogen sulphide.

3. The presence of pyrite is possible up to the temperatures of 650°—700°C at low pressures of hydrogen and at higher concentrations of hydrogen sulphide. Its formation along open cracks is thus possible in a wide temperature range.

4. According to curve 4, representing the equilibrium curve of covellite — chalcocite, these minerals occur together in a stable association at a partial pressure of 1 atm. hydrogen sulphide and 2.42 atm. partial pressure of hydrogen at a temperature of 50°C. In the presence of 1 atm. pure hydrogen sulphide the equilibrium point should be at 390°C. Perhaps it will be possible to control this point by experimental investigations. In that case the course of the dissociation curve of hydrogen sulphide could be corrected and the precision of the foregoing results revised.

On account of the ascertained importance of hydrogen sulphide in the formation of sulphide ores, especially of the iron pyrites, it can be considered possible that the relative amounts of pyrite and pyrrhotite in an ore are not always the primary ones, as is often supposed. Not rarely is the adjacent rock altered metasomatically into sericite schist and pyritized in addition, obviously owing to the influence of hydrogen sulphide. Certainly the fragments of sericite schist sometimes included in the sulphide ore seem to indicate that the sericitization has preceded the sulphide mineralization but it is possible, too, that it has survived this.

In the belt of sulphide mineralization of the Otravaara region in Eastern Finland pyrite is the main sulphide mineral but in addition pyrrhotite is also important as is indicated by the almost continuous zone of magnetic anomalies. It occurs locally as a compact ore with phenocrysts of pyrite and fragmental inclusions of sericite schist (Saksela, 1923). These formations, e. g. the occurrences at Räsvaaranlaki (Rasivaaranlaki) and Pihlajavaara, closely resemble eruptive breccias, but the pyrite ores are devoid of such structures. The pure pyrite ores could be comprehended as posterous replacements of the pyrrhotite ores and be formed by an addition of sulphur from hydrogen sulphide. Inasmuch as these ores locally contain microscopic grains of pyrrhotite the pressure of hydrogen sulphide should, in the last phase of this process, have dropped to such an extent that the equilibrium between pyrite and pyrrhotite was restored but then only minute particles of pyrrhotite could be formed.

Such a formation of pyrite ore may be considered secondary or metamorphic, but it can also be comprehended as a result of the last, hydrotermal phase in a continuous process of ore formation at the beginning of which compact deposits of pyrrhotite ore have been produced at higher temperatures. More conspicuous does such an influence of hydrogen sulphide seem to have been in the formation of the Hevoskumpu ore, 9 km to the east from Otravaara (Aurola and Vähätalo, 1939). It is included in highly metamorphic rocks of quartzitic character, brecciated on the hanging wall side and containing epidote and plagioclase, whereas the footwall rock is strongly sheared, schistose, pyritized, and rich in sericite. The composition of the ore body is variable insofar as the hanging wall side consists mainly of pyrrhotite and gradually turns to the chiefly pyritic ore on the footwall side. It can be thought that the pyritic composition is a secondary feature owing to the metasomatic replacement of pyrrhotite by pyrite in connection with the sericitization and shearing of the footwall rock.

The porphyritic structure of the sulphidic ores occurring, except for the aforementioned instance, e. g. at Outokumpu and Sulitjelma were earlier considered to be due to the primary crystallization of sulphidic ore magma but by some authors also as a porphyroblastic feature (Carstens, 1941). It is known that pyrite occurring in schists has the form of most beautiful idioblasts. It is thus possible that the often rounded crystals of pyrite in the sulphidic ores are not primary phenocrysts but secondary porphyroblasts formed through the influence of hydrogen sulphide.

Besides, some other parageneses between the sulphide minerals are similarly influenced by the presence of hydrogen sulphide. The relations between chalcocite and covellite have been briefly broached upon in the foregoing. Other dissociating minerals are, according to Allen and Lombard, $\mathrm{CuFeS_2}$, $\mathrm{Cu_3FeS_4}$ and $\mathrm{PtS_2}$. But in regard to the first of these, the chalcopyrite, several possibilities are to be considered:

$$\begin{array}{ll}
\mathrm{CuFeS_2 + H_2S = CuS + FeS_2 + H_2} & \text{(covellite)} \\
\mathrm{2\ CuFeS_2 + H_2S = Cu_2S + 2\ FeS_2 + H_2} & \text{(chalcocite)} \\
\left.\begin{array}{l}\mathrm{3\ CuFeS_2 + H_2S = Cu_3FeS_3 + 2\ FeS_2 + H_2} \\ \mathrm{9\ CuFeS_2 + 4H_2S = Cu_9FeS_6 + 8\ FeS_2 + 4\ H_2}\end{array}\right\} & \text{(bornite)} \\
\mathrm{3\ CuFeS_2 + FeAsS + 5\ H_2S = Cu_3AsS_4 + 4\ FeS_2 + 5\ H_2} & \text{(enargite)}
\end{array}$$

REFERENCES

Allen, E. T., Grenshaw, J. L. and Johnston, J. (1912) The mineral sulphides of iron. Am. J. Sci. 33, p. 203.

—»— and Lombard, Robert H. (1917) A method for the determination of dissociation pressures of sulphides and its application to covellite (CuS) and pyrite ($\mathrm{FeS_2}$). Ann. Rep. Geophys. Lab. 1914, p. 156 and Am. J. Sci. (4) 43, p. 175.

Aurola, Erkki and Vähätalo, Veikko (1939) The pyrite deposits of Hevoskumpu in Tuupovaara. Bull. Comm. géol. Finlande 125, p. 65.

Breuner, G. und Schupp, W. (1909) Dissoziation des Schwefelwasserstoffes. Zeitschr. f. phys. Chemie 68, p. 161.

Carstens, C. W. (1941), Zur Frage der Metamorphose der Schwefelkieserze. Kong. Norske Vid. Selsk. Forh. XIV, p. 9.

Saxén, Martti (1923) Über die Petrologie des Otravaaragebietes im östlichen Finnland. Bull. Comm. géol. Finlande 65.

Väyrynen, Heikki (1935) Über die Mineralparagenesis der Kieserze in den Gebieten von Outokumpu und Polvijärvi. Bull. Comm. géol. Finlande 109.

—»— (1951) Rikkikiisu — magneettikiisuyhtymän tasapainosuhteista. Geologi N:o 2—3, p. 7.

5

ORIGIN OF ULTRABASICS IN THE ORIJÄRVI REGION [1]

BY

TOIVO MIKKOLA

Suomen Malmi Osakeyhtiö, Otaniemi, Finland

ABSTRACT

Since the re-investigation of the Orijärvi region the ultrabasic rocks, previously thought to be magmatic peridotites, are interpreted as metamorphic. Ultrabasics originated as a product of the metamorphic differentiation of a marly clay. The differentiation was brought about by strong penetrative movements during which the excess elements were removed. The present mineral composition is the result of a complete re-crystallization.

CONTENTS

INTRODUCTION

The ultrabasics occurring in the Orijärvi region in the Svecofennidic range were described petrographically by Eskola (1914, p. 89). According to him they are peridotites of magmatic origin. Parras (1941, p. 491) has written about the mineral composition and the mode of occurrence of these rocks. He also regards their origin as magmatic.

Usually the ultrabasic rocks in every part of the world, especially those containing olivine, have been regarded as magmatic. Some serpentinites,

---

[1] Received December 31, 1954.

however, have been supposed to be alteration products of dolomite. Barth (1947) seems to be the first to apply the idea of metamorphic differentiation to the ultrabasic rocks. However, I cannot agree with him that »the mechanical force was less important than the force which Scheerer called chemical attraction» (Op. cit. p. 37). Recently Sørensen (1953) published a paper in which the problem is discussed in a way very similar to that of the present study.

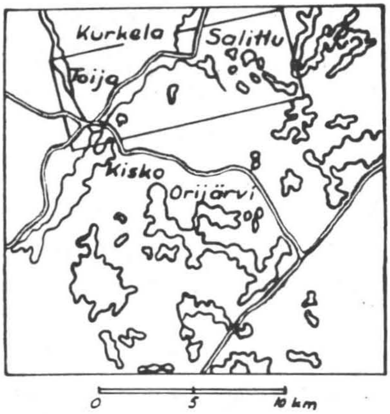


Fig. 1. Map of Kisko. The rectangle shows the area in Fig. 2.

It is evident that there are still uncertainties, especially concerning the mechanism of selective diffusion of ions. But, as shown in the present paper, kinematically controlled metamorphic differentiation seems to be the phenomenon best accounting for the formation of the ultrabasics in question.

The study is based on the field work carried out in the years 1946—50. The first manuscript was presented to the Editor on December 31, 1951.

COMPOSITION OF THE ULTRABASICS

The ultrabasics of Orijärvi resemble in their chemical and mineral composition the magmatic hornblende peridotites. The most abundant types were re-investigated under the microscope and two additional chemical analyses were made (Tabl. 1, Anal. 1 and 3). Some other analyses are cited from Eskola (1914) for comparison. The microscopic study has shown that there are no very great differences in composition or mineralogy in the ultrabasics occurring between Toija and Salittu (Figs. 1 and 2). Because Eskola has already described the petrography of these rocks, it has not been discussed in greater detail in this paper.

The texture of the groundmass of the ultrabasics in question is granoblastic and usually very fine-grained. Occurring as larger crystals there are olivine, augite, diopside, biotite, anthophyllite, and cummingtonite. Hypersthene occurs in the breccia of Salittu. Olivine seems to exist in every variant in association with one or two of the other minerals mentioned. The groundmass consists of greenish edenitic amphibole. Serpentine, chlorite, talc, carbonate, and magnetite evidently occur as alteration products. Accessories are spinel, pyrite, pyrrhotite, chalcopyrite, sphalerite, and apatite.

According to the description of Eskola (1914, p. 89), the peridotite from Salittu (Table 1, Anal. 2) seems to be mineralogically and chemically very similar to the rock from Toija (Anal. 1). The analyzed rock from Toija belongs to a bed more than 2 km in length. The thickness is variable but it never exceeds 50 m. In a fine-grained hornblende groundmass of this rock there occur larger crystals of forsterite with a diameter ranging from 0.5 to 3 mm. They are partly altered to serpentine and magne-

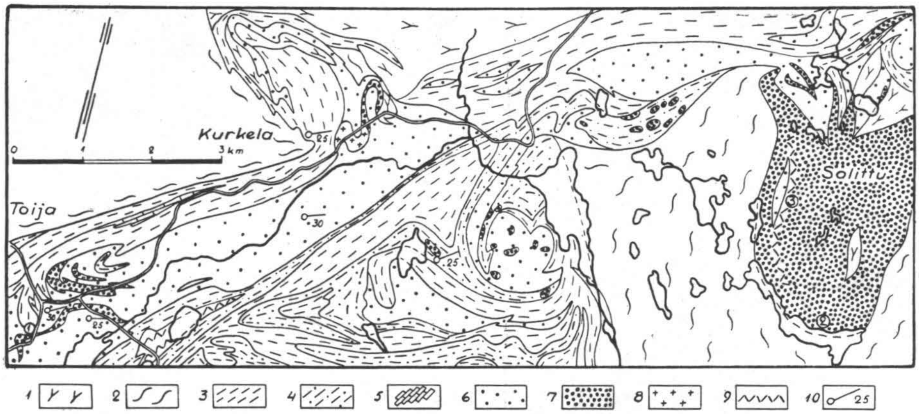


Fig. 2. Map of Toija-Kurkela-Salittu. Based on the geological detail map made in the scale 1 : 4000. Legend: 1. Granodiorites and associated rocks. 2. Microcline granites. 3. Mica schists. 4. Amphibolites. 5. Marble and skarn. 6. Diopside-amphibolites. 7. Ultrabasics. 8. Cordierite- anthophyllite rocks. 9. Breccias and mylonites. 10. Fold axis. The numbers in circles show the places from which the analyses were made.

tite. The amount of forsterite including the alteration products is 32 %. Another mineral forming larger crystals is augite ($\gamma \Lambda C = 42^\circ$, dispersion strong, $\varrho > v$), diameter between 0.1 and 0.5 mm. Its amount is 15 %. In places augite seems to replace forsterite. It is practically free from inclusions. Besides serpentine, there occurs another common alteration product viz. colorless chlorite (almost uniaxial, opt. posit.) as pseudomorph after biotite. Some remnants of biotite are still visible. Biotite is abundant in sheared variants, for example, at Kurkela.

Table 1

	1	2	3	4	5	6	7
SiO_2	42.49	42.96	45.25	49.61	51.5	42.25	48.00
TiO_2	0.98	1.27	1.45	0.56	0.9	0.75	2.06
Al_2O_3	8.13	8.11	16.23	15.21	16.3	13.40	18.62
Fe_2O_3	5.52	3.31	2.83	0.89	5.0	4.10	1.07
FeO	8.02	10.36	9.58	8.77	2.9	2.40	16.18
MnO	0.22	0.23	0.14	0.05	—	—	0.13
NiO	0.11	0.05	0.09	—	—	—	—
MgO	20.70	19.30	12.19	5.02	5.6	4.60	11.85
CaO	9.40	9.65	8.47	16.32	14.8	12.10	0.64
Na_2O	0.60	0.24	1.22	1.20	1.2	2.25	0.23
K_2O	0.02	0.11	0.17	1.36	1.2	2.25	0.01
P_2O_5	0.25	—	0.17	0.19	—	—	0.09
Cr_2O_3	0.18	0.24	—	—	—	—	—
H_2O+	3.20	3.93	2.19	0.55	0.6	5.35	1.50
CO_2	0.00	—	—	—	—	10.55	—
	99.82	99.82	99.98	99.73	100.0	100.00	100.57

1. Ultrabasic rock, Toija, Kisko, Finland. Analyst A. Heikkinen.
2. Peridotite, Salittu, Suomusjärvi, Finland. Analyst P. Eskola (1914).
3. Hypersthene-bearing fragment of breccia in ultrabasic rock. Salittu, Suomusjärvi, Finland. Analyst H. B. Wiik.
4. Diopside-amphibolite, Vetjo, Kisko, Finland. Analyst P. Eskola (1914).
5. »Diopside-amphibolite», calculated from No. 6; CO_2, 90 % of water and more than half of alkalies removed.
6. Calculated marl, one part of average limestone and three parts of glacial clay. After Pettijohn (1949).
7. Cordierite-anthophyllite rock, Träskböle, Perniö, Finland. Analyst P. Eskola (1914).

Fig. 3. Boudins of ultrabasics (dark) in diopside-gneiss (light). The boudins have distinct contacts and internal layering of diopside-gneiss. Toija, Ilves.

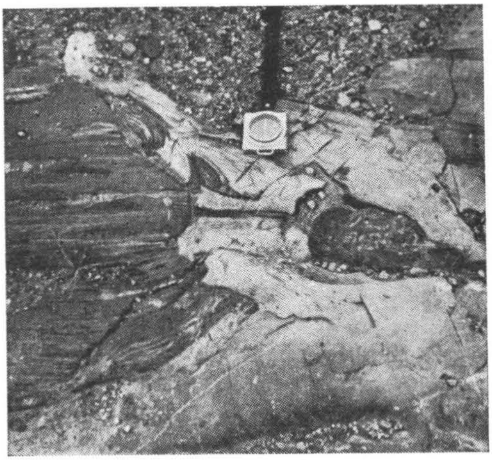


Fig. 4. Detail from Fig. 3. Small boudin completely massive, larger boudin has a weak interbedding of diopside-gneiss.

MODE OF OCCURRENCE

In this connection it should be mentioned that according to my stratigraphical investigations the rocks of Orijärvi belong to a sedimentary zone (Mikkola, 1950 and Simonen, 1953). The following series, from top to base, has been observed:

Geosynclinal facies
  1. basic volcanites? (amphibolites)
  2. greywacke with mica-schist (leptite)
  3. polymictic conglomerate

Atypical foreland facies
  1. banded iron ore
  2. black schist

Foreland facies
  1. limestone (marble)
  2. marl (diopside-gneiss and -amphibolite)
  3. mica-schist (leptite)

This is based on the sequence of sedimentation given by Pettijohn (1949, pp. 454—461). The sedimentary origin of these rocks is evident (excluding the topmost amphibolites).

Usually the ultrabasics are associated with diopside-amphibolite or -gneiss, rarely with skarn, marble or other rock rich in lime (Fig. 2). They always exist in the zone of foreland facies.

Commonly the ultrabasics seem to occur as concordant layers in the rocks referred to. In some places, e. g. at Kurkela, the ultrabasics occur as great lenses in these rocks. At Toija small lenses of ultrabasics form boudins in diopside-gneiss (Figs. 3 and 4). It may be that the occurrence of ultrabasics at Kurkela is the same as at Toija but of greater dimensions.

In appearance the ultrabasics may be massive or layered. All transitional forms occur. Layers of skarn and marble are rare but do exist as at Toija. Layers of other surrounding rocks are common (Figs. 3 and 4). There are pseudoagglomeratic breccias, tectonic by nature, such as those at Toija and Salittu (Fig. 5). At Salittu the brecciated zone is 3—4 km in length and some 2—300 m in breadth. The fragments are usually lighter in color and less basic than the matrix (Table 1, Anal. 3) but they may also contain abundant olivine.

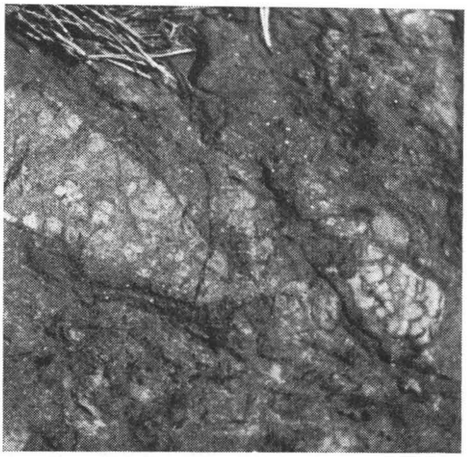


Fig. 5. Breccia in ultrabasic rock. Fragments are lighter in color than the matrix. White patches are hypersthene crystals. The small fragment is crowded with hypersthene. Larger fragment has a sharp contact upwards. Lower contact is gradual. Salittu, Suomusjärvi.

Usually the contacts of ultrabasics are distinct (Figs. 3 and 4) but sometimes gradual. An excellent outcrop for the study of the contact phenomena between ultrabasics and diopside-amphibolites occurs at Kurkela, 150 m W. from the highway and 30 m S. from the road to the lodge of the Kisko Labor Association. No distinct contact exists but in some layers of diopside-amphibolite olivine crystals occur (Fig. 6). The occurrence of olivine is not continuous in the direction of strata. The olivine-bearing parts of the rock form irregular lenses or there are only some scattered olivine crystals visible. Statistically the amount of olivine increases towards the contact of ultrabasic rock, and there is a gradation zone about 1 m broad where it is impossible to say whether the rock is ultrabasic or diopside-amphibolite. The ultrabasic rock some metres away is completely massive, showing no layered structure. It is sheared in places, however.

In diopside-amphibolites there are also transitional forms which range mineralogically between ultrabasics and diopside-amphibolites but do not show gradations into the true ultrabasic bodies. In appearance they are like diopside-amphibolites but contain olivine and serpentine. One occurrence is beside the highway between Kisko and Toija.

Many more bodies of ultrabasics were located in detail mapping than appear on the older maps. On the other hand, the peridotite massif of Kurkela is only a group of ultrabasic lenses in diopside-amphibolite. The massif of Salittu, though almost homogeneous, seems to be a core-filling in a large fold. However, it also contains thick layers of diopside-amphibolite.

The color of the ultrabasics is dark greenish. Very often they have rusty patches on the surface representing weathered olivine.

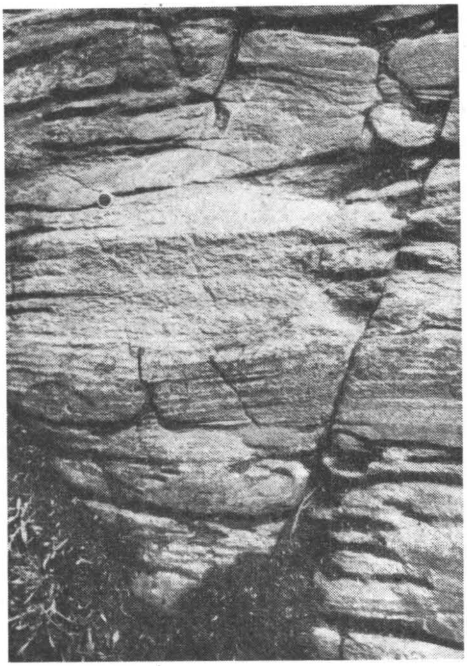


Fig. 6. Diopside-amphibolite close to the contact of ultrabasics. The layered rock contains irregular lenses of ultrabasics with strongly weathered surface. Some layers contain only scattered olivine crystals (dark weathered patches). A weak boudinage structure is visible. Kurkela, Kisko.

ORIGIN OF THE ULTRABASICS

In composition the diopside-amphibolite shows gradations to the ultrabasics. Since the ultrabasics are usually associated with this rock and because it is important in the discussion of the origin of ultrabasics, I shall describe it briefly.

The rock called diopside-amphibolite is very heterogeneous. In appearance it is greenish grey, sometimes massive but most often layered. (Definitions of layer and stratum according to Schrock, 1948). The thickness of layers vary from a few mm up to 5 cm. The layers may be strata but a part of them could have been differentiared mechanically (Schmidt, 1932). The light leptitic layers in diopside-amphibolite are undoubted strata because graded bedding is often evident. This feature has not been preserved in every variant. Usually there occur alternately only dark and light layers. The dark layers resemble the common amphibolites or amphibole schists of the Orijärvi region; in other words, they consist of fine-grained hornblende, basic plagioclase, and accessories. Very often diopside crystals are visible. The light layers are regionally of at least two types, viz., such consisting of plagioclase, microcline, and diopside, and others of scapolite, plagioclase, diopside, and carbonate. The texture is crystalloblastic. The carbonate content is characteristic of all variants of diopside-amphibolite.

Especially from the stratigraphical point of view, the sedimentary origin of this rock is apparent. The source rock has been a marl (Table 1, Anal. 4—6), which also occurs as interbeds of a thin layered marble and mica-schist at Kuovila (Eskola, 1914, Map 1).

Probably the fragment analyzed (Anal. 3) represents an intermediate stage between ultrabasics and diopside-amphibolites. In addition to fragments there are amphibolitic and diopside-amphibolitic layers in the breccia of Salittu, only partially brecciated. In composition they likewise are intermediate between the basic and ultrabasic rocks similar to the fragment in question. Because the layers are brecciated mechanically, the entire zone seems to be tectonic by nature. The minerals, however, are quite undeformed and thus formed after the cessation of movements; in other words, the present mineral paragenesis cannot be original.

In the groundmass of these fragments (Anal. 3), there now occur clear and unaltered prisms of anthophyllite, 0.5 to 5 mm in length. A small amount of nearly colorless biotite is visible. Magnetite occurs in abundance. Some serpentinized olivine crystals also have been found. The rock is characterized, however, by hypersthene crystals. In another sample no anthophyllite is visible; instead it contains more olivine. Hypersthene occurs as reddish grey nodules, 0.5 to 3 cm in diameter (Fig. 5). Under the microscope, the crystals are colorless and not as well shaped as they seem to be with the unaided eye. They are crowded with inclusions of hornblende. The optical properties of this hypersthene are: $2V\alpha = 78°$, $\alpha = 1.685$, $\gamma = 1.694$. For chemical analysis the hypersthene (Table 2) was separated by means of centrifuge and the powder was completely pure under the microscope.

Table 2. Chemical Composition of the Hypersthene from Salittu. (Analyst H. B. Wiik)

SiO_2	51.33
TiO_2	0.27
Al_2O_3	4.50
Fe_2O_3	0.52
FeO	16.38
MgO	27.50
CaO	0.10
H_2O+	0.00
H_2O-	0.00
Total	100.60

The amount of hypersthene is variable but in the variety analyzed it is about 10 % (determined by the combined integration method, the crystals on the rock surface by the point method, the inclusions under the microscope). Because this mineral contains 4.5 % Al_2O_3, the Al_2O_3-content in the groundmass must consequently be about 17—18 %. Since the groundmass consists mainly of fine-grained amphibole, this mineral must be much richer in Al_2O_3 than the variety analyzed by Eskola (1914, p. 91), which contains Al_2O_3 about 9 %. Mineralogically it seems to be an edenitic hornblende rich in CaO and Al_2O_3. No analysis has been made as yet.

The ultrabasic parts of diopside-amphibolites cannot be thought to have a magmatic origin because they are only variants of a metamorphic sediment without any possible connection with plutonic rocks. Thin separate layers of ultrabasics and small boudins with internal layering, occurring only in diopside-amphibolite or -gneiss, can hardly be magmatic, either. Since the occurrences of other bodies of ultrabasics in the region are similar, it is likely that they, too, are of non-magmatic origin. As contributive evidence, no gabbros or diorites are visible.

The possibility of volcanic origin is discussed by Eskola (1914, p. 94) and found hardly tenable. There is no evidence of volcanic origin in these rocks. The existence of some gradual contacts and transitional forms make a volcanic origin impossible.

The existence of ultrabasics in the limestone horizon, especially in the diopside-amphibolites, together with the phenomena discussed in the foregoing, suggest that they should be regarded as metamorphic derivatives of rocks rich in lime. However, they are not sedimentary rocks of unchanged bulk composition. Very probably no sediments of peridotitic composition do exist. In any event, these rocks have undergone a metamorphism that has changed their composition as shown in the next section.

MECHANISM OF METAMORPHISM

All the rocks of the Svecofennian range are strongly metamorphic. But in what way has the metamorphism taken place in this particular case? If we take as our starting point some kind of marl, from which the diopside-amphibolites are supposed to have originated, it seems to me that there are two principal possibilities. The first is an addition of Mg and Fe and simultaneous removal of the constituents in excess, that is, Mg metasomatism; the second is simply the removal of excess elements, the alkalies, Ca, and Si, that is, some kind of metamorphic differentiation.

Metasomatic processes are generally supposed to be characterized by gradual contacts, but unfortunately not always. On the other hand, differentiation is not limited to a certain type of contact. Therefore it is not possible to draw any final conclusions from contact relations alone. Because the contacts most often are distinct, metasomatism does not seem probable. Of course there is the possibility that gradual contacts have existed everywhere and that gliding movements have destroyed them. However, the ultrabasics had attained their present bulk composition already before the cessation of folding or achieved a greater competency than the clayey part of the surrounding schists. Had they not, there could be no ultrabasic boudins. Because the boudins are not sheared, the gliding could not have been very intense after the formation of the boudins and ultrabasics and the alteration in the contact relations has been insignificant. In other words, gradual contacts have not existed everywhere. Further the fragments of breccias give evidence in favor of differentiation. If the process had been metasomatic, how could the fragments have their different compositions, compared to those of the matrix? However, there is still another and better way to decide between metasomatism and differentiation.

In general the mode of occurrence of these ultrabasic olivine-bearing rocks is very similar to that of Mg Fe rich cordierite-anthophyllite rocks of the Orijärvi region. At Kurkela the cordierite-anthophyllite rocks and olivine rocks have a parallel distribution. They are separated only by a narrow area lacking outcrops. It is not probable that the addition of Mg in these two rocks has taken place in different ways. Thus the problem of olivine rocks seems to be similar to that of cordierite-anthophyllite rocks.

Eskola (1914, p. 252) has based his hypothesis of Mg metasomatism on the cordierite-anthophyllite rocks. For the reasons given by Tuominen and Mikkola (1950) and Tuominen (1951) it is unlikely that Mg metasomatism took place. If the cordierite-anthophyllite rocks are not metasomatic, the olivine rocks with the same mode of occurrence could hardly be metasomatic, either. In addition, by supposing differentiation during folding instead of metasomatism under quiet conditions, there is no »room problem», because during the folding the rock could have the volume it needs.

These cordierite-anthophyllite rocks have been described as products of kinematical metamorphic differentiation (Tuominen and Mikkola, 1950). Some principles presented in that paper extend much further than the cordierite-anthophyllite rocks. We could show that a rock can, by the agency of mechanical processes, split into two or several components; in other words, new rocks with new bulk compositions are formed during folding. Taking for granted that there exists no sedimentary rock or even a stratum with a primary bulk composition of cordierite-anthophyllite rock, it is probable that besides purely mechanical differentiation the folding and shear movements call forth a migration of easily moving elements and a continuous re-crystallization, which all act simultaneously. Since this has not been experimentally investigated, it is only a hypothesis which is in good agreement with the field observations.

When comparing the formation of olivine-bearing rocks with that of cordierite-anthophyllite rocks, it is evident that the process is not quite the same. There are in evidence no very clear flow structures characteristic of the environment of cordierite-anthophyllite rocks. However, traces of such a deformation are met with (Fig. 3). On the other hand, the flowage is not necessary, for the strong penetrative movements alone, very likely, can change the composition of a marl into that of an ultrabasic rock by removing excess elements. Obviously the water contained in these sediments has acted as a carrier. Further, there is the possibility that some portion of the incompetent layers has glided away, thus facilitating the alteration of the bulk composition; but there is no definite evidence of this gliding. Perhaps the strong shear movements, which clearly have existed in the region and seem to be necessary for the formation of peridotitic composition, have destroyed the traces of flowage probably occurring during the early stage of deformation.

In comparing Analysis 7 with 3 (Table 1), a great similarity is apparent. Only the content of CaO is decidedly lower in cordierite-anthophyllite rock. But, since the amount of CaO in the residuum always decreases during the process of this kind of metamorphic differentiation, the highest possible content of CaO depends, in principle, only upon the initial material. Therefore the content of CaO in cordierite-anthophyllite rocks, formed from clays, cannot be as high as in the rocks which originate from marls.

It is possible that the brecciation at Toija and Salittu occurred in an early stage of deformation. The gliding movements have not had a great enough influence upon the fragments. Consequently they could not attain the peridotitic composition completely.

In general peridotites are rich in Cr_2O_3 and NiO. Their content in the ultrabasics of Orijärvi is much lower, about 0.25 and 0.10 % compared with the average 0.5 and 0.4. An exogenic enrichment of Cr and Ni occurs in clays (Rankama and Sahama, 1950). The NiO content in the clayey rocks of Orijärvi is, according to spectrographic determinations, about 0.03 %. The quantity of Cr_2O_3 in these same rocks has not been determined but, according to Sahama (1945), it is in clayey rocks, in general, about 0.05 %. Depending upon the carbonate content of marl, the amount of Cr and Ni in the source material of ultrabasics could be considerably lower. Ni is very similar to Fe and Mg geochemically. The FeO content in recent clays is approximately 3 %. Excluding the carbonate part, the enrichment of FeO and NiO in these ultrabasics is nearly the same, about 3-fold. The stage of Cr enrichment is slightly greater, perhaps 5-fold. It is not, however, greater than the enrichment of Mg. Though Cr and Mg differ geochemically, Cr substitutes for Mg (Rankama and Sahama, 1950). Consequently the great enrichment of Cr is not only possible but is to be expected. The encichment of Cr and Ni cannot be offered as evidence here but it does not contradict the hypothesis in question.

ORIGIN OF OLIVINE AND PYROXENES

What the mineral composition of the ultrabasics of Orijärvi might have been at each stage of deformation, is impossible to say; but at present it represents the result of thorough re-crystallization and neo-mineralization.

The last re-crystallization took place after the folding, as there are only slight traces of deformation of minerals, even in places where strong movements have evidently occurred, e. g. in breccias and boudins. The major minerals evidently are formed at great depths in a geosyncline and correspond to a bulk composition obtained in the way described. Later the diaphthoretic processes have changed a part of the olivine into serpentine.

The formation of olivine and pyroxenes metamorphically is possible in the PT conditions of a geosyncline. According to the laboratory experiments of Bowen and Tuttle (1949) on the system $MgO—SiO_2—H_2O$, forsterite is stable at all pressures down to a temperature of about 430°C in the presence of an excess of water, and olivine containing FeO is stable to yet lower temperatures. Below 430°C forsterire is stable in the region deficient in water. Excluding the cummingtonite, all the other Mg minerals occurring in the ultrabasics of Orijärvi have been formed in the experiments of Bowen and Tuttle. The mineral parageneses in equilibrium pointed out by them also seem to occur in these rocks, e. g. talc is found only in assemblage with anthophyllite.

CONCLUDING REMARKS

With regard to the purpose of the present study, the structure of Orijärvi is already sufficiently dealt with by Tuominen and Mikkola (1950). As seen in the map (Fig. 2), the fold axes plunge gently to the E. or E. N. E. The central part of the Orijärvi area is characterized by isoclinal folding, and the other parts by open folds.

In the center of the Orijärvi region there occurs a great amount of diopside-amphibolite but no olivine-bearing rocks. The absence of ultrabasics probably arises from the nature of tectonic movements. In the center, flow folding is common. However, the horizon is the same as in the regions where ultrabasics exist but stronger penetrative movements have dominated there. Between Toija and Kurkela the ultrabasics seem to be connected with a gliding zone. Almost all the diopside-amphibolites in this region contain some olivine or layers of ultrabasics.

The ultrabasics of Orijärvi do not seem to be host rocks of nickel ores. One occurrence at Toija with diopside and apatite porphyroblasts contains a considerable amount of pyrite and chalcopyrite but no more nickel (0.05 %) than is usually found in the region (0.05—0.1 %). Therefore, the formation of copper ores in connection with these ultrabasics seems possible but that of nickel ores remains uncertain. A small amount of zinc ore has recently been detected in the ultrabasics of Toija.

Though the olivine-bearing ultrabasics of Orijärvi apparently have been formed from marls, other rocks rich in Ca could have been the source material. But, as the cordierite-anthophyllite rocks are most easily formed from argillites, the metamorphic formation of ultrabasics in question seems to be associated with marls.

Basic and ultrabasic plutonic rocks have been classified by Turner and Verhoogen (1951) into three groups:

1. Gabbros, peridotites, and associated rocks of lopoliths and thick sheets.
2. The peridotite serpentinite association of orogenic zones (Alpine type).
3. The pre-Cambrian anorthosite association.

According to their composition and mode of occurrence, excluding the origin, the ultrabasics of Orijärvi are rocks of the second group. But, because they are metamorphic sedimentary rocks, there is a possibility that some part of the second group in other regions, too, is of a non-plutonic origin.

In this paper I have used the name ultrabasics, because the name peridotite belongs to the rocks of magmatic origin, as yet.

Acknowledgements — This study was carried out for the prospecting company Suomen Malmi Osakeyhtiö as a part of its program in the Orijärvi region; and I am indebted to the chief geologist of the Company, Mr. H. Tuominen, for helpful discussions and suggestions during the investigation.

With Messrs. V. Hyppönen, I Salli, and Dr. M. Härme, I have had an opportunity to investigate several ultrabasic areas mapped by them. Mrs. Toini Mikkola and Mr. V. Makkonen have made some optical determinations. The analyses were made by the Geol. Survey of Finland. Dr. S. Gavelin and Dr. H. S. Yoder, Jr. have kindly read the manuscript and contributed some valuable remarks. The constructive criticisms and suggestions made by Prof. P. Eskola over a period of four years are appreciated.

REFERENCES

BARTH, T. F. W. (1947) The nickeliferous Iveland-Evje amphibolite and its relation. Norges Geologiske Undersökelse 168 a.

BOWEN, N. L. and TUTTLE, O. F. (1949) The system MgO—SiO_2—H_2O. Bull. Geol. Soc. Amer. 60, p. 439.

ESKOLA, P. (1914) On the petrology of the Orijärvi region in south-western Finland. Bull. Comm. géol. Finlande 40; Fennia 37, No. 3.

MIKKOLA, T. (1950) Orijärven alueen rakennetta ja stratigrafiaa. Geologi, p. 43.

PARRAS, K. (1941) Das Gebiet der Pyroxenführenden Gesteine im westlichen Uusimaa in Südfinnland. Geol. Rundschau 32, p. 484.

PETTIJOHN, F. J. (1949) Sedimentary rocks. New York.

RANKAMA, K. and SAHAMA, TH. G. (1950) Geochemistry. Chicago.

SAHAMA, TH. G. (1949) Spurenelemente der Gesteine im südlichen Finnisch-Lappland. Bull. Comm. géol. Finlande 135.

SCHMIDT, W. (1932) Tektonik und Verformungslehre. Borntraeger, Berlin.

SHROCK, R. (1948) Sequence in layered rocks. New York.

SIMONEN, A. (1953) Stratigraphy and sedimentation of the svecofennidic, early archean supracrustal rocks in south-western Finland. Bull. Comm. géol. Finlande 160.

SØRENSEN, H. (1953) The ultrabasic rocks at Tovqussaq, west Greenland. Meddelelser om Grønland 136, No. 4.

TUOMINEN, H. and MIKKOLA, T. (1950) Metamorphic Mg-Fe enrichment in the Orijärvi region as related to folding. C. R. Soc. géol. Finlande 23; Bull. Comm. géol. Finlande 150, p. 67.

TUOMINEN, H. (1951) Metamorphic concentration of magnesium and iron in the Orijärvi region: A reply. C. R. Soc. géol. Finlande 24; Bull. Comm. geol. Finlande 154, p. 233.

TURNER, F. and VERHOOGEN, J. (1951) Igneous and metamorphic petrology. New York.

6

SULPHATIC CANCRINITE FROM IIVAARA, KUUSAMO, FINLAND [1]

BY

MAUNO LEHIJÄRVI

Geological Survey of Finland, Helsinki

ABSTRACT

Chemical analyses, physical properties, powder diffraction data and unit cell dimensions for sulphatic cancrinite from Iivaara are presented.

In 1937 Mr. V. Pääkkönen, M. A., of the Geological Survey of Finland, collected a specimen of ijolite from the northern slope of Iivaara, a hill in Kuusamo, Finland. In a microscopic examination made by the author, this specimen (No. 5737 Geologia Fennoscandiae, Institute of Geology, University of Helsinki) was found to contain a mineral that has a lower refringence than nepheline and a low birefringence. The author separated and analyzed this mineral and calculated its unit cell dimensions. The mineral proved to be a sulphatic cancrinite that had not been discovered at Iivaara before.

A planimetric analysis of the rock, determined on the Leitz Integration Stage, gave the following results:

Nepheline	36.4	Cancrinite	2.6
Pyroxene	37.0	Apatite	2.3
Sulphatic cancrinite	20.8	Sphene ..	0.9

Under the microscope, the sulphatic cancrinite shows a dirty grey colour caused by pigment (Fig. 1). Some of the crystals are hexagonal in form. The inclusions consist of carbonatic cancrinite, zeolite and opaque ore.

The separation was performed by centrifuging the material in Clerici solution. The result of the analysis, including the atomic proportions, optical properties, and specific gravity appear in Table 1.

---

[1] Received February 4, 1955.

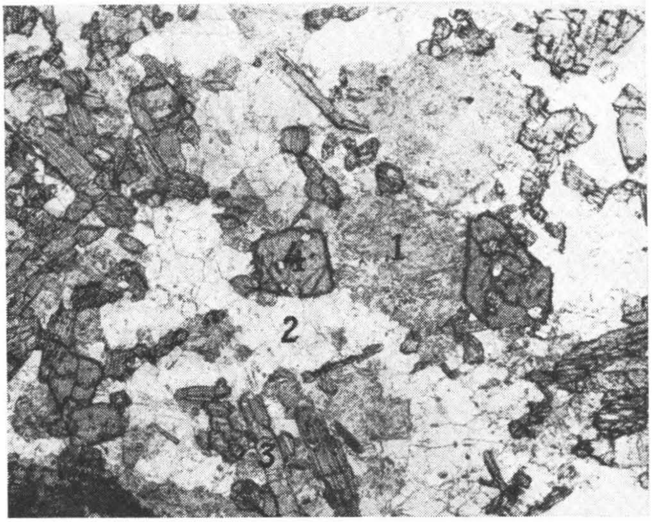


Fig. 1. Ijolite from Iivaara. 1, Sulphatic cancrinite; 2, Nepheline; 3, Pyroxene; 4, Sphene. Nic. ||. Magn. 20×. Photo A. Matisto.

Table 1. Chemical composition and physical properties of sulphatic cancrinite from Iivaara

Constituent	%	Mol.prop.	Atomic prop. (0=24)		
SiO_2	35.40	5891	Si	6.12	12.00
TiO_2	0.00	—	Al	6.07	
Al_2O_3	29.78	2921	Ca	0.21	7.57
Fe_2O_3	0.05	3	Na	6.96	
FeO	0.42	58	K	0.21	
MnO	0.00	—	CO_3	0.51	1.22
MgO	0.13	32	SO_4	0.61	
CaO	1.12	200	Cl	0.10	
Na_2O	20.77	3350	O	24.00	
K_2O	0.96	102	H_2O	2.40	2.40
Cl	0.35	99			
SO_3	4.70	587			
CO_2	2.16	491			
H_2O+	4.17	2315			
H_2O-	0.00	—			
	100.01	ω = 1.499			
Less O for Cl	0.08	ε = 1.493			
Total	99.93	Sp.gr. = 2.402 (pycnometer)			

The atomic proportions of cancrinite analyses have been usually calculated on a basis of Si + Al = 12. In this paper the basis 0 = 24 has been used in agreement with Phoenix and Nuffield (1949). The atomic proportions calculated in Table 1 fit very well into the general chemical formula of cancrinite as it is used by Phoenix and Nuffield:

$$(\mathrm{Na,K,Ca,Al})^{10+}_{6-8}\,\mathrm{Si}_6\mathrm{Al}_6\mathrm{O}^{6-}_{24}\cdot(\mathrm{SO}_4,\mathrm{CO}_3)^{4-}_{1-2}\cdot 1\text{—}5\ \mathrm{H_2O}$$

As mentioned by Stewart (1941), the optical properties of cancrinite are affected by variation in chemical composition. The birefringence, in particular, decreases rather regularly with increasing SO_3. Phoenix and Nuffield (op.cit.) arranged the analyses available to them with decreasing birefringence and found an increase of SO_3 and, on the other hand, a decrease of Ca and Ca/(Na+K). In Table 2, a number of chemical and physical data on cancrinite collected from literature have been arranged in order of increasing SO_3 expressed as molecular per cent of (SO_3 + CO_2).

Table 2. Cancrinite: Atomic proportions (0=24), optical properties, specific gravities and SO_3 as molecular per cent of (SO_3+CO_2)

	1	2	3	4	5	6	7	8	9	10	11	12	13
Si	5.94	5.97	6.28	6.17	6.04	6.11	5.99	6.12	5.94	6.25	6.21	6.20	6.03
Al ...	6.21	5.96	5.94	6.23	5.95	5.73	6.16	6.07	6.21	6.01	6.06	5.99	5.98
Na ...	5.27	5.97	3.38	6.59	5.54	6.20	6.38	6.96	6.26	5.37	5.40	4.87	5.59
K	0.11	0.04	0.28	0.15	0.34	0.10	0.33	0.21	0.28	0.94	1.12	1.36	1.14
Ca ...	1.63	1.48	2.22	0.93	1.16	1.13	0.80	0.21	0.71	0.28	0.24	0.34	0.28
CO_3 ..	1.64	1.36	1.50	1.74	0.94	0.91	0.77	0.51	0.44	0.24	0.22	0.15	0.06
SO_4 ..	0.01	0.01	0.03	0.19	0.15	0.20	0.62	0.61	0.76	0.77	0.84	0.81	0.74
Cl	—	0.01	0.02	0.13	0.01	0.01	—	0.10	—	—	—	—	0.03
O	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00
H_2O ..	2.71	1.54	3.78	2.63	2.78	2.36	2.51	2.40	1.31	4.50	4.17	3.26	3.12
ω	1.524	1.5238	1.528	1.515	1.514	1.515	1.509	1.499	1.502	1.492 — 1.493		—	1.489
ε	1.501	1.5015	1.503	1.496	1.499	1.499	1.500	1.493	1.497			—	
ω—ε .	0.023	0.0223	0.025	0.019	0.015	0.016	0.009	0.006	0.005	0.0007	0.0007	0.0007	0.0007
G	2.51	2.44	2.476	2.422	2.430	2.447	2.443	2.402	2.423	2.35	2.35	—	—
SO_3 mol. % of (SO_3+ CO_2)	0.26	0.74	1.3	9.7	11.5	16.5	44.56	56.4	63.18	75.8	79.4	84.2	91.5

1. Iron Hill, Gunnison Co., Colorado. Anal. Larsen and Foshag (1926). Incl. Mg 0.10, Ti 0.01, Mn, Cl trace.
2. Dôdô, Korea. Anal. Kôzu (1931). Incl. Mg 0.01, Fe''', P trace.
3. Bancroft, Ont. Anal. Meen (1938).
4. Blue Mountain, Methuen, Ontario, Anal. Phoenix (1949).
5. Vishnevyh gor, Ural. Anal. Arest-Jakubowitsch 1946 (Bonshtedt-Kupletskaya and Arest-Jakubowitsch, 1947). Incl. Mn 0.01.
6. Vishnevyh gor, Ural. Anal. Arest-Jakubowitsch 1946 (Bonshtedt-Kupletskaya and Arest-Jakubowitsch, 1947). Incl. Mn 0.01.
7. Beaver Greek, Colorado. Anal. Steiger Incl. Sr 0.01, Ti 0.01.
8. Iivaara, Kuusamo, Finland. Anal. Lehijärvi (this paper). Incl. Fe''' 0.01, Fe'' 0.06, Mg 0.03.
9. Allt á Mhuillin, Loch Borolan, Assynt, Scotland. Anal. Stewart (1941). Incl. Sr 0.03.
10. Ilmen Mts., Ural. Anal. Morachevsky. Incl. Mg 0.03, Fe''' 0.03.
11. Ilmen Mts., Ural. Anal. Morachevsky. Incl. Mg 0.03, Fe''' 0.01.
12. Ilmen Mts. Ural. Anal. Kosakova 1936 (Bonshtedt-Kupletskaya and Arest-Jakubowitsch, 1947). Incl. Fe''' 0.01, Mg 0.07.
13. Vishnevyh gor, Ural. Anal. Knipowitsch 1931 (Bonshtedt-Kupletskaya and Arest-Jakubowitsch, 1947). Incl. Ti 0.01, Fe''' 0.04, Fe'' 0.02, Mn 0.01, Mg 0.02.

Figures in columns 1—4, 7, 9—11 in Table 2 have been taken from Phoenix and Nuffield (1949).

This table shows the same general tendency. The few exceptions are evidently caused by impurities in the material used for analysis and also by considerable variation in the water content, the effect of which on the much more irregular changes of specific gravity is obvious.

In order to test the indentification of the mineral in question, a powder pattern was prepared of the material used for analysis and, for comparison, of a cancrinite from Mitchfield, Maine, U.S.A. (No. 2727, Institute of Geology collection, Helsinki). The two powder patterns are identical. The apparatus used was the Philips Geiger counter x-ray spectrometer, CuKα-radiation with Ni-filter.

Table 3. Powder diffraction data and unit cell dimensions for sulphatic cancrinite from Iivaara

hkl	d	I	hkl	d	I	Unit cell dimensions
110	6.34	5	002	2.589	4	$a_0 = 12.74$
200	5.49	3	302	2.521	2	$c_0 = 5.22$
101	4.69	7	112	2.428	4	$c_0/a_0 = 0.4098$
210	4.15	3	321	2.280	2	
300	3.665	7	411	2.173	2	
211	3.238	10	330	2.114	4	
310	3.050	2	222	2.021	2	
301	2.992	2	402	1.883	2	
400	2.755	4	430	1.807	2	
311	2.627	3	520	1.759	3	

Acknowledgments — The author here wants to express his indebtedness to Dr. K. J. Neuvonen for his good counsel in interpreting the powder x-ray diagrams, and to Mr. A. Matisto, M. A., for photographing the thin section.

REFERENCES

Бонштедт-Куплетская, Э. М. и Арест-Якубович, Р. Е., (Bonshtedt-Kupletskaya, E. M. and Arest-Jakubowitsch, R. E.) (1947) О минералах группы канкринита из Вишневых гор на Урале. Доклады Академии Наук СССР. Том LVI. № 5, p. 507.

Phoenix, R and Nuffield, E. W. (1949) Cancrinite from Blue Mountain, Ontario. Am. Mineral., 34, p. 452.

Stewart, F. H. (1941) On sulphatic cancrinite and analcime (eudnophite) from Loch Borolan, Assynt. Min. Mag., 26, p. 1.

7

SANDSTONES IN FINLAND [1]

BY

AHTI SIMONEN AND OLAVI KOUVO

Geological Survey of Finland, Helsinki and Outokumpu Oy., Outokumpu

ABSTRACT

Sedimentary petrographic features and conditions of deposition of the sandstones in Finland are briefly described. The late pre-Cambrian, Jotnian arkosic sandstone of Satakunta shows characteristics of the postgeosynclinal piedmont facies of sedimentation. Arkosic sandstones showing petrographic features similar to those of the Jotnian sandstone also occur in the Muhos formation south of the Oulu river, where the sandstones are closely associated with red silstones and shales. On the basis of a tectonic classification of the sandstones the Muhos formation is closely related to the Jotnian formation, which is postgeosynclinal (postorogenic) in relation to the Karelidic orogeny. In addition to the arkosic sandstones, many small occurrences of pure quartz sandstone are known in Finland. These quartz sandstones are nonorogenic deposits of a stable platform, and they represent the transgression of the Cambrian sea over a peneplained surface.

CONTENTS

INTRODUCTION

Finland is part of an old pre-Cambrian Shield where only very few occurrences of sandstones and associated sedimentary rocks overlie the

---

[1] Received February 12, 1955.

metamorphosed and plutonic pre-Cambrian rocks. The oldest unmetamorphosed sandstone in Finland is the late pre-Cambrian Jotnian formation (Sederholm, 1897). A fossiliferous Cambrian sandstone occurs in Åland (Tanner, 1911) and many small occurrences of pure quartz sandstone in Finland are believed by various authors to be deposits of the transgressive Cambrian sea. Different opinions have been presented, however, regarding the geological age of many unfossiliferous sandstone deposits in Finland. The sandstone occurrences and opinions presented concerning their geological age are listed in Table 1, and the geographic distribution of the rocks is shown in Fig. 1.

Table 1. Sandstone occurrences in Finland

Occurrence	Proposed geological age and references
Satakunta	Cambrian (Gylling, 1887); Jotnian (Sederholm, 1897; Laitakari, 1925; Kahma, 1951)
Muhos	Mesozoic (Brenner, 1941 and 1944); Jotnian (Ödman, Härme, Mikkola and Simonen, 1949; Enkovaara, Härme and Väyrynen, 1953)
Saltvik	Cambrian (Tanner, 1911; Metzger, 1922)
Lemland	Cambrian (Kaitaro, 1949)
Kökar	Cambrian (Sederholm, 1924)
Eckerö	Cambrian (Asklund and Kulling, 1926)
Brändö	Cambrian (Sederholm, 1934)
Åva	Cambrian (Kaitaro, 1953)
Parainen	Mesozoic (Hausen, 1934)
Vestanfjärd	Cambrian (Eskola, 1913); Mesozoic (Hausen, 1934)
Hiittinen	Cambrian (Edelman, 1949)
Hanko	Cambrian (Sederholm, 1913 b)
Lauhavuori	Cambrian (Sederholm, 1913 a); Jotnian (Laitakari, 1925)
Karstula	Cambrian (Sauramo, 1916)

The main purpose of this investigation is to present the sedimentary petrographic characteristics of the unmetamorphosed sandstones in Finland. The new data presented are chiefly based on the microscopical study of thin sections. New, more detailed field work for this study was carried out only in the sandstone area of Lauhavuori in southern Ostrobothnia. On the basis of the sedimentary petrographic features the depositional history of the sandstones is briefly discussed and some suggestions on the correlation and geological age are presented. The lower Cambrian quartz sandstones in Enontekio, Lapland, thoroughly described by Hausen (1942) were not taken for petrographic study presented in this paper.

The present study was planned by the senior author (A. S.), who has written the entire manuscript and is responsible for the conclusions. The junior author (O. K.) did most of the laboratory work.

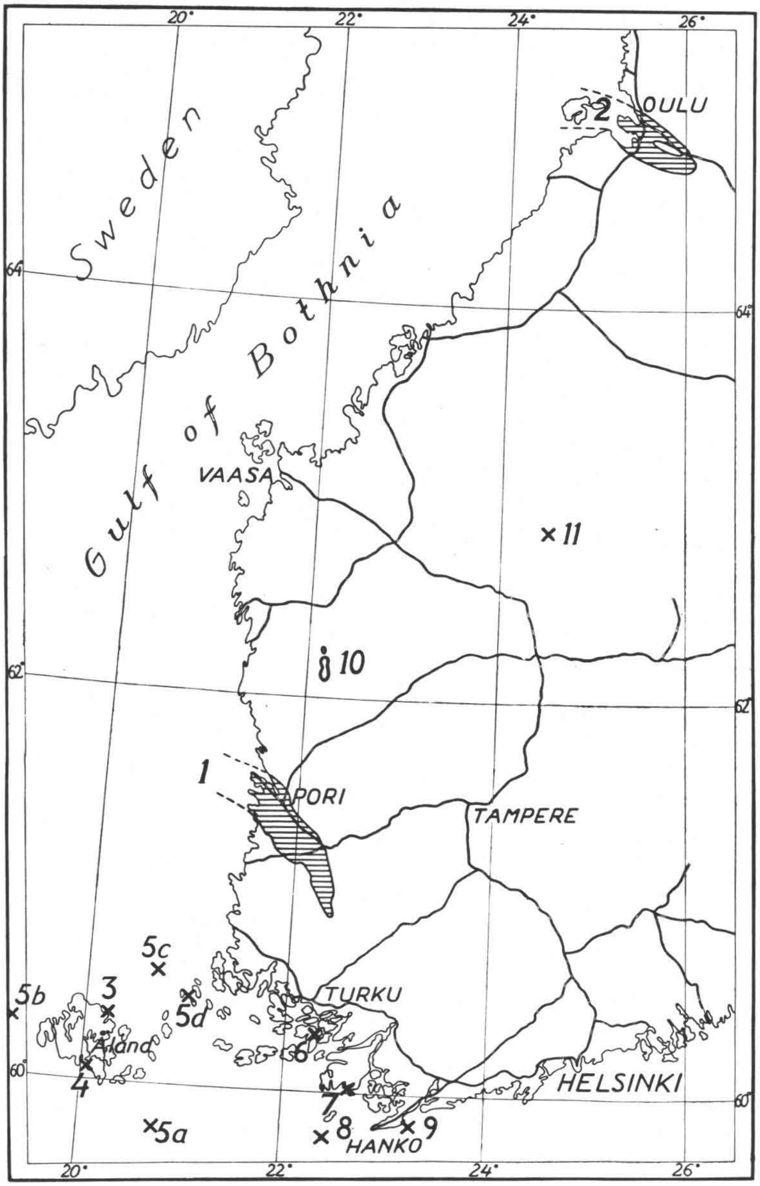


Fig. 1. Location of the unmetamorphosed sandstones in Finland. 1, Jotnian sandstone of Satakunta; 2, Muhos formation; 3—11 quartz sandstones; 3, Saltvik, Åland; 4, Lemland, Åland; 5a, Kökar, Karlbybådar; 5b, Eckerö, Yttre Borgen; 5c, Brändö, Bredan; 5d, Åva, Kummelören; 6, Parainen; 7, Vestanfjärd; 8, Hiittinen; 9, Hanko, Skarvkyrkan; 10, Lauhavuori; 11, Karstula.

JOTNIAN SANDSTONE OF SATAKUNTA

The Jotnian sediment formation of Satakunta occupies a wide area extending from the town of Pori to Lake Pyhäjärvi. The distribution and the field description of the sandstone are given by Laitakari (1925). Red-colored, stratified arkosic sandstones with thin intercalated beds of red or black shales predominate. A conglomerate with quartz pebbles has been found, and conglomeratic sandstone with Archean granite pebbles occur among the Pleistocene glacial boulders derived from the sandstone formation.

The stratification of the sandstone is generally horizontal; the thickness of the deposit is unknown. The gravimetric measurements carried out by the Geodetic Institute suggest, however, a thickness of some hundred meters (cf. Kahma, 1951 p. 10). Many sedimentary structures indicating terrestrial conditions of deposition are described by Laitakari (1925). Current bedding is common and ripple marks (Plate II, Fig. 7). occur in the fine-grained and quartz-rich varieties of the sandstone. Furthermore, mud cracks, clay galls, and rain drop impressions (Plate II, Fig. 8) have been found.

The ripple marks, made by water, in the Jotnian sandstone are symmetric or asymmetric. No definite aeolian ripple marks have been found. The measurements of the wave lengths and amplitudes of the ripple marks from 6 specimens are presented in Fig. 2, which shows that the ripple index (wave length: amplitude) ranges from 7 to 10. The ripple index found in the Jotnian sandstone is typical for the aqueous ripple marks, whereas the index of the aeolian ripple marks ranges usually from 20 to 50 (cf. Twenhofel, 1950).

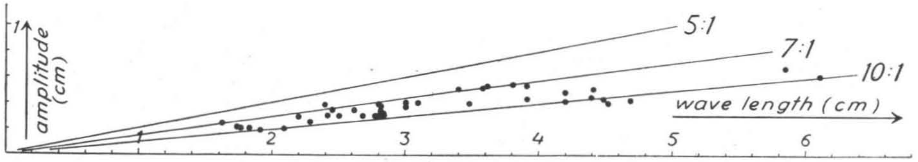


Fig. 2. Wave lengths and amplitudes of the Jotnian ripple marks.

Fourteen specimens of the Jotnian sandstone in Satakunta were selected for more detailed petrographic study. The mineralogical composition of these specimens was determined from the thin sections on the integration stage and the grain size distribution of 4 specimens was measured. The heavy mineral fraction of 2 specimens was separated by centrifuging in Clerici solution. The identification of triclinic and monoclinic modification of the potash feldspar was made by means of the powder X-ray diffraction data using a Geiger Counter Recording X-ray Spectrometer (Norelco).

The afore-mentioned methods of investigation were also used in the petrographic study of the other Finnish sandstones.

Table 2. Mineralogical composition (volume per cent) of the Jotnian sandstone in Satakunta

Locality	Particles					Cement and intergranular matrix
	Quartz	Potash feldspar	Plagioclase	Rock fragments	Accessories	
1. Nakkila, Leistilänjärvi	49.2	20.9	17.4	0.3	0.3	11.9
2. Ulvila	57.1	15.2	24.0	—	—	3.7
3. Säkylä	48.6	30.0	2.3	2.9	1.8	14.4
4. Nakkila, Leistilänjärvi	56.3	21.2	14.8	0.6	1.1	6.0
5. Kauttua	57.0	27.5	1.4	1.1	0.2	12.8
6. Säkylä	64.2	27.9	0.9	5.9	—	1.1
7. Harjavalta	53.1	12.9	9.1	4.5	3.0	17.4
8. Eura, Kiperjärvenoja	65.8	24.5	—	0.4	0.3	9.0
9. Luvia, Korvenkylä, Markkula	65.0	19.6	1.6	0.7	1.0	12.1
10. Kiukainen, Panelia	56.5	16.8	0.3	1.7	0.1	24.6
11. Kiukainen, Panelia	50.2	12.8	1.6	4.3	0.1	31.0
12. Eura, Kiperjärvi	71.3	16.8	0.7	—	0.2	11.0
13. Luvia, Lappäng	70.9	15.5	1.7	2.0	0.2	9.7
14. Yläne	78.2	0.7	4.9	—	2.3	13.9

The mineralogical composition of the Jotnian sandstone is presented in Table 2. Quartz is always the main component of the mineral particles. The content of feldspar varies greatly. The composition of the various specimens ranges from arkose (feldspar over 25 %) through feldspathic sandstone (feldspar 10—25 %) into quartz sandstone (feldspar under 10 %). The quartz to feldspar ratio is presented in Fig. 13. The feldspar is predominantly potash feldspar, but in some few specimens (Table 2, specimens 1, 2 and 4) the content of plagioclase is high. The feldspar is usually rich in hematite pigment, and is clouded and altered. Many specimens contain, however, both clouded and fresh feldspar grains. This indicates that the erosion which acted upon the feldspar-bearing rocks derived detrital material from both unweathered basement complex and its partly decomposed mantle.

The composition of the plagioclase grains varies greatly. Variations from An_5 to An_{30} are observed in the same specimen. To obtain the relative abundance of the different plagioclases the composition of 50 plagioclase grains was determined by means of refractive indices or on the universal stage. The results of measurements are presented in Fig. 3. The plagioclase particles showing the composition An_{7-10}, An_{14-16}, and An_{29-31} are most common. The source of these plagioclase grains is, however, impossible to determine, because the composition of the plagioclase varies irregularly between wide limits in the surrounding country rocks. The absence of basic plagioclase is probably due to the better resistance of alkali feldspars during weathering and transportation.

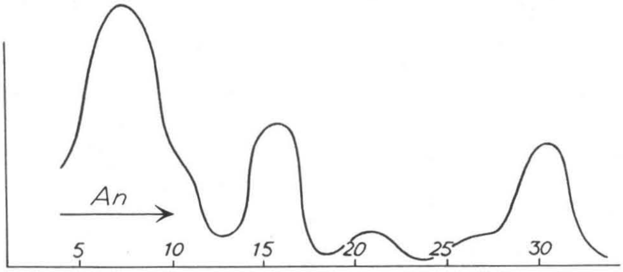


Fig. 3. The relative abundance of different plagioclases in the Jotnian sandstone. Abscissa, composition of plagioclase; ordinate, relative abundance (number of particles).

Altogether 35 potash feldspar particles from four different localities (Harjavalta, Luvia, Sassilanjuopa, and Säkylä) were studied by means of the X-ray powder method to determine the monoclinic or triclinic symmetry. All the potash feldspar grains showed the diffraction lines typical of triclinic modification; no grains with monoclinic symmetry were detected. In this case the presence of the triclinic potash feldspar suggests the source rock of the deposited material, because the potash feldspar of the Archean basement complex surrounding the sandstone area is microcline and that of the normal rapakivi granite, occurring south of the sandstone area, is predominantly orthoclase[1]. The study of the potash feldspar particles in the Jotnian sandstone indicates that the detrital material of the sandstone is derived from the Archean basement. This conclusion is in harmony with the fact that the Glacial boulders of the conglomeratic sandstone contain pebbles of Archean granites.

Biotite, muscovite, and chlorite are the most common accessory minerals of the Jotnian sandstone in Satakunta, but small quantities of many other minerals are also present. Laitakari (1932) studied the heavy minerals of 2 sandstone specimens from Yläne and found the following: garnet, ilmenite, magnetite, zircon, topaz, and monazite. According to Laitakari (1932), the presence of topaz suggests that the rapakivi granite south of the sandstone deposit was the source rock of the detrital material, because topaz occurs in some rapakivi varieties. As, however, topaz also occurs in certain Archean pegmatites, the conclusion is not quite decisive. On the other hand, the garnet is surely derived from the metamorphic Archean rocks.

The present authors have separated the heavy minerals from two sandstone specimens by means of Clerici solution (sp. gr. = 2.9). The heavy minerals observed in the specimens studied are listed below:

---

[1] The study of the potash feldspars of the Finnish rapakivi granites is in the research program of Dr. K. J. Neuvonen, Geological Survey of Finland. According to an oral communication of Dr. Neuvonen, the potash feldspar in the normal Laitila rapakivi, south of the sandstone occurrence, is predominantly monoclinic.

Sandstone, Harjavalta: biotite, chlorite, magnetite, garnet, rutile, sphene, apatite, sillimanite, tourmaline, and epidote;

Sandstone, Nakkila, Leistilänjärvi: biotite, hornblende, magnetite, rutile, zircon, tourmaline, sphene, and apatite.

All these minerals are present in the Archean rocks surrounding the sandstone occurrence and the presence of the garnet and sillimanite is especially diagnostic of the high-rank metamorphic source rocks. No heavy minerals definitely indicating detrital material derived from the rapakivi granites have been found in the specimens studied by the present authors. In view of the study of heavy minerals and feldspars it seems reasonable that at least the main part of the detrital material of the sandstone is derived from the Archean basement complex.

Small rock fragments of black shale and fine-grained quartzite also form particles of the sandstone. The fragments of black shale are angular and those of the quartzite are rounded. The black shale particles are petrographically similar to the interbeds of black shale in the sandstone and they are considered as products of channeling of the deposited sandstone with interbeds of clayey material. The quartzite particles were derived from the metamorphic terrain.

The intergranular matrix is rich in quartz. Minute flakes of sericite and chlorite are also sometimes abundant. They probably were formed from original argillaceous material. The content of mica minerals in the matrix is especially abundant in the sandstones whose intergranular matrix occupies over 20 volume per cent of the whole rock. In some few cases introduced siliceous cement has caused the secondary outgrowth of the quartz particles.

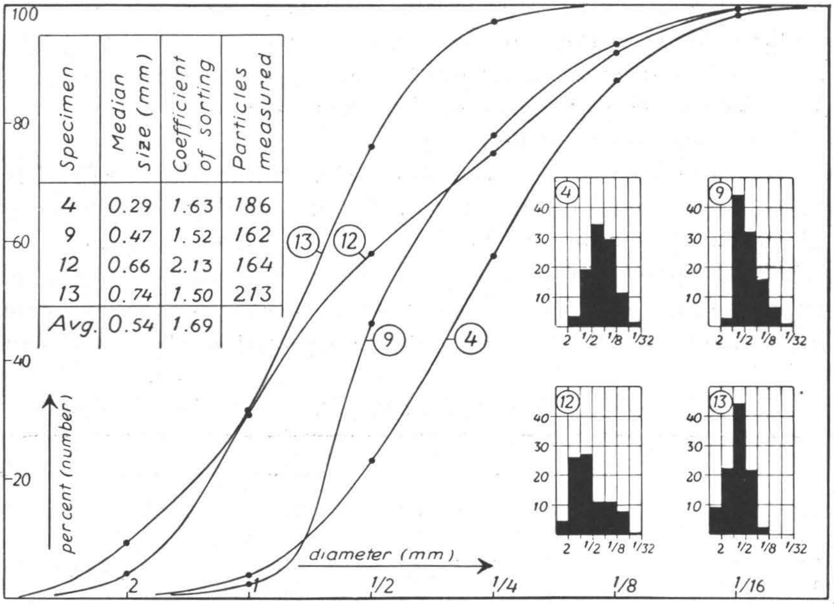
Table 3. Chemical composition of the arkosic sandstones

	1	2	3	4	5
SiO_2	79.30	81.10	82.14	73.32	75.5
TiO_2	0.22	0.09	—	—	—
Al_2O_3	9.94	9.64	9.75	11.31	11.4
Fe_2O_3	1.00	0.28	1.23	3.54	2.4
FeO	0.72	0.65		0.72	
MnO	0.02	0.03	—	tr.	0.2
MgO	0.56	0.76	0.19	0.24	0.1
CaO	0.38	0.80	0.15	1.53	1.6
Na_2O	2.21	1.78	0.50	2.34	2.0
K_2O	4.32	3.79	5.27	6.16	5.6
P_2O_5	0.05	0.08	0.12	—	tr.
H_2O+	0.55	0.77	0.64	0.30	0.6
H_2O-	0.41	0.17			
CO_2	—	—	0.19	0.92	0.4
	99.68	99.94	100.18	100.38	99.8

1. Jotnian sandstone. Köyliö, Muurunmäki, Anal. H. B. Wiik (Lokka, 1950).
2. Jotnian sandstone. Köyliö, Tuiskula. Anal. H. B. Wiik (Kahma, 1951).
3. Torridonian sandstone. Kinlochewe. Mackie (1901).
4. Old Red Sandstone. Foyers, Loch Ness. Mackie (1901).
5. Average arkose. Pettijohn (1949).

Chemical analyses of the Jotnian sandstone in Satakunta are presented in Table 3 (Anal. 1—2). These analyses are very similar and are also similar to the chemical composition of arkoses in thick accumulations of other geological formations. The chemical analyses of the well-known classical Torridonian and Devonian Old Red Sandstone arkose deposits and the average chemical composition of arkose given by Pettijohn (1949) are presented for comparison (Table 3, Anal. 3—5). Chemically the Jotnian sandstone is characterized by the high content of alkalies. The low value of the ratio $Al_2O_3 : Na_2O$ shows that the effectiveness of the weathering has not been strong.

The Jotnian sandstone in Satakunta is usually coarse-grained. Most of the particles are subangular or subrounded (Plate I, Figs. 1—2). These textural features suggest that the abrasion history was rather brief.



Specimen	Median size (mm)	Coefficient of sorting	Particles measured
4	0.29	1.63	186
9	0.47	1.52	162
12	0.66	2.13	164
13	0.74	1.50	213
Avg.	0.54	1.69	

Fig. 4. Grain size distribution of the Jotnian sandstone in Satakunta. The numbers of the specimens refer to Table 2.

The grain size distribution of 4 specimens is presented in Fig. 4, which shows cumulative grain size curves, histograms, median grain size, and coefficients of sorting. The number of sand particles in different size groups was determined from the thin sections. The grain size varies between wide limits and the size distribution indicates that the Jotnian sandstone is not well sorted.

Mineralogical and chemical composition of the Jotnian sandstone in Satakunta is typical for sandstones of the arkose suite. Some mineralogical characteristics (triclinic symmetry of the potash feldspar particles and presence of garnet and sillimanite as heavy minerals) show that the source of the detrital material was mainly the metamorphic Archean basement complex. The quartz to feldspar ratio and the textural features, poor roundness and sorting, show that the Jotnian sandstone belongs to the class of immature sands which are products of rapid erosion and deposition controlled by the conditions of steep relief. The sedimentary structures (ripple marks, cross bedding, and mud cracks) and the red color indicate a terrestrial oxidizing environment in the source area and at the place of deposition. All these features of the Jotnian sandstone in Satakunta are similar to the arkoses that are characteristic of thick terrestrial piedmont facies in postgeosynclinal basins (cf. Pettijohn, 1949).

MUHOS FORMATION

Unmetamorphosed conglomerates, sandstones, siltstones, and shales occupy a wide area in northern Ostrobothnia south and southeast from the town of Oulu. These so-called Muhos sediments are protected from erosion, because they are downfaulted along the northern boundary. The Muhos formation is covered by thick accumulation of the Pleistocene deposits and only one exposure is found in the valley of the Oulu river at Kieksi in the parish of Muhos. The distribution of the Muhos rocks has been studied

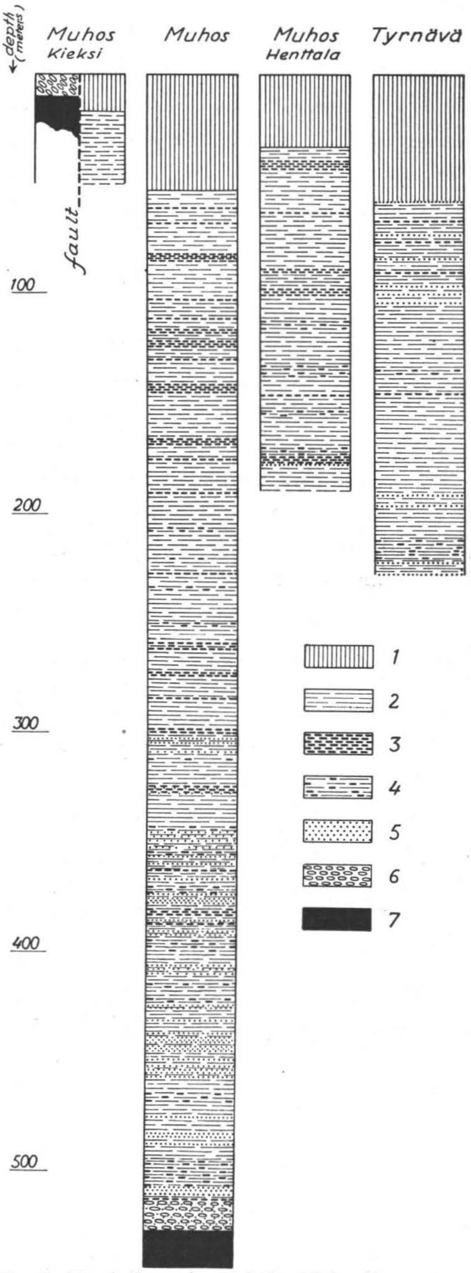


Fig. 5. Vertical sections of the Muhos formation. 1, Pleistocene deposits; 2, red siltstones and shales; 3, grey and green shales; 4, intimate alternation of red and grey shales; 5, sandstone; 6, conglomerate; 7, Archean rocks. This figure is based on the published data of Th. Brenner (1941 and 1944) and on the unpublished reports of the drill holes made by the Geological Survey and Pargas Kalkberg A. B.

by means of drill holes, geomorphological, and glaciogeological data. The material of these investigations has been studied by Veikko Okko, M. A., who will present his observations on the distribution and manner of occurrence of the Muhos formation in the near future. Preliminary reports of the Muhos sediments have been published by Brenner (1941 and 1944).

The vertical variations in the stratigraphic column are presented in Fig. 5. The Muhos formation begins with basal conglomerates and coarse-grained arkoses lying unconformably upon the Archean basement complex. According to Brenner (1944), the polymictic basal conglomerate at Kieksi is separated by a vertical fault from the corresponding basal beds which are situated in the drill hole at Muhos at a level about 500 meters deeper. The basal conglomerate contains rounded and angular pebbles of granite and schist and its matrix is brown-colored arkose rich in silty material. The conglomerate-arkose association forms an accumulation about 20 meters thick at the base of the Muhos formation and it is overlain by thick deposits of red, brown or greyish green siltstones and shales with thin interbeds of arkosic sandstone.

Table 4. Proportions of common types of sedimentary rocks in drill holes of the Muhos formation

%	Muhos	Muhos Henttala	Tyrnävä
Red and brown siltstones and shales	81.5	86.8	91.0
Grey and green shales	11.1	13.2	5.3
Sandstones and conglomerates	7.4	—	3.7

The proportions of common types of sediment in the Muhos formation are presented in Table 4. The red and brown siltstones and shales are predominant, forming 80—90 per cent of the total thickness. Grey and green beds do not form thick, continuous accumulations, but they occur as thin intercalations through the whole section. Sandy material is present in many siltstones and it is associated predominantly with the red beds. The arkosic sandstone occurs as thin interbeds especially in the lower part of the stratigraphic column (cf. Fig. 5).

Altogether 19 complete chemical analyses of the sedimentary rocks in the drill hole Muhos were published by Lokka (1950). These analyses, giving a good idea of the chemical characteristics of the Muhos rocks are presented graphically in Fig. 6. Most of the analyses (Anal. 3—16) represent red or green siltstones and shales. Three analyses (anal. 17—19) represent arkosic sandstones and two (Anal. 1—2) are gypsiferous shales.

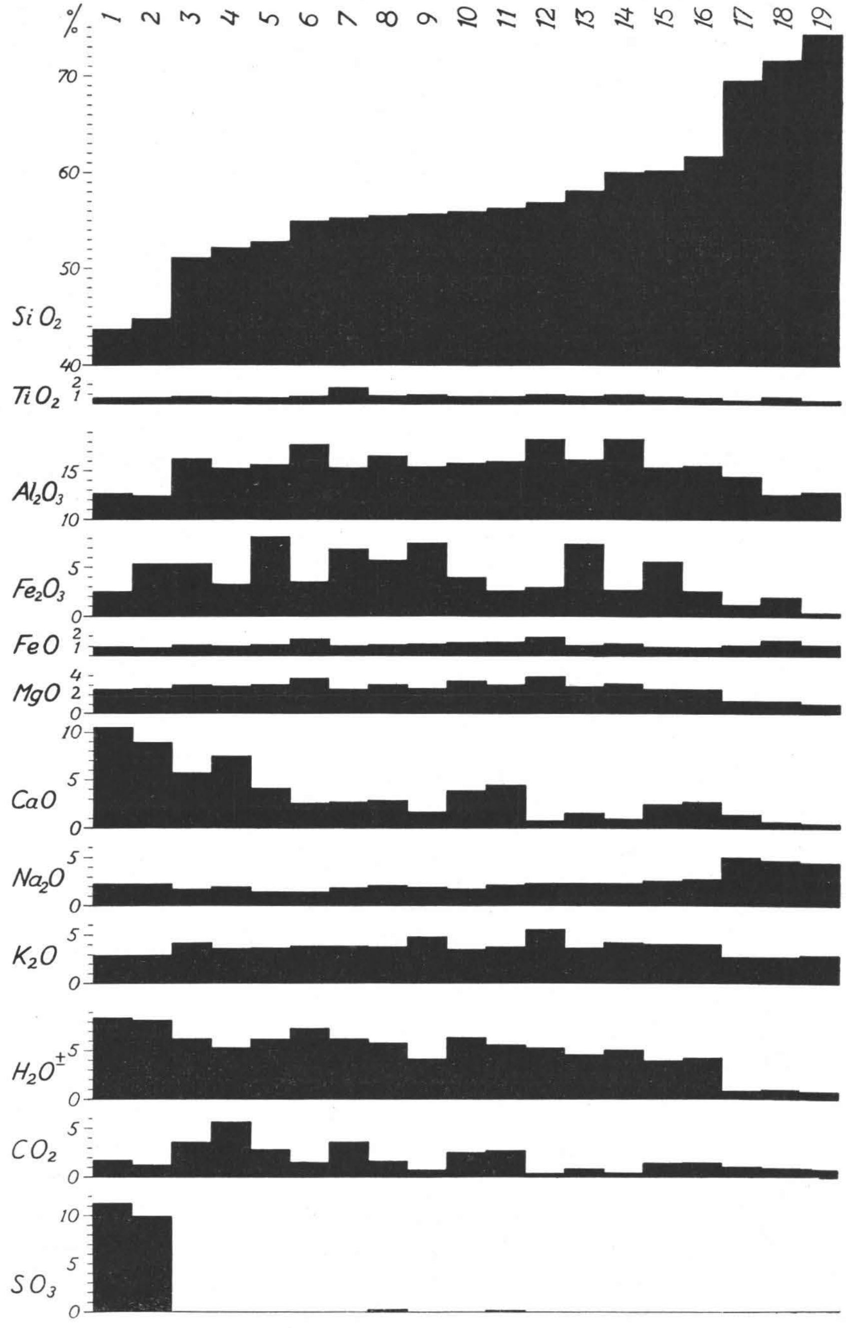


Fig. 6. Chemical composition of the Muhos rocks. Based on the chemical analyses published by Lokka (1950). 1—2 gypsiferous shales; 3—16, shales and siltstones; 17—19, sandstones. 3, 5, 7, 9 and 13, red beds; 4, 6, 10, 12 and 14, greenish grey beds; 8, 11, 15 and 16, red beds with grey spots.

The most remarkable variations in the chemical composition of the shales and siltstones are caused by the different contents of Fe_2O_3, CaO and CO_2. The red and brown beds are regularly characterized by a higher content of Fe_2O_3 than those of grey and green shales. A small amount of CO_2 is always present and the variations in the content of CO_2 appear clearly also in the content of CaO. The percentage of K_2O in the shales is higher than that of Na_2O, but this ratio is reverse in the sandstones. The relatively high content of Na_2O in the shales and siltstones is due to the presence of unweathered feldspar, which shows that the sediment had not undergone complete chemical weathering.

Table 5. Average compositions of the Muhos rocks

	1	2	3	4	5
SiO_2	56.28	54.68	56.09	58.10	71.86
TiO_2	0.88	0.98	0.85	0.65	0.50
Al_2O_3	16.32	15.83	17.14	15.40	13.28
Fe_2O_3	4.91	7.14	3.30	4.02	1.17
FeO	1.21	1.10	1.43	2.45	1.29
MnO	0.08	0.07	0.10	—	0.07
MgO	3.03	2.81	3.42	2.44	1.20
CaO	3.11	3.14	3.07	3.11	0.83
Na_2O	2.05	1.87	1.93	1.30	4.75
K_2O	4.06	4.06	4.14	3.24	2.75
P_2O_5	0.38	0.41	0.33	0.17	0.48
H_2O	5.40	5.33	5.89	5.00	0.89
CO_2	2.00	2.11	2.05	2.63	0.55
SO_3	0.02	—	—	0.64	—
BaO	—	—	—	0.05	—
C	—	—	—	0.80	—
Cl	0.01	—	0.01	—	—
	99.74	99.53	99.75	100.00	99.62

1. Average shale, Muhos formation. Anal. 3—16 in Fig. 6.
2. Average red shale, Muhos formation. Anal. 3, 5, 7, 9 and 13 in Fig. 6.
3. Average greyish green shales, Muhos formation. Anal. 4, 6, 10, 12, and 14 in Fig. 6.
4. Average shale. Clarke (1924).
5. Average sandstone, Muhos formation. Anal. 17—19 in Fig. 6.

Average composition of the shales in the Muhos formation (Table 5) is very similar to the average shale presented by Clarke (1924). The most striking difference between the average composition of red and green beds appears in the content of Fe_2O_3.

On the basis of the chemical analyses published by Lokka (1950) the ratio of iron in Fe_2O_3 and FeO in the red and green shales of the Muhos

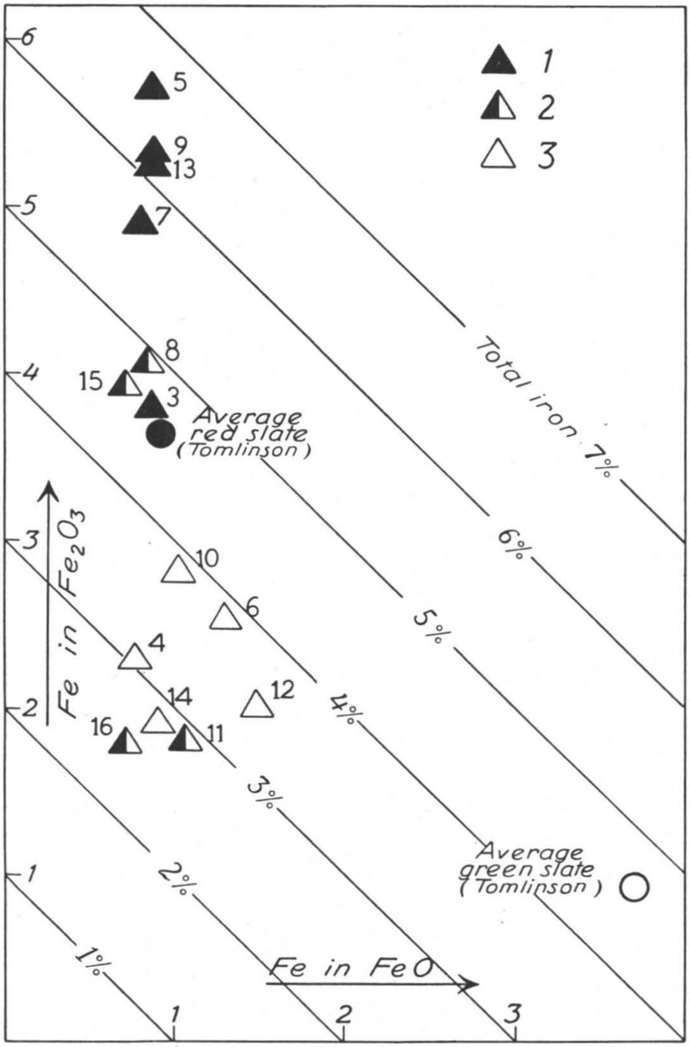


Fig. 7. Relation of iron in Fe_2O_3 and FeO in the red and greyish green shales of the Muhos formation. 1, red beds; 2, red beds with grey spots; 3, greyish green beds. The numbers 1—16 refer to the analyses in Fig. 6.

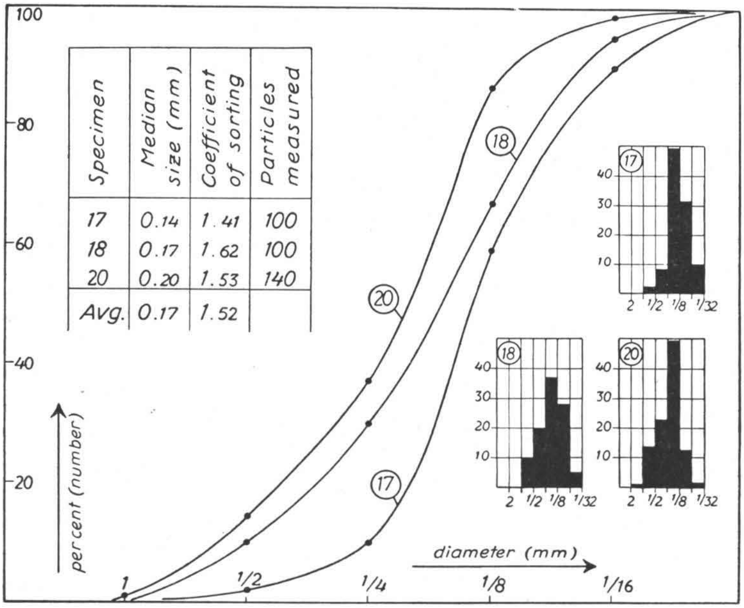
formation is plotted in Fig. 7. The total content of iron in the red beds is much higher than that of the greyish green beds. These data contradict the observations of Tomlinson (1916) who has shown that the red shales do not contain more total iron than the green and black shales, but the different color is only due to the different degree of oxidation.

Table 6. Mineralogical composition (volume per cent) of the sandstones and sandy siltstones in the Muhos formation

<table>
<tr><th rowspan="2">Locality</th><th colspan="5">Particles</th><th rowspan="2">Cement and intergranular matrix</th><th rowspan="2">Calcite</th><th rowspan="2">Gypsum</th></tr>
<tr><th>Quartz</th><th>Potash feldspar</th><th>Plagioclase</th><th>Rock fragments</th><th>Accessories</th></tr>
<tr><td>15. Muhos, Kieksi</td><td>20.4</td><td>1.5</td><td>3.2</td><td>15.6</td><td>2.8</td><td>56.5</td><td>—</td><td>—</td></tr>
<tr><td>16. Muhos, depth 53.2 m</td><td>88.8</td><td>1.4</td><td>—</td><td>—</td><td>0.1</td><td>9.7</td><td>—</td><td>—</td></tr>
<tr><td>17. Muhos, depth 254 m</td><td>30.9</td><td>—</td><td>8.9</td><td>0.2</td><td>1.2</td><td>57.1</td><td>1.7</td><td>—</td></tr>
<tr><td>18. Muhos, depth 305 m</td><td>12.5</td><td>0.2</td><td>7.3</td><td>4.8</td><td>0.4</td><td>74.8</td><td>—</td><td>—</td></tr>
<tr><td>19. Muhos, depth 310 m</td><td>21.7</td><td colspan="2">23.6</td><td>3.3</td><td>0.8</td><td>49.2</td><td>1.4</td><td>—</td></tr>
<tr><td>20. Muhos, depth 347.8 m</td><td>40.4</td><td colspan="2">18.4</td><td>4.8</td><td>1.7</td><td>8.4</td><td>—</td><td>26.3</td></tr>
<tr><td>21. Muhos, depth 509.3 m</td><td>26.4</td><td>6.3</td><td>9.7</td><td>34.0</td><td>0.9</td><td>22.7</td><td>—</td><td>—</td></tr>
<tr><td>22. Liminka, depth 65 m</td><td>40.4</td><td>10.3</td><td>28.5</td><td>—</td><td>0.5</td><td>15.2</td><td>5.1</td><td>—</td></tr>
<tr><td>23. Liminka, depth 109 m</td><td>8.9</td><td>1.9</td><td>2.7</td><td>2.0</td><td>0.1</td><td>84.2</td><td>0.2</td><td>—</td></tr>
<tr><td>24. Liminka, depth 237 m</td><td>25.0</td><td>3.4</td><td>9.5</td><td>15.0</td><td>0.2</td><td>46.9</td><td>—</td><td>—</td></tr>
</table>

Altogether 10 specimens of the Muhos rocks containing sandy particles were selected for more detailed petrographic study. The mineralogical composition presented in Table 6 shows that the sandstones range from arkoses through silty sandstones into siltstones. Only one thin bed of a quartz sandstone 7 cm in thickness (Table 6, n:o 16) has been found. Quartz, feldspar, and small rock fragments of granite and schist form the sandy darticles of the arkoses. Feldspar is mainly plagioclase which is unusual for arkoses generally. This indicates that the detrital material is derived from a rock complex where the content of plagioclase has been higher than that of potash feldspar. Furthermore, decomposition of the deposited material was very incomplete. The composition of the plagioclase varies from An_{10} to An_{34}. Grains showing the composition An_{20-30} are most common. Fresh and much altered particles of plagioclase occur together. The ratio of quartz to feldspar in the specimens studied is presented in Fig. 13 and the values obtained are very similar to those of the Jotnian sandstone in Satakunta. The high content of the intergranular silty material is characteristic of the sandstones in the Muhos formation.

The chemical composition of the Muhos sandstone (cf. Fig. 6 and Table 5) is related to that of the arkose sandstone in Satakunta. The abundance of silty matrix in the Muhos sandstone produces the higher content of Al_2O_3 and the predominance of plagioclase among the detrital feldspar causes an extremely high content of Na_2O, which indicates the immaturity of the Muhos sandstone.



Specimen	Median size (mm)	Coefficient of sorting	Particles measured
17	0.14	1.41	100
18	0.17	1.62	100
20	0.20	1.53	140
Avg.	0.17	1.52	

Fig. 8. Grain size distribution of the sandstones in the Muhos formation. The numbers of the specimens refer to Table 6.

Stratification and cross bedding are the only sedimentary structures observed in the Muhos sandstone. The sandy particles are mostly subangular or subrounded (Plate I, Figs. 3—4) and the grain size distribution (Fig. 8), determined from the thin sections of 3 specimens, shows that the sandstone is not well sorted. These textural features also point to the immaturity of the Muhos sandstone.

The petrographic characteristics of the sandstone in the Muhos formation are related to those of the Jotnian arkosic sandstone in Satakunta. Accordingly, the Muhos sandstone belongs to the immature sandstones which represent tectonic accumulations of slightly decomposed detrital material. The sandstones of the Muhos formation deviate, however, from those in Satakunta through a higher content of intergranular matrix and detrital plagioclase. Some silty varieties resemble graywackes, but their red color and extremely high content of Fe_2O_3 are not characteristic of sandstones of the graywacke suite. The predominance of red beds in the Muhos formation indicates an oxidizing condition of terrestrial sedimentation. The association of red shales, siltstones, and sandstones is typical of many floodplain deposits in the piedmont facies in postgeosynclinal basins.

QUARTZ SANDSTONES

COASTAL AREA OF SOUTHWESTERN FINLAND

Many small exposures of quartz sandstone are known from the coastal area of southwestern Finland. All of these sandstones are protected from erosion, because they occur as fillings of fault cracks in the pre-Cambrian rocks or are fillings of cavities in the pre-Cambrian crystalline limestone. The ancient peneplain on which these sandstones were deposited has stood slightly above the present surface. A brief petrographic description of the different sandstone occurrences is given below.

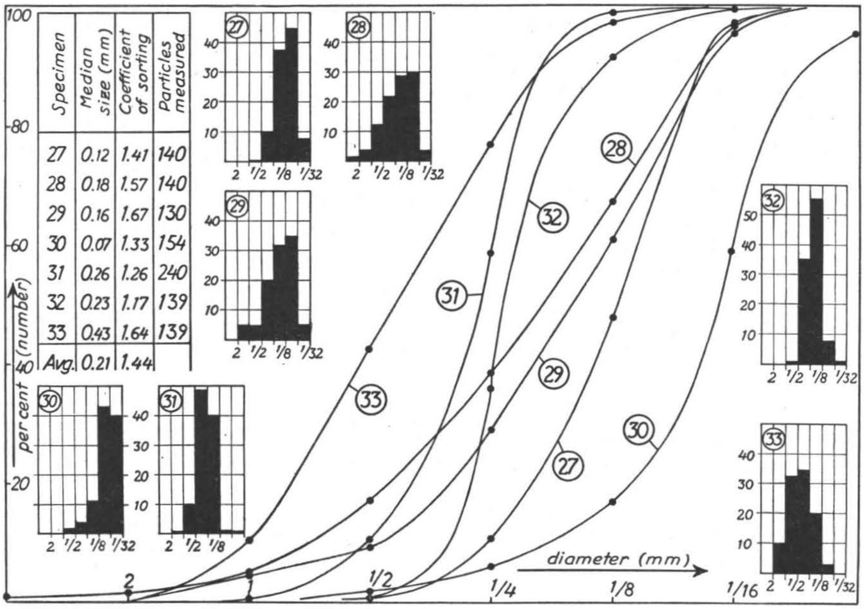
Saltvik. This yellowish grey sandstone occurs as filling of vertical fissures in the late pre-Cambrian rapakivi granite. Altogether 8 sandstone dikes, measuring from a few centimeters up to 60 cm in thickness, have been found in Långbergsöda, Saltvik. One of these dikes contains fossil Brachiopods of Cambrian age. The mode of occurrence of the sandstone has been reported by Tanner (1911) and the fossils have been described by Metzger (1922).

Table 7. Mineralogical composition (volume per cent) of the quartz sandstones in the coastal area of southwestern Finland

Locality	Particles				Cement and intergranular matrix
	Quartz	Feldspar	Rock fragments	Accessories	
25. Saltvik	73.8	5.0	8.8	1.0	11.4
26. Saltvik	86.2	1.9	0.5	0.5	10.9
27. Saltvik	72.2	1.5	0.5	1.1	24.7
28. Saltvik	66.6	1.3	12.9	0.6	18.6
29. Saltvik	81.3	1.3	1.5	0.5	15.4
30. Lemland, Lemböte	74.8	9.5	1.7	0.8	13.2
31. Vestanfjärd, Lammala	90.9	—	—	0.3	8.8
32. Hiittinen, Sisan	86.3	0.2	1.5	—	12.0
33. Hanko, Skarvkyrkan	81.2	0.4	0.3	0.4	17.7

The mineralogical composition of 5 sandstone specimens from Saltvik is presented in Table 7. The sandstone is rich in quartz and poor in feldspar. The quartz particles are rounded or subrounded. The feldspar is red-colored potash feldspar whose X-ray diffraction data shows a monoclinic symmetry, indicating that the feldspar is derived from the surrounding rapakivi granite. Small rock fragments of rapakivi granite are also present. The angularity of most of the feldspar particles and rock fragments indicates that they are not primary components of the highly decomposed sandstone, but are probably derived from the walls of the sandstone dike. The rich occurrence of tourmaline as heavy mineral is characteristic. Also minute amounts of zircon, rutile, and apatite have been found in the heavy mineral fraction. The cement is siliceous and a secondary outgrowth of the quartz particles is common.

The grain size distribution and sorting (Fig. 9) varies greatly in different specimens. It is probable that the observed grain size distribution does not represent the original grain size distribution, because many different sizes may have been mixed together when the sand filled the open vertical fissure.



Specimen	Median size (mm)	Coefficient of sorting	Particles measured
27	0.12	1.41	140
28	0.18	1.57	140
29	0.16	1.67	130
30	0.07	1.33	154
31	0.26	1.26	240
32	0.23	1.17	139
33	0.43	1.64	139
Avg.	0.21	1.44	

Fig. 9. Grain size distribution of the sandstones in the coastal area of southwestern Finland. The numbers of the specimens refer to Table 7.

Lemland. Two sandstone dikes cutting the rapakivi granite of Åland have been found by Kaitaro (1949) in Lemland. The mineralogical composition of the sandstone at Lemböte in Lemland is given in Table 7. Quartz is the main component, but the content of potash feldspar is also remarkably high. The sandstone is rather fine-grained and most of the quartz particles are rounded. The sandstone also contains larger subangular particles of quartz, potash feldspar, and graphic rapakivi granite. It seems probable that these subangular particles were derived from the surrounding rapakivi, and that they did not take part in the long abrasion history of the small quartz particles. The introduced cement is siliceous and a secondary outgrowth of the quartz particles is common. Sorting (Fig. 9) is much better than in the sandstone of Saltvik.

Other sandstone dikes in the Åland area. In addition to the occurrences mentioned, narrow sandstone dikes cutting the rapakivi granite of Åland have been found in the following places: on the skerry of Karlbybådar in Kökar (Sederholm, 1924), on the skerry of Yttre Borgen in Eckerö (Asklund and Kulling, 1926), and in the islands of Bredan and Venningbådan in the northeastern part of the Åland area (Sederholm, 1934). Furthermore, Kaitaro (1953) has found a sandstone dike in the late pre-Cambrian Åva granite on the skerry of Kummelören. The petrographic characteristics of these quartz sandstones are similar to those of the afore-described sandstones in Saltvik and Lemland.

Parainen. The petrography of the sandstone in Parainen has been described by Hausen (1934) and the following short description is based on his investigation. The sandstone bed is as much as 1 meter thick and overlies a thin kaolin bed. Both of these beds occur as fillings of a cavity in the pre-Cambrian limestone. The sandstone is a pure quartz sandstone with no feldspar. The heavy minerals are psilomelane, zircon, and magnetite. Limonite and introduced siliceous matter form the cement. The sandstone is rather coarse-grained, containing mostly particles measuring from 0.3 to 1 mm in diameter. The roundness of the particles varies greatly. Cross bedding is the only sedimentary structure observed.

Vestanfjärd. Two occurrences of sandstone filling cavities in the pre-Cambrian limestone at Illo and Lammala in the parish of Vestanfjärd have been thoroughly described by Eskola (1913). The sandstone is brownish yellow and quartz is the main component of the mineral particles (cf. Table 7, N:o 31). Only very few strongly weathered feldspar particles have been found. Small pebbles of quartz, 4—20 mm in diameter, and fragments of grey shale occur occasionally. The heavy minerals observed by Laitakari (1932) are as follows: zircon, rutile, magnetite, and ilmenite. Limonite and introduced siliceous matter usually form the cement, but in a few cases calcite, too, occurs as large crystals. A secondary outgrowth of quartz particles is common. The particles are extremely well rounded and the sandstone is well sorted (cf. Fig. 9).

Hiittinen. A sandstone dike cutting the Archean migmatitic basement has been found by Edelman (1949) on the skerry of Sisan in the parish of Hiittinen. This sandstone is also very pure quartz sandstone (Table 7, N:o 32) with introduced siliceous cement. Small fragments of chert also occur as sandy particles. The particles are subrounded or rounded and the sorting of the sandstone (cf. Fig. 9) is the best among the specimens from the coastal area studied.

Hanko, Skarvkyrkan. This sandstone dike cutting the pre-Cambrian migmatites was found by Sederholm (1913 b) on the skerry 20 km east of the town of Hanko. Petrographically the sandstone is very similar to the other exposures in the coastal area. Quartz is the main detrital component (Table 7, N:o 33). The cement is siliceous and secondary outgrowths on the quartz particles are common. The particles are subrounded and rounded and grades of many sizes are present (Fig. 9).

In addition to the outcrops mentioned, the Glacial boulders of sandstone are common in the coastal area of southwestern Finland. Most of the sandstone boulders consist of red-colored, arkosic sandstone of Jotnian type, but also fossiliferous Cambrian sandstone with calcareous cement is usual in the drift material. Observations on the distribution of the sandstone boulders have been given by many authors (Hausen, 1911; Eskola, 1913; Laitakari, 1925; Metzger, 1927; Hellaakoski, 1930; Edelman, 1951; Marmo and Laitakari, 1952) and it seems probable that sandstone occurs in many places at the bottom of the sea, especially north of the Åland Islands.

The sandstones in the outcrops of the coastal area are very similar to each other and their ratio of quartz to feldspar (Fig. 13) shows a maturity produced by an intense chemical weathering. The textural features, as the good rounding and sorting, are also characteristic of mature sandstones. The cement of the sandstones in the coastal area is commonly siliceous. Calcareous cement has been found only in some varieties of sandstone in Vestanfjärd and in Glacial boulders of the sandstone. These kinds of mature quartz sandstones are typical of the foreland facies of sedimentation during transgression of a stable platform with a low relief. The petrographic similarity to Cambrian sandstones in Sweden thoroughly described by Hadding (1929) and the Cambrian fossils found in the sandstone of Saltvik and in the Glacial boulders of calcitic sandstone indicate a Cambrian age for the sandstones of the coastal area. The mature sandstones of a foreland facies are commonly associated with limestones. It should be noted that Cambro-Silurian limestone is known in Lumparn, Åland, and Glacial boulders of fossiliferous Cambro-Silurian limestone are common in the coastal area of southwestern Finland. The probable association of quartz sandstone and limestone, representing the same facies of sedimentation, also provides evidence that the sandstone outcrops belong to the early Paleozoic age. The fossils of the Cambro-Silurian age found in sandstones and limestones in southwestern Finland have been described by Metzger (1922 and 1927). A Mesozoic age was suggested by Hausen (1934) for the sandstones in Parainen and Vestanfjärd, but no valid evidence has been found for a marine transgression at that time in southwestern Finland.

SANDSTONE OF LAUHAVUORI

The widest occurrence of quartz sandstone in Finland is in Lauhavuori, Southern Ostrobothnia. This sandstone is very resistant to erosion, and forms the wide Lauhavuori hill, which is the highest place (about 230 meters above sea level) in this part of the country.

A geological map of the sandstone area is presented in Fig. 10. This map is based on field work carried out by K. Kauranne, M. A., in the summer of 1951 under the supervision of the senior author (A. S.) of the present paper.

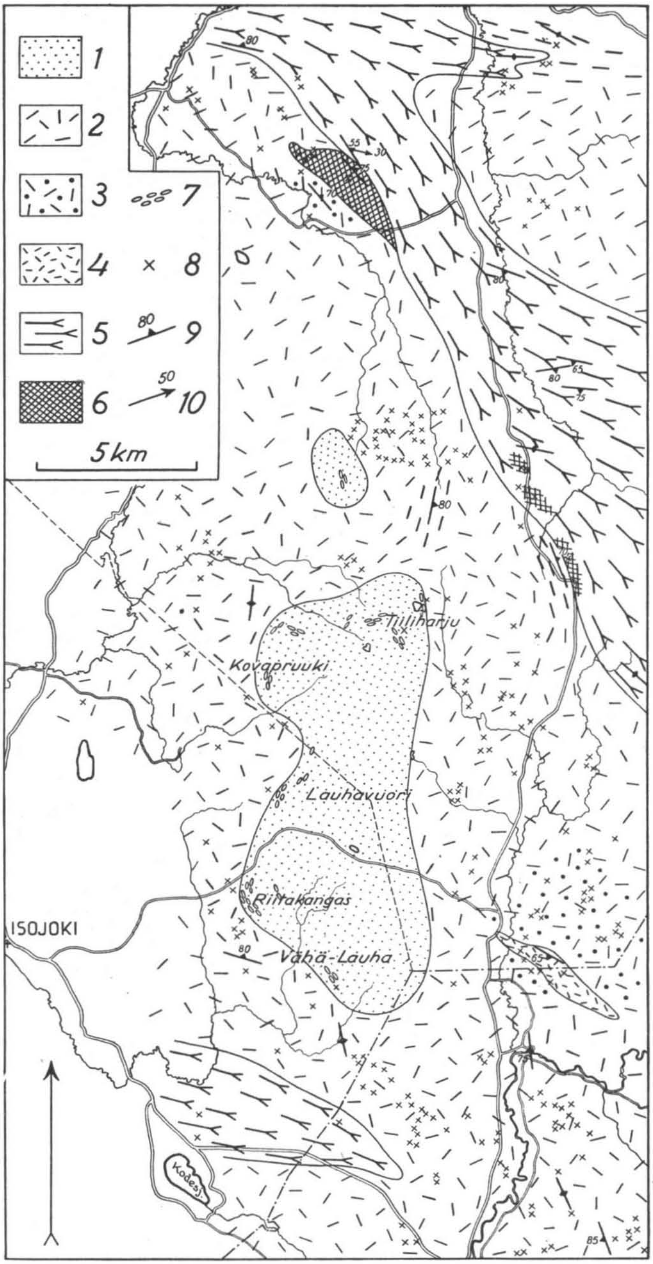


Fig. 10. The geological map of the Lauhavuori sandstone area. 1, sandstone; 2, porphyritic granite; 3, hornblende-bearing porphyritic granite; 4, even-grained granite; 5, gneissose granodiorite; 6, gabbro; 7, boulder field of the sandstone; 8, outcrop; 9, strike and dip of foliation; 10, lineation.

The sandstone occupies about 60 square kilometers. The surrounding country rock consists of pre-Cambrian igneous rocks belonging to the infracrustal rock complex of Central Finland. The sandstone is covered by Pleistocene deposits and only two outcrops of it have been found (Plate III, Fig. 9).

Boulder fields of the late-Glacial shores along the slopes of the Lauhavuori hill have been useful for geological mapping, because the angular boulders of these boulder fields consist exclusively of the local sandstone (Plate III, Fig. 10). These boulder fields, which give an idea about the distribution of the sandstone, are up to 100—250 meters long and 20—50 meters wide and the boulders measure from some dm³ to 2 m³ in volume. The largest boulders have been locally used as millstones.

The boulder train of the Lauhavuori sandstone owing to the movements of the Pleistocene ice-sheet also suggests a wide distribution of the sandstone (Fig. 11). The glacial boulders of the sandstone are most abundant close to the sandstone area, making up more than 1/3 of the total drift material, but farther away they become sparse. The direction of the boulder train agrees very well with the striae observed in the region around the sandstone occurrence.

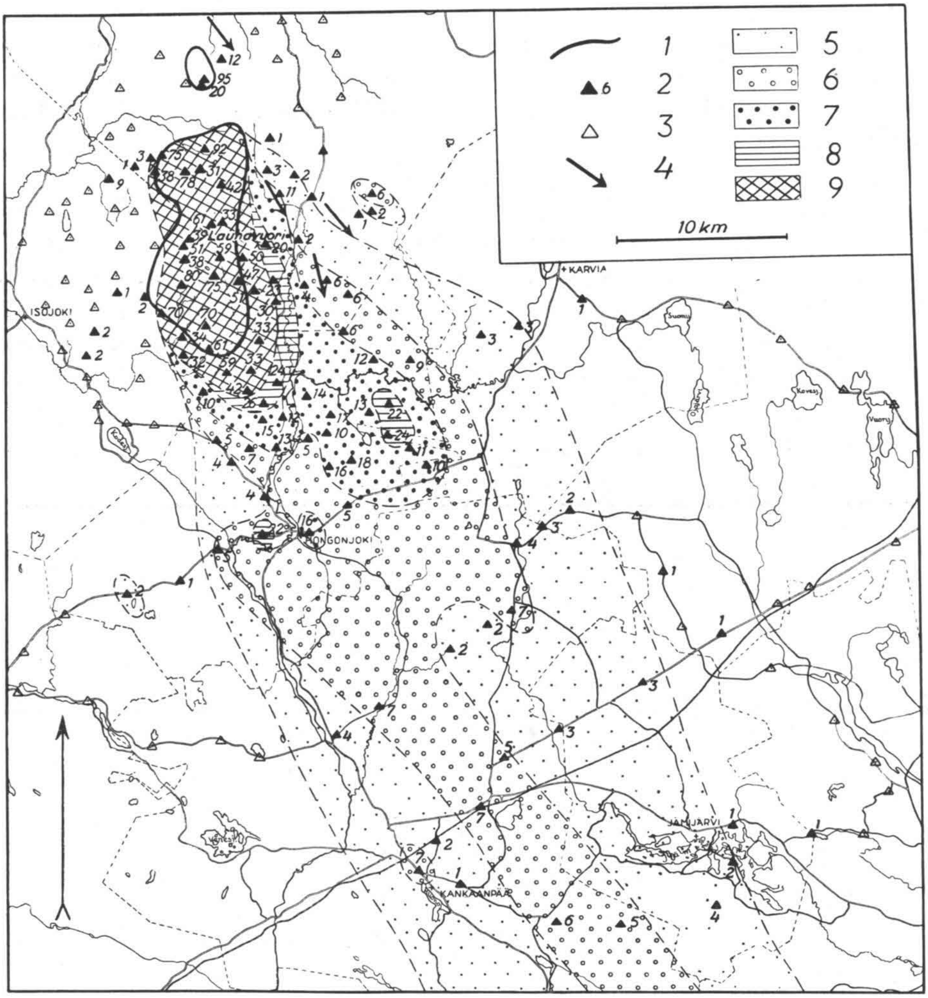


Fig. 11. The Glacial boulder train of the Lauhavuori sandstone. 1, boundary of the sandstone area; 2, percentage of Lauhavuori sandstone boulders in the drift material; 3, no boulders of Lauhavuori sandstone in the drift material; 4, striae; 5—9, contours for the frequence of distribution of Lauhavuori sandstone boulders; 5, 2—5 %; 6, 5—10 %; 7, 10—20 %; 8, 20—30 %; 9, over 30 %.

The sandstone occupies the highest places in the present topography and it occurs as a thin sheet upon the pre-Cambrian basement. The thickness of the sandstone formation is not exactly known, but on the basis of the topographic map of the Lauhavuori hill made by Olander (1934) it is only some ten meters. Outcrops of Archean porphyritic granite appear immediately below many boulder fields of sandstone blocks.

The Lauhavuori sandstone is light-colored with shades of pale red, yellow or brown. The size of the sand particles varies greatly in different beds. Usually the grain size ranges from 0.2 to 0.5 mm, but conglomeratic varieties with well-rounded quartz pebbles also occur. The diameters of the quartz pebbles are commonly 0.5—2 cm but the largest quartz cobble observed was 15 cm in diameter. Conglomeratic sandstones are abundant in the boulder fields of Riitakangas and Vähä-Lauha in the southern part of the sandstone area. A few boulders containing angular fragments of very much altered porphyritic granite have also been found in the boulder field of Vähä-Lauha (Plate IV, Fig. 13). The red-colored porphyritic granite boulders are strongly silicified and the feldspar grains are entirely kaolinized. The matrix is quartz sandstone rich in kaolin. These conglomerate boulders probably represent the basal part of the Lauhavuori sandstone deposited on a deeply weathered surface of porphyritic granite.

Table 8. Mineralogical composition (volume per cent) of the quartz sandstones in Lauhavuori and Karstula

Locality	Particles				Cement and intergranular matrix
	Quartz	Feldspar	Rock fragments	Accessories	
Lauhavuori:					
34. Vähä Lauha	74.3	1.3	5.6	1.0	17.8
35. Riitakangas	71.9	1.4	2.9	0.5	23.3
36. Iso Lauha	73.7	1.1	0.1	—	25.1
37. Tiiliharju	78.7	0.5	0.1	0.1	20.6
38. Riitakangas	73.6	0.4	0.5	0.2	25.3
39. Kovapruuki	70.3	0.2	0.1	0.8	28.6
40. Riitakangas	75.7	0.1	5.6	0.3	18.3
41. Riitakangas	72.8	0.1	0.7	0.2	26.2
Karstula:					
42. Vahankajärvi	75.4	1.1	0.2	0.7	22.6
43. Vahankajärvi	89.3	1.0	0.8	0.3	8.6

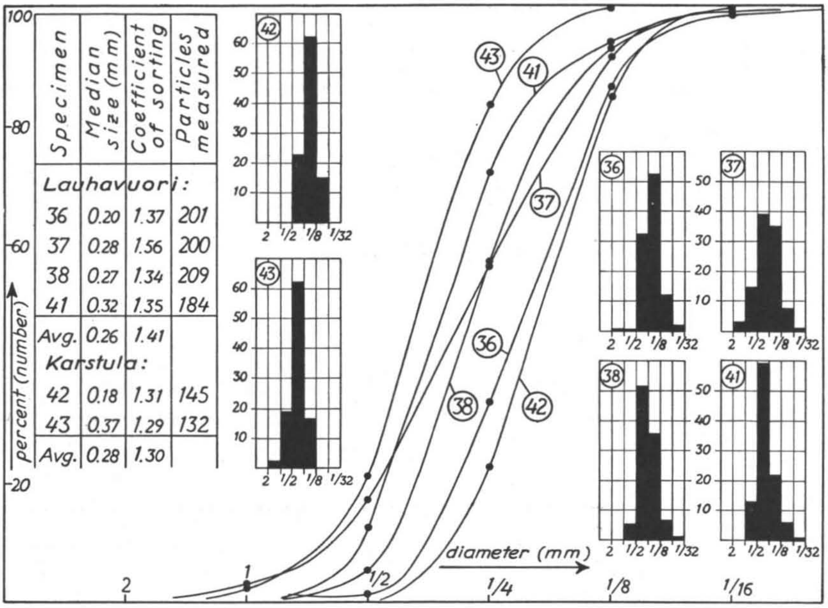
The mineralogical composition of the Lauhavuori sandstone presented in Table 8 is typical of pure quartz sandstones. The content of strongly kaolinized feldspar is very low. Some small siliceous rock fragments are present. The content of the accessory minerals is very low. Laitakari (1932) has found the following heavy minerals: zircon, ilmenite, rutile, and magnet ite. The heavy minerals observed by the present authors in the Lauhavuori sandstone from Tiiliharju are as follows: magnetite, tourmaline, zircon, sphene, and apatite. The cement is siliceous and a secondary outgrowth of the quartz particles is common. The enlargement of the primary quartz particles by secondary outgrowth can easily be observed microscopically in the sandstone specimens, in which the original boundaries of the quartz particles are extremely clearly seen. This increase varies from 10—20 volume per cent.

Table 9. Chemical composition of the Lauhavuori sandstone from Tiiliharju (1) and composition of a clay ball (2) in the sandstone boulder from Riitakangas. Anal. Pentti Ojanperä

	1	2
SiO_2	97.36	70.21
TiO_2	0.05	0.58
Al_2O_3	0.73	10.46
Fe_2O_3	0.63	4.28
FeO	0.14	0.29
MnO	0.01	0.01
MgO	0.01	0.23
CaO	0.04	0.84
Na_2O	0.08	0.20
K_2O	0.19	0.82
P_2O_5	0.02	5.78
H_2O+	0.54	5.18
H_2O-	0.14	0.82
	99.94	99.70

The highly siliceous character of the Lauhavuori sandstone appears also in the chemical analyses presented in Table 9. The analyzed specimen from Tiiliharju contains small amounts of kaolin material.

Stratification (Plate IV, Fig. 12) and cross bedding are common structural features of the Lauhavuori sandstone. Clay balls (Plate IV, Fig. 11) in the sandstone are also found, the greatest diameter of the well-rounded or flattened balls varies from 1 to 6 cm. The material of the clay balls is very fine-grained so that microscopic determination of the minerals is impossible. The chemical composition (Table 9, Anal. 2) shows, however, that the material is a product of strong chemical weathering. The high content of P_2O_5 is remarkable, but it was not possible for the authors to identify by the microscope and X-ray powder diagram the phosphate mineral present in the analyzed clay ball. No microfossils were found in the clay ball separations made and investigated by Dr. K. Mölder.



Specimen	Median size (mm)	Coefficient of sorting	Particles measured
Lauhavuori:			
36	0.20	1.37	201
37	0.28	1.56	200
38	0.27	1.34	209
41	0.32	1.35	184
Avg.	0.26	1.41	
Karstula:			
42	0.18	1.31	145
43	0.37	1.29	132
Avg.	0.28	1.30	

Fig. 12. Grain size distribution of the sandstones in Lauhavuori and Karstula. The numbers of the specimens refer to Table 8.

The mineral particles of the Lauhavuori sandstone are extremely well-rounded (cf. Plate I, Figs. 5—6) and the sorting (Fig. 12) is good.

Mineralogical, chemical, as well as textural features of the Lauhavuori sandstone are typical of mature sandstones which are products of prolonged chemical and mechanical processes during weathering and sedimentation. Determination of the geological age is difficult, because valid evidence is lacking. On the basis of the petrographic features the conditions of sedimentation were, however, similar to the Cambrian quartz sandstones in the coastal area of southwestern Finland. The Lauhavuori hill probably is the place where the Paleozoic (sub-Cambrian) peneplain has not been destroyed by later erosion as is the case in the coastal area where the sandstone dikes indicate that the ancient peneplain originally was a little above of the present surface.

SANDSTONE OF KARSTULA

This sandstone is known only as a rich local occurrence of Glacial boulders. Its main petrographic features were given by Sauramo (1916). The mineralogical composition (Table 8) shows that the Karstula sandstone is very pure quartz sandstone. The cement is siliceous and the secondary outgrowth of the quartz particles is common. Stratification and cross bedding are the structural features observed. The sand particles are extremely well-rounded and the sorting (Fig. 12) is very good. The petrographic characteristics of the Karstula sandstone are very closely related to those of the other quartz sandstones in Finland. The sandstone of Karstula marks, so far as known, the eastern limit of the early Paleozoic transgression in Central Finland.

SUMMARY AND CONCLUDING REMARKS

The sedimentary petrographic features described in this paper show that the unmetamorphosed sandstones in Finland are either arkoses or pure quartz sandstones. Mineralogically the most remarkable difference between these sandstone types is seen in the ratio of quartz to feldspar, which is the index of maturity giving an idea of the effectiveness of the chemical and mechanical processes. Fig. 13 shows that the ratio of quartz to feldspar in the sandstones of Satakunta and Muhos varies within the same limits and the feldspar occupies commonly 20—50 per cent of the total amount of sand particles. Detrital feldspar of the Satakunta sandstone is predominantly potash feldspar, but in the Muhos sandstone feldspar is mainly plagioclase, which is unusual for arkoses generally. The quartz sandstones in Finland form a well separated group, where the content of the feldspar particles is only a few per cent. In two specimens of quartz sandstones (Fig. 13, No 25 and 30) the feldspar content is remarkable high, but a petro-

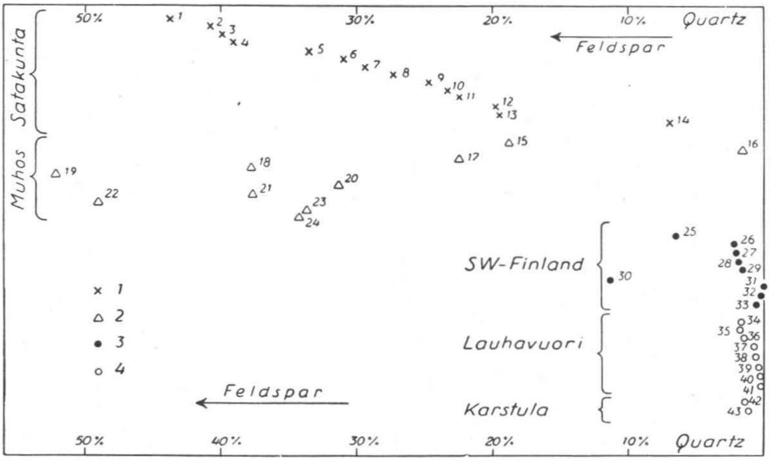


Fig. 13. The ratio of quartz to feldspar in the sand particles of the Finnish sandstones. 1, Jotnian arkose sandstones in Satakunta; 2, sandstones and sandy siltstones in Muhos; 3, quartz sandstones of the coastal area of southwestern Finland; 4, quartz sandstones in Lauhavuori and Karstula. The numbers refer to the specimens in Tables 2, 6, 7, and 8.

graphic study of these specimens (see p. 72—73) revealed that the feldspar has not undergone long abrasion, as have the quartz particles.

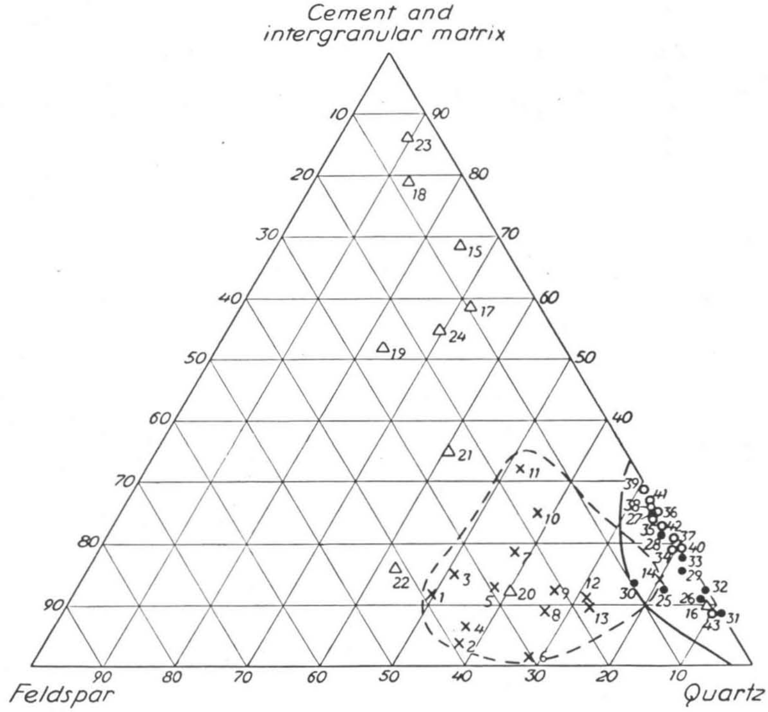


Fig. 14. The proportions of quartz and feldspar particles and matrix in the Finnish sandstones. The key is like that of Fig. 13.

The petrographic characteristics of the sandstones in Finland are also seen in the triangle diagram (Fig. 14), which shows the proportions of the quartz and feldspar particles and the matrix. The quartz sandstones form a clearly separated field in the quartz corner of the diagram. The sandstones of the Muhos formation deviate from those in Satakunta through higher content of intergranular matrix. The sandstones of Satakunta are for the most part typical arkoses, whereas the Muhos sandstones pass into impure sandstones and siltstones. Some impure varieties of the Muhos sandstone resemble graywackes, but their red color and high content of Fe_2O_3 are not characteristic of the sandstones of the graywacke suite, which are deposits of the reducing condition of sedimentation. The cement of the quartz sandstones is introduced siliceous matter and a secondary outgrowth of quartz particles is common. The intergranular matrix of the arkoses chiefly contains silty material and a secondary outgrowth of quartz particles is observed only occasionally.

The afore-mentioned mineralogical differences of the sandstones in Finland naturally cause differences in the chemical composition (cf. Tables 3, 5, and 9). The immature arkoses are characterized by a remarkable content of alkalies and Al_2O_3, whereas the quartz sandstones are highly siliceous, containing over 90 per cent of SiO_2. Furthermore, the Muhos sandstones characterized by the abundance of the silty matrix and detrital plagioclase contain more Al_2O_3 and Na_2O than the arkoses of Satakunta.

The mineralogical differences of the Finnish sandstones appear also in the quality of the heavy minerals. The Jotnian arkose sandstone contains many different species of heavy minerals, whereas the quartz sandstones contain only a few species especially resistant to weathering.

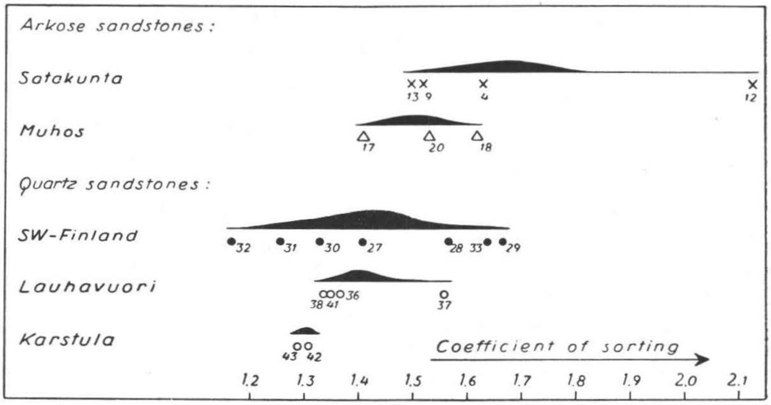


Fig. 15. The coefficient of sorting in the sandstones of Finland. Based on the determinations presented in Figs. 4, 8, 9, and 12.

Textural differences in the sandstones of Finland are found in the degree of roundness and sorting of the particles. The sand grains of the arkoses are mainly subangular or subrounded, whereas those of the quartz sandstones are usually rounded or extremely well rounded. The coefficients of sorting plotted in Fig. 15 show that the quartz sandstones are usually better sorted than the arkoses.

The petrographic characteristics of the sandstones depend on the conditions of origin and deposition controlled by tectonics. The members of the three major classes of sandstones (graywackes, arkoses, and quartz sandstones) represent certain stages in the pulse of the earth characterized by the alternation of orogenic and nonorogenic phases of evolution. The graywackes are typical orogenic sandstones deposited in geosynclinal troughs. Thick accumulations of arkose sandstone are postgeosynclinal deposits usually associated with uplift of the source area after orogenic folding. The pure quartz sandstones are typical for the nonorogenic peneplanation evolutionary phase. The three tectonic stages, controlling the deposition of the sandstones, have been emphasized especially by Krynine (1941).

In the light of the tectonic classification of sandstones we may state that the late pre-Cambrian Jotnian arkose sandstone in Satakunta is postgeosynclinal in relation to the pre-Cambrian, Karelidic orogeny. This has been generally agreed upon by Swedish and Finnish geologists. The importance of the continued vertical tectonic movements, controlling the Jotnian sedimentation, has been stressed by Eckermann (1937 a, b) in his comprehensive studies of the Fennoscandian Jotnian formations. The fault movements connected with the intrusion of the rapakivi granites (latest postkinematic granites of the Karelidic orogeny) have caused rifts and depressions where the Jotnian sediments accumulated. It should be noted that the Jotnian sandstone area in Satakunta forms a downfaulted block in the Archean basement complex.

The petrographic features of the Muhos formation support the conclusion that it belongs to the postgeosynclinal deposits. The basal conglomerates containing pebbles of the Karelidic schists indicate that the Muhos formation is postgeosynclinal in relation to either pre-Cambrian Karelidic or early Paleozoic Caledonidic orogeny. That the Muhos formation belongs to the late pre-Cambrian age is plausible, because no fossils have been found. On the basis of a tectonic classification of sandstones, the Muhos formation is related to the Jotnian formations. The correlation of the Muhos rocks with the Jotnian formation in Satakunta is suggested also by similar petrographic features and a similar manner of occurrence in downfaulted blocks of the crystalline basement. Brenner (1941 and 1944) suggested that the Muhos formation belongs to the Mesozoic age, but no valid evidence was presented. Brenner based his conclusion on the occurrence of a Belemnite fossil said to have been found in the Glacial drift material at Kankari. Later the finder of this Belemnite denied the report (cf. Viluksela, 1951).

The petrographic characteristics of quartz sandstones in Finland deviate strongly from those of the arkosic sandstones and are typical for the sandstones of the peneplanation stage. Similar sandstones of the Cambrian age occur abundantly in the Baltic countries and Sweden and it seems therefore probable that the sporadic occurrences in Finland indicate a Cambrian transgression over a stable peneplained platform. The occurrence of the fossiliferous quartz sandstone in Saltvik affords direct evidence of the Cambrian age.

Acknowledgements — The authors wish to express their thanks to Dr. S. Kaitaro and Dr. N. Edelman for placing the new field material of the sandstone occurrences in the coastal area of southwestern Finland at their disposal. To Mr. V. Okko, M. A., the authors are very grateful for information and helpful discussions concerning the Muhos formation. The X-ray powder diagrams were made by Dr. K. J. Neuvonen and two chemical analyses were carried out by Mr. O. Ojanperä, M. A. Dr. van Houten, Princeton University, critically read the manuscript. The authors wish to express their sincere gratitude to all these persons.

REFERENCES

ASKLUND, B. och KULLING, O. (1926), Nya data till Ålands geologi. Geol. Fören. i Stockholm Förhandl. 48, p. 498.

BRENNER, THORD (1941) Ein ungewöhnliches Kalk-Schlammsteinsediment von Muhos in Mittelfinnland. Geol. Rundschau 32, p. 535.

—»— (1944) Die Bodenbildungen des Muhos-Sediments bei Kieksi. C. R. Soc. géol. Finlande 16; Bull. Comm. géol. Finlande 132, p. 189.

CLARKE, F. W. (1924) Data of geochemistry. U. S. geol. Surv. Bull. 770, p. 34.

VON ECKERMANN, HARRY (1937 a) The Jotnian formation and the sub-Jotnian unconformity. Geol. Fören. i Stockholm Förhandl. 59, p. 19.

—»— (1937 b) The genesis of the Jotnian sediments. Geol. Fören. i Stockholm Förhandl. 59, p. 548.

EDELMAN NILS (1949) Sandsten. Geologi 2, p. 7.

—»— (1951) Glacial abrasion and ice movement in the area of Rosala-Nötö, S. W. Finland. C. R. Soc. géol. Finlande 24; Bull. Comm. géol. Finlande 154, p. 157.

ENKOVAARA, A., HÄRME, M. ja VÄYRYNEN, H. (1953) Suomen Geologinen Yleiskartta. Lehdet C5—B5. Oulu—Tornio. Kivilajikartan selitys.

ESKOLA, PENTTI (1913) On phenomena of solution of Finnish limestones and on sandstone filling cavities. Bull. Comm. géol. Finlande 36.

GYLLING, K. H. N. (1887) Zur Geologie der Cambrischen Arkosenablagerung des westlichen Finnlands. Zeitschr. d. deutsch. geol. Ges. 39, p. 770.

HADDING, ASSAR (1929) The pre-Quaternary sedimentary rocks of Sweden. III. The Paleozoic and Mesozoic sandstones of Sweden. Medd. Lunds Geol. Min. Inst. 41.

HAUSEN, H. (1911) Stenräkningar på Åland. Geol. Fören. i Stockholm Förhandl. 33, p. 495.

—»— (1934) Über ein neunentdecktes Kaolin-Sandstein-Vorkommen im kristallinen Kalkstein auf Pargas—Ålön, Gegend von Åbo—Turku, SW. Finnland. Acta Acad. Aboensis Mathem. et Phys. 8, p. l.

—»— (1942) Der Abschnitt Finlands der kaledonischen Überschiebungszone. Soc. Scient. Fennica. Comm. Phys.—Math. 11.

HELLAAKOSKI, AARO (1930) On the transportation of materials in the esker of Laitila. Fennia 52, No. 7.

KAHMA, AARNO (1951) On contact phenomena of the Satakunta diabase. Bull. Comm. géol. Finlande 152.

KAITARO, SIMO (1949) Lisää kambrisia hiekkakivijuonia. Geologi 3—4, p. 13.

—»— (1953) Geologic structure of the late pre-Cambrian intrusives in the Åva area, Åland Islands. Bull. Comm. géol. Finlande 162.

KRYNINE, PAUL. D. (1941) Differentiation of sediments during the life history of a landmass. Bull. Geol. Soc. Am. 52, p. 1915.

LAITAKARI, AARNE (1925) Über das Jotnische Gebiet von Satakunta. Bull. Comm. géol. Finlande 73.

—»— (1932) Suomen kivien raskaista mineraaleista. Tekn. Aikakauslehti 9, p. 408.

LOKKA, LAURI (1950) Chemical analyses of Finnish rocks. Bull. Comm. géol. Finlande 151.

MACKIE, W. (1901) Seventy chemical analyses of rocks. Trans. Edin. Geol. Soc. 8, p. 33.

MARMO, VLADI ja LAITAKARI, AARNE (1952) Lounais-Suomen rikkikiisukonkretioista. Geoteknillisiä julkaisuja 53.

METZGER, A. TH. (1922) Beiträge zur Paläontologie des nordbaltischen Silurs im Ålandgebiet. Bull. Comm. géol. Finlande 56.

—»— (1927) Zur Kenntnis des nordbaltischen Kambro-Silurs auf Åland und im südwestlichen Küstengebiet Finnlands. Fennia 47, No. 12.

OLANDER, A. J. (1934) Lauhavuoren vanhoista rantamuodostuksista. Terra, p. 185.

PETTIJOHN, F. J. (1949) Sedimentary rocks. New York.

SAURAMO, MATTI (1916) Über das Vorkommen von Sandstein in Karstula, Finland. Fennia 39, No. 7.

SEDERHOLM, J. J. (1897) Über eine archäische Sedimentformation in südwestlichen Finland und ihre Bedeutung für die Erklärung der Entstehungsweise der Grundgebirges. Bull. Comm. géol. Finlande 6.

—»— (1913 a) Suomen Geologinen Yhteiskartta. Lehti B2. Tampere, Kivilajikartan selitys.

—»— (1913 b) Weitere Mitteilungen über Bruchspalten mit besonderer Beziehung zur Geomorphologie von Fennoskandia. Bull. Comm. géol. Finlande 37.

—»— (1924) Granit-gneisproblemen belysta genom iakttagelser i Åbo—Ålands skärgård. Geol. Fören. i Stockholm Förhandl. 46, p. 129.

—»— (1934) On migmatites and associated pre-Cambrian rocks of southwestern Finland. Part III. The Åland Islands. Bull. Comm. géol. Finlande 107.

TANNER, V. (1911) Über eine Gangformation von fossilführendem Sandstein auf der Halbinsel Långbergsöda—Öjen im Kichspiel Saltvik, Åland—Inseln. Bull. Comm. géol. Finlande 25.

TOMLINSON, C. W. (1916) The origin of red beds. J. Geol. 24, p. 153.

TWENHOFEL, W. H. (1950) Principles of sedimentation. New York.

VILUKSELA, ERKKI (1951) Der Belemnitenfund bei Kankari. C. R. Soc. géol. Finlande 24; Bull. Comm. géol. Finlande 154, p. 241.

ÖDMAN, OLOF H., HÄRME, M., MIKKOLA, A. och SIMONEN, AHTI (1949) Den svensk-finska geologiska exkursionen i Tornedalen sommaren 1948. Geol. Fören. i Stockholm Förhandl. 71, p. 113.

EXPLANATION OF PLATES

PLATE I

Fig. 1. Jotnian arkosic sandstone of Satakunta. Eura, Kiperjärvenoja. 10 x.
Fig. 2. Jotnian arkosic sandstone of Satakunta. Luvia, Lappäng. 10 x.
Fig. 3. Muhos sandstone. Diamond drill hole Liminka, depth 217.5 m. 10 x.
Fig. 4. Muhos sandstone. Diamond drill hole Muhos, depth 347.8 m. 10 x.
Fig. 5. Quartz sandstone of Lauhavuori. Riitakangas. 10 x.
Fig. 6. Quartz sandstone of Lauhavuori. Tiiliharju. 10 x. Photo: Halme.

PLATE II

Fig. 7. Ripple marks in the Jotnian sandstone. Orientation of the ripples in the successive beds is vertical to each other. Glacial boulder collected by Dr. Edelman. Nauvo. Photo: Halme.

Fig. 8. Rain impressions in the ripple-marked Jotnian sandstone. Glacial boulder collected by Dr. Edelman. Nauvo. Photo: Halme.

PLATE III

Fif. 9. Exposure of the Lauhavuori sandstone. Tiiliharju. Photo: Härme.
Fig. 10. Boulder field consisting exclusively of the Lauhavuori sandstone boulders. Kovapruuki. Photo: Härme.

PLATE IV

Fig. 11. Clay balls in the Lauhavuori sandstone. Riitakangas. ½ nat. size. Photo: Halme.
Fig. 12. Stratification in the Lauhavuori sandstone. ½ nat. size. Photo: Halme.
Fig. 13. Angular fragments of strongly altered granite in the Lauhavuori sandstone. Vähä-Lauha. ⅓ nat. size. Photo: Halme.

PLATE I.

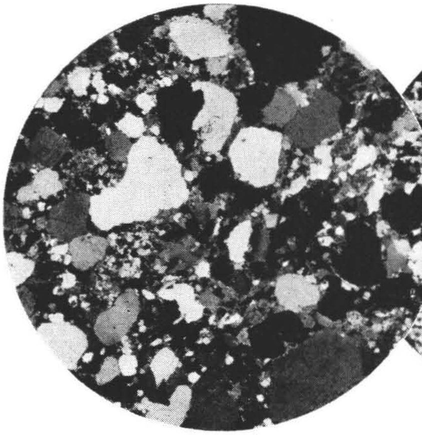


Fig. 1

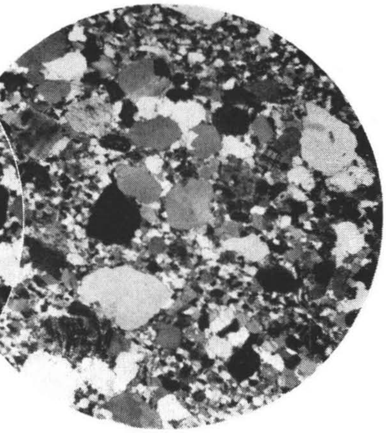


Fig. 2

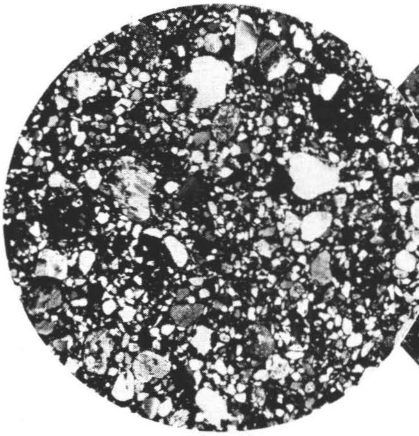


Fig. 3

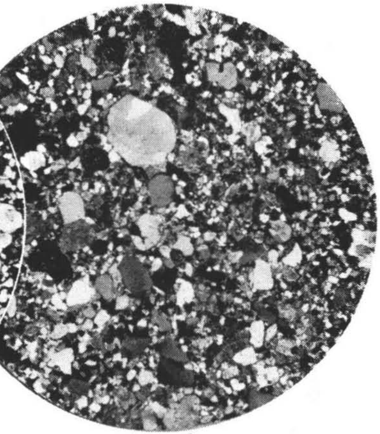


Fig. 4

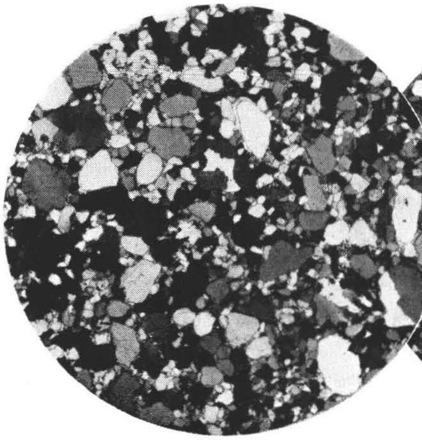


Fig. 5

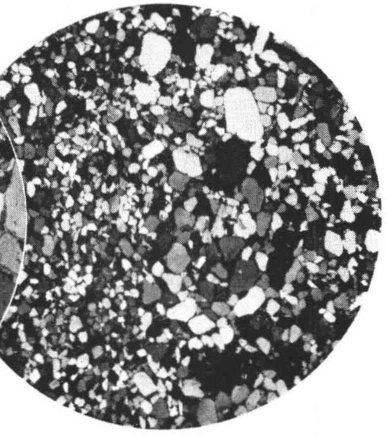


Fig. 6

Ahti Simonen and *Olavi Kouvo:* Sandstones in Finland.

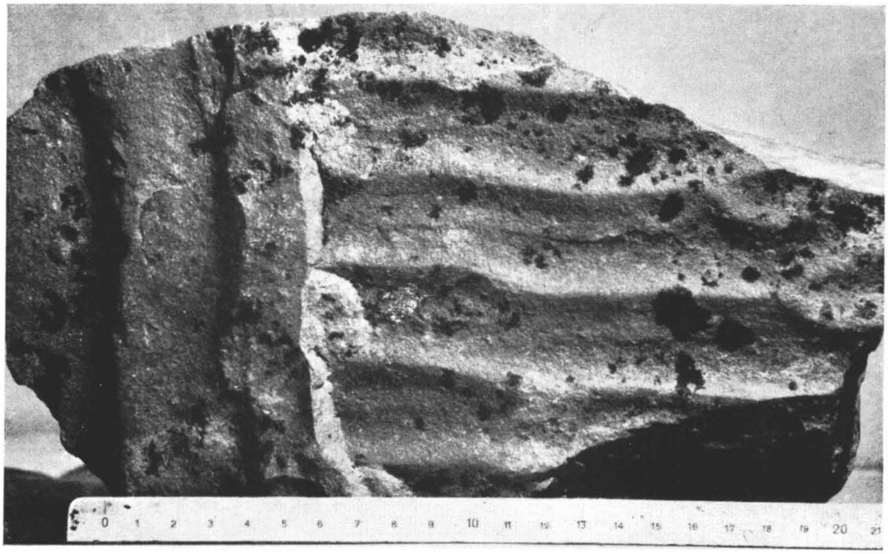


Fig. 7

Fig. 8

Ahti Simonen and *Olavi Kouvo:* Sandstones in Finland.

Fig. 9

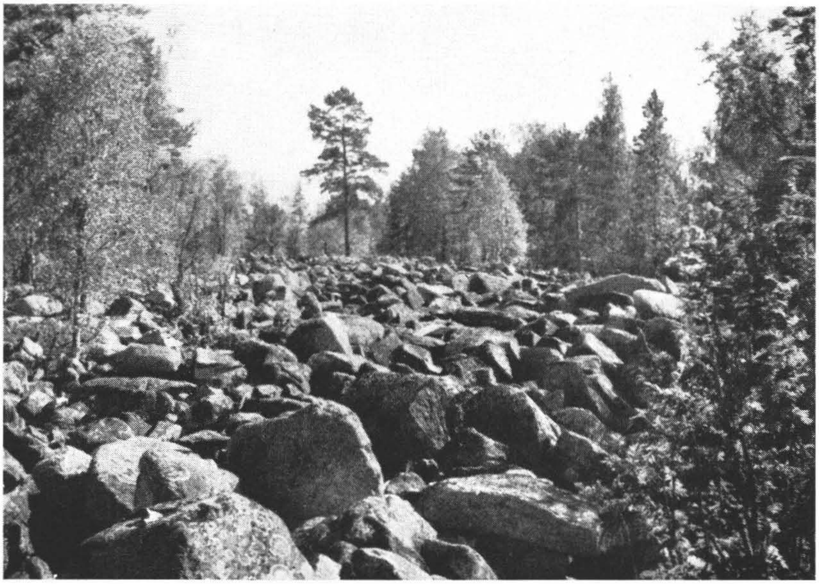


Fig. 10

Ahti Simonen and *Olavi Kouvo:* Sandstones in Finland.

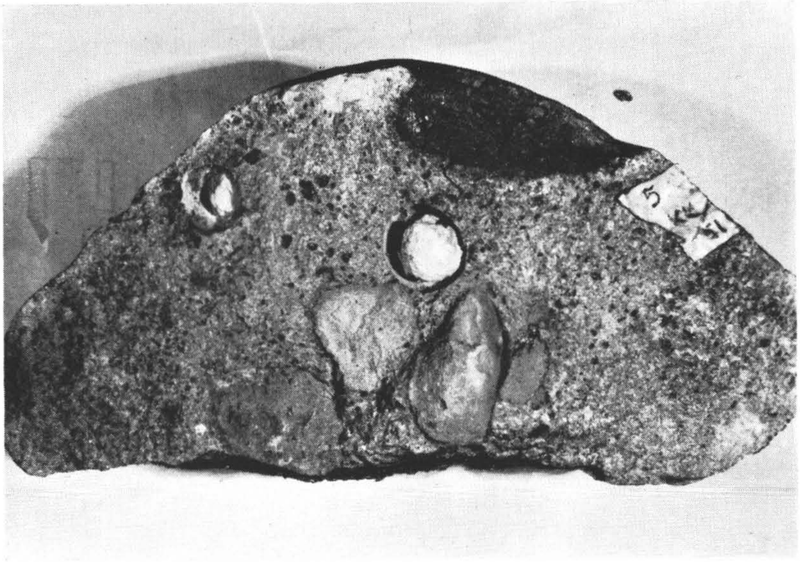


Fig. 11

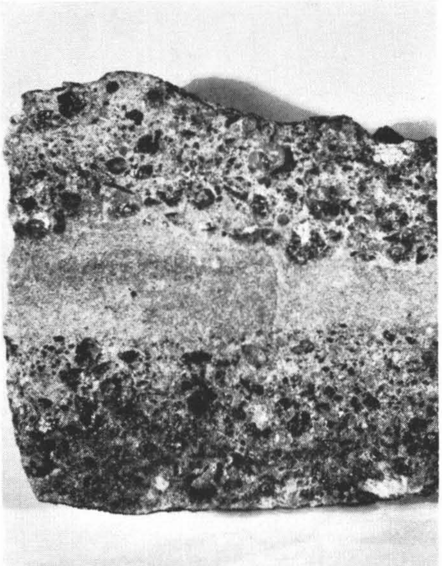


Fig.12

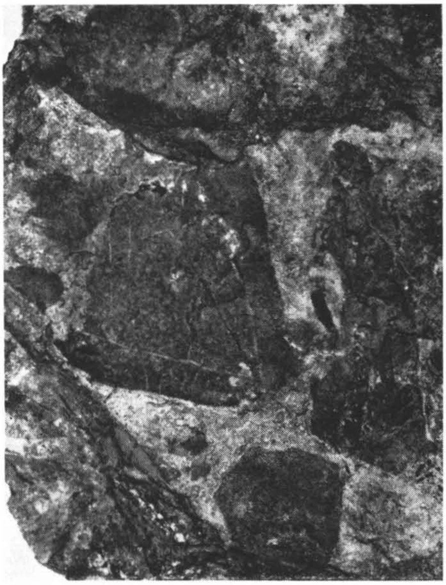


Fig. 13

Ahti Simonen and *Olavi Kouvo:* Sandstones in Finland.

8

ON THE GRAPHIC INTERGROWTHS OF SILICATIC MINERALS WITH ORE OXIDES [1]

BY

OKE VAASJOKI

Geological Survey of Finland, Helsinki

ABSTRACT

In subsilisic rocks accompanied by titaniferous iron ores, graphic intergrowths of ore oxides with silicatic constituents have been noted in some cases, although they have been mostly ignored, especially in studies of a petrographic character. In the following these textures are described from a couple of Finnish titaniferous occurrences while attempting to show the significance of these textures in outlining the order and mechanism of the deposition.

CONTENTS

INTRODUCTION

In general there are few authors only who have paid attention to graphic intergrowths of ore minerals with silicates, although such textures must be rather common in subsilisic rocks. One beautiful example is given by Vogt (1921) of intergrowths of ilmenite and magnetite with olivine in specimens from Fiskö, Söndmöre, Norway. Schwartz (1930) in his study on the Duluth gabbro accompanied by titaniferous iron ores offers an illuminating example of graphic intergrowths of ilmenite with hornblende. Ödman (1932) has established the graphic intergrowths of magnetite with pyroxene in the lavas of Mt. Elgon and reproduced a picture showing

---

[1] Received March 11, 1955.

magnetite occurring within the corona surrounding a grain of olivine. In that corona magnetite has a worm-like appearence, relating to a graphic intergrowth against the corona-forming mineral. Further, in his publication on ore minerals in common igneous rocks, Newhouse (1936) presents two excellent pictures of implication textures of magnetite-ilmenite or magnetite with ferromagnesian silicates and reports, that those textures in which opaque oxides form a component were observed most often in the subsilisic rocks, such as gabbros and diabases, less often in the dioritic rocks, and not at all in granite and syenite. The specimens in the first instance presented by Newhouse were taken from Duluth gabbro and that in the second from an andesite at Kyoshe, Japan. Later reprinting the aforementioned microphoto taken of the Duluth gabbro, Ramdohr (1940) states that in general the intergrowths of magnetite-ilmenite with silicates belong to the »synantetic rims» in the sense of Sederholm (1916). In the study at hand the corresponding textures have been considered especially with an eye to the paragenetic properties of the rocks investigated.

LOCATIONS AND MINERALOGICAL OUTLINE

The specimens for the present study have been collected from the localities of Susimäki and Riuttamaa in southern Finland about 150 km west-northwest from Helsinki and at a distance of about 4 km from each other. Both of these occurrences, which as yet have no economic importance, are situated within a rather uniform zone of hornblende gabbros and diorites. A microscopical and chemical study of the rocks and associated titaniferous iron ores in the Susimäki area has been carried out in detail by Palmunen (1925). In the same connection he has also given a tentative description of the Riuttamaa ore from an ore-microscopical point of view.

Palmunen has considered that the rocks in the Susimäki area are typical representatives of rocks produced by magmatic differentiation. Accordingly, in successive series of the rocks, starting from the most mafic member, he has distinguished four distinct types of rock in the area as follows:

— Ilmenite-magnetite olivinite, i.e. corresponding to the richest ore of the area.

— Peridotite
— Hornblende-ilmenite gabbro
— Hornblende-biotite gabbro

The specimens from Susimäki and Riuttamaa, which have been of special value for this study, belong to the first and the second group of the rocks as listed above.

The rock called ilmenite-magnetite olivinite is composed almost entirely of ore oxides and olivine. The proportion between the opaque material and silicates in the best ore of Susimäki is about 2 : 3, while the best ore of Riuttamaa is poorer, the host rock being close to gabbro in composition. As a usual minor component in both cases there occur small grains of hercynite. The olivine is slightly serpentinized along the cracks and according to Palmunen its composition in Susimäki corresponds to the formula $Fa_{45}Fo_{55}(2V_{\alpha} = 80° \pm 5°)$.

The content of olivine and ore material decrease while those of hypersthene ($2V_{\alpha} = 70°$) and hornblende ($2V_{\alpha} = 2V_{\gamma} = 90° \pm 2°$; $c \Lambda \gamma = 14°$) increase in the rock called peridotite by Palmunen. Anorthite-rich plagioclase (An_{65-75}) is noted as one of the usual components sometimes occurring as large, weakly zoned, corroded crystals. Besides the ore minerals, brown biotite, apatite and hercynite occur as accessories. On the basis of the microscopical properties, the rock in question must be called gabbro rather than peridotite.

The contents of ilmenite and magnetite are rather evenly distributed in the specimens from both the Susimäki and Riuttamaa occurrences. Small grains of pyrite, pyrrhotite and pentlandite are occasionally met with among the oxidic ore minerals.

Similarly in both occurrences the ilmenite does not contain hematite as a product of exsolution so typical of many ilmenites of titaniferous iron ores. This possibly means weaker oxidizing factors during the time of crystallization (Newhouse, op.cit.). In the magnetite, however, there occur an abundance of small disks of spinel and very minute grains of some semi-opaque mineral not possible to analyze. The ilmenite forms very narrow, lath-shaped exsolution products only occasionally in the magnetite.

IMPLICATION TEXTURES

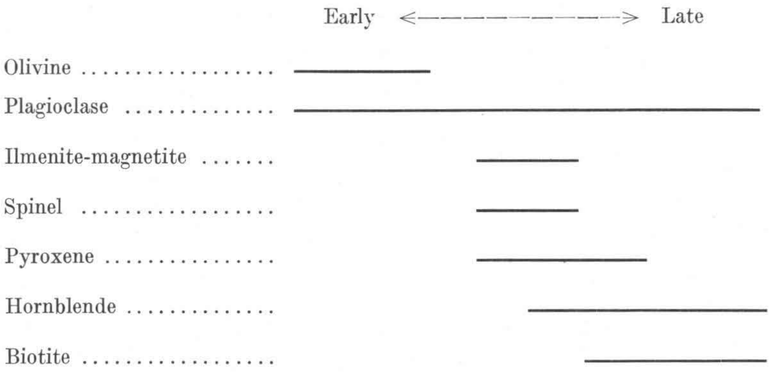
In his highly esteemed publication on synantetic minerals Sederholm has made a contribution to an extensive literature, presenting many observations of his own on reaction rim textures met with around the olivine grains of gabbroic rocks in southern Finland. Such a phenomenon is a typical one in the gabbros now under consideration from the Susimäki and Riuttamaa occurrences (the peridotite in the nomenclature used by Palmunen). One example of these formations, generally called coronas in the literature, is shown in Fig. 1 of Plate I. Usually the innermost part is build up of olivine, the next zone of colorless, enstatitic pyroxene, followed by a mixture of hypersthenic pyroxene and needle-shaped hornblende. In some cases the colorless zone is absent and the nucleus made up of the olivine is then surrounded by a corona composed of hypersthene and hornblende. It should be especially pointed out that sometimes the large, scattered crystals of plagioclase are likewise surrounded by coronas composed of green, needle-shaped hornblende.

The interesting textural relations of the ore oxides with silicatic constituents, which form the main topic of this study, appear most conspicuously in gabbro with a relatively poor content of ore. Some unexceptional textural behavior was observed already by Palmunen (op.cit., p. 13), who stated: »At the boundaries of the larger crystals of olivine a myrmekitelike intergrowth of olivine and iron ore is frequently seen. In it the diverging, fingershaped worms of ore, as well as the convex sides of the whole growth, point towards the olivine». This statement does not make clear the fact that the myrmekitelike formations, as seen in Fig. 2 of Plate I, are always associated with some other mineral outside the olivine and have never been found in the olivine itself. By means of polished thin sections this circumstance in especial is easily confirmed (Pl. II. Fig. 3) Even in case the myrmekitelike texture is observed inside a grain of olivine, it can be established that the host for the intergrowth is a mineral, which has replaced the olivine, e.g. in an intersection of grain cracks.

PARAGENESIS

The order of crystallization of ore oxides and silicatic constituents in subsilisic rocks accompanied by titaniferous iron ores show essential differences as regards different occurrences. That is obviously true also regarding the occurrences referred to here as examples showing graphic intergrowths. According to Schwartz, in the ore associated with Duluth gabbro, a part of the ilmenite crystallized with the silicates presumably later than the greater part of the magnetite and ilmenite. A similar tendency is present also in the occurrence of titaniferous iron ore at Otanmäki, Finland

Table 1



(Vaasjoki, 1947). In the lavas of Mt. Elgon Ödman distinguished three different generations of magnetite. He concluded, however, that these are not products of three different epochs, but merely indicate the maxima during the course of the separation of magnetite. Newhouse (op.cit.) insisted that the magnetite-ilmenite, in part, finished crystallizing late in many gabbros formed by the process of fractional crystallization differentiation. He subsequently accepted and developed further the idea of filterpress (Osborne, 1928) — a mechanism by which a still fluid, interstitial ore-material was pressed out from an earlier crystallized silicate mass and was intruded into a horizon of less pressure to crystallize. This view, however, has not been fully approved, and especially Ramdohr (1940) has made some sceptical comments as regards the aforementioned interpretation.

In the ore-bearing rocks of Susimäki, as well as in those from Riuttamaa, the crystallization of the olivine, and probably of a part of the anorthite rich plagioclase, took place prior to the crystallization of the ore oxides. This is suggested in the first place by the invariable occurrence of graphic intergrowths outside the grains of olivine, and still better when the ore-material brecciates the grains of olivine, separately embedded in the mass of ore oxides (Pl. II, Fig. 4) in specimens of the ilmenite-magnetite olivinite. Starting from these observations, the total paragenetic succession of the minerals in the present cases is outlined in Table 1.

CONCLUDING REMARKS

The most wide-spread concept in the literature concerning the genesis of titaniferous iron ore deposits suggests that these deposits have been segregated by early crystallization of the oxides within the course of the differentiation of the subsilisic magmas. It has been pointed out (Schneiderhöhn, 1940), however, that in many instances the original character might be confused e.g. by later autohydratic processes. Considering those formations in which the ore-material has the appearence of younger injections, the opinion has been advanced that the magmatic ore oxides originated partly by a liquid immiscibility. The total process has thus given rise to the formation of independent oxidic ore magmas (e.g. Schneiderhöhn, 1940). In summary, all the exceptions to the first-mentioned concept have advanced the idea that the final concentration and crystallization of magmatic ore oxides in very many instances took place at a relatively late stage of magmatic development.

In the present study it has been shown how the ore-material has brecciated and corroded the olivine crystals. This apparently emphasizes a relatively long period between the crystallization of the olivine and that of the ore fluid. On the other hand, the formation of the coronas and the graphic intergrowths, as described in the foregoing, are to be held as related phenomena, which presumably were fixed by simultaneous crystallization during a period when the temperature apparently favored the crystallization of pyroxenes as well. Both the aforementioned features obviously suggest a late magmatic development of the Susimäki and Riuttamaa ores along the general lines recently presented by Bateman (1951) and others.

REFERENCES

BATEMAN, ALAN M. (1951) The formation of late magmatic oxide ores. Econ. Geol. 48, p. 404.

NEWHOUSE, W. H. (1936) Opaque oxides and sulphides in common igneous rocks. Bull. Geol. Soc. Amer. 47, p. 1.

OSBORNE, F. F. (1928) Certain titaniferous magnetite deposits and their origin. Econ. Geol. 23, p. 724 and p. 859.

PALMUNEN, M. K. (1925) On the ilmenite-magnetite olivinite of Susimäki, Fennia 45, No. 9.

RAMDOHR, P. (1940) Die Erzmineralien in gewöhnlichen magmatischen Gesteinen. Abh. preuss. Akad. Wiss. Math.-nat. Kl., No. 2.

SCHNEIDERHÖHN, HANS (1940) Lehrbuch der Erzlagerstättenkunde. I Band. Jena.

SCHWARTZ., G. M. (1930) The relations of magnetite and ilmenite in the magnetite deposits of the Duluth gabbro. Am. Min. 15, p. 243.

SEDERHOLM, J. J. (1916) On synantetic minerals and related phenomena. Bull. Comm. géol. Finlande 48.

VAASJOKI, O. (1947) On the microstructure of titaniferous iron ore at Otanmäki. Bull. Comm. géol. Finlande 140, p. 107.

VOGT, J. H. L. (1921) The physical chemistry of the crystallization and magmatic differentiation of igneous rocks. J. Geol. 29, p. 627.

ÖDMAN, OLOF H. (1932) Minerographic study on the opaque minerals in the lavas from Mt. Elgon, British East Africa. Geol. Fören. i Stockholm Förh. 54, p. 285.

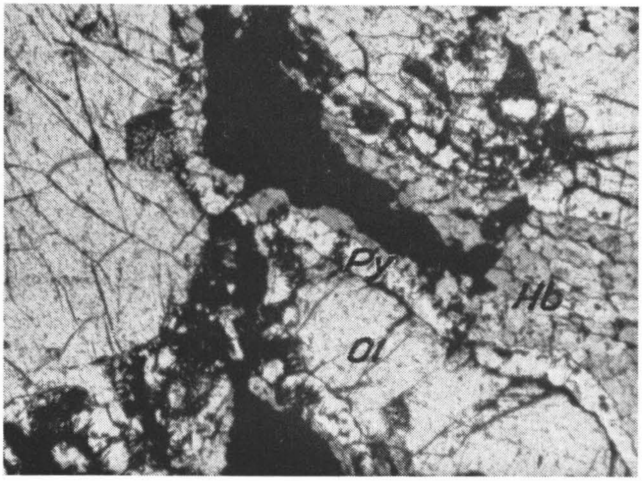


Fig. 1. Gabbro, Riuttamaa. Reaction rims including groups of myrmekitelike ore are visible around the crystals of olivine. Ol = olivine, Py = pyroxene, Hb = hornblende. Thin section. 40×. Photo E. Halme.

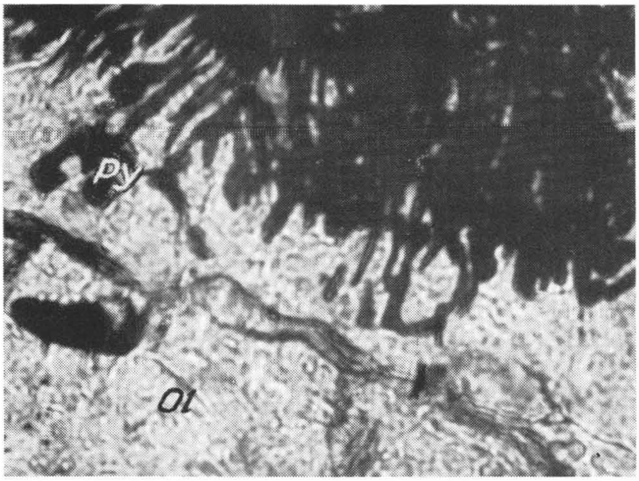


Fig. 2. Gabbro, Riuttamaa. Myrmekitelike ore in pyroxene. Ol = olivine, Py = pyroxene. Thin section. 100 ×. Photo E. Halme.

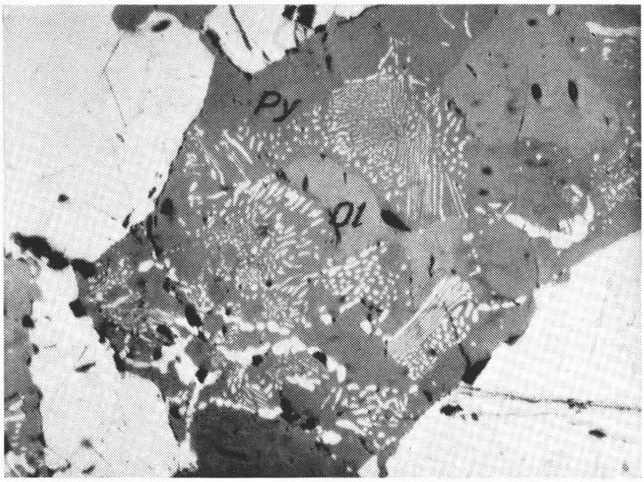


Fig. 3. Graphic texture of magnetite with pyroxene, Riuttamaa. Ol = olivine, Py = pyroxene, white = ilmenite-magnetite. Polished section. 100 ×. Photo E. Halme.

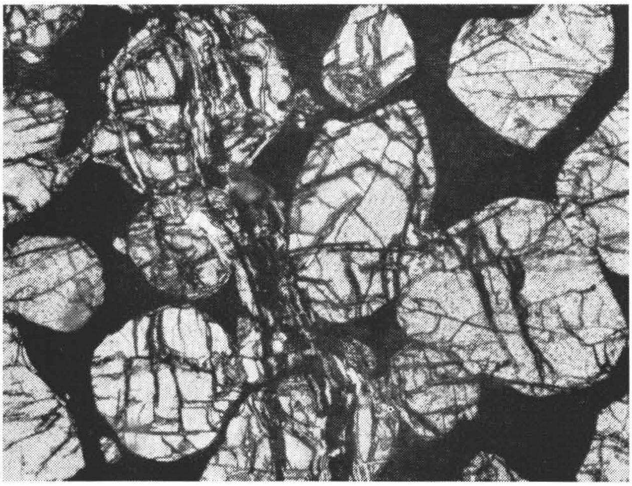


Fig. 4. Ilmenite-magnetite olivinite, Susimäki. The crystals of olivine partly brecciated by the ore material. Thin section. 20 ×. Photo E. Halme.

Oke Vaasjoki: On the graphic intergrowths of silicatic minerals with ore oxides

AN EXAMPLE OF GRANITIZATION[1]

BY

MAUNU HÄRME and MATTI LAITALA

Geological Survey of Finland, Helsinki

ABSTRACT

This paper deals with the granitization of gabbro fragments in late-kinematic microcline granite. The chemical and mineralogical changes caused by the metasomatic addition of potassium and removal of calcium are described.

During the recent years the authors have done geological mapping in the coastal area of southern Finland. In 1952 the mapping of the archipelago area south of Snappertuna parish was carried out by the junior author (M.L.) under the supervision of the senior author. The Archean basement complex of this area belongs to the Svecofennian formations, and the migmatites described by Sederholm in his numerous papers are common here. Recently Laitala (1954) carried out his pro gradu investigation dealing with the general geology of the Snappertuna archipelago.

In connection with the field work attention was also focussed on granitization, which is a common phenomenon in this area. An interesting example of this process was found at Karlholm, a little island south of Stakanäs in Snappertuna. The rock complex of this island consists of migmatized gneisses, breccias, etc. The migmatite contains some subangular gabbro fragments, which are surrounded by darker reaction rims (Plate I, Fig. 1). The middle part of such a fragment is gabbro and the darker rim consists of garnet-bearing biotite-plagioclase-quartz rock.

On the basis of the field observations as well as of its mineralogical and chemical composition, it seems evident that the whole fragment had originally been one coherent and homogeneous gabbro fragment, a block in the breccia. Intrusive microcline granite surrounding the fragment subsequently caused metasomatic processes in the outer parts of it. The size of such fragments is about 30 × 50 cm and the darker rim is about 2—5 cm broad. In many fragments it is seen that the dark rim produced by metasomatism is equally thick on each side of the fragment. This shows that the metasomatic processes have simultaneously influenced the fragment from all sides.

---

[1] Received March 12, 1955.

Specimens from the central part of the gabbro fragment (1), from the boundary line between the central part and the darker outer zone (2) and from the dark marginal zone (3) were selected for more detailed mineralogical and chemical study. The points where the specimens studied were taken are marked in Plate I, Fig. 1. The mineralogical composition of the different zones of the fragment are presented in Table 1.

Table 1. Mineralogical composition of the different zones of the fragment, as determined by means of the point counter method

Constituent	1	2	3
Plagioclase	48.9	45.7	34.6
Amphibole	32.3	—	—
Chlorite (alteration product of the amphibole)	0.2	33.2	—
Biotite	7.4	8.5	40.9
Quartz	9.6	10.7	19.1
Garnet	—	—	3.6
Accessories	1.6	1.9	1.8
Total	100.0	100.0	100.0

The gabbro in the central part of the fragment (1 in Table 1) is medium-grained and only weakly orientated (Plate I, Fig. 2). The amphibole is colorless and its optical properties are as follows (Laitala, 1954): $\alpha = 1.640$, $\gamma = 1.667$; $2V\gamma = 82°$, $c \wedge \gamma = 17°$. The amphibole is strongly twinned, the twinning plane being (100). In part the amphibole has altered into biotite. Some amphibole grains still show the crystal form of pyroxene, indicating that they are products of uralitization. The composition of the plagioclase is about An_{80}. The texture is granoblastic, although some blastohypidiomorphic features occur.

Along the boundary line between the central part and the darker outer zone (2 in Table 1), the rock is very different. The amphibole is strongly altered into chlorite and biotite. The composition of the plagioclase is no longer determinable.

In the dark outer rim of the fragment (3 in Table 1) the rock is foliated and recrystallized (Plate I, Fig. 3). The texture is granoblastic. The composition of plagioclase is about An_{60}. Reddish garnet (almandite, $N = 1.798$; Laitala, 1954) occurs irregularly as small grains but in places it forms big porphyroblasts, as is also observable in Plate I, Fig. 1.

Apatite, carbonate and ore occur as accessories in all the specimens studied.

The chemical composition of the afore-described specimens is presented in Table 2a in weight percentages and in Table 2b in one-cation oxide percentages (= cation percentages; Eskola, 1954). The weight norms, Niggli numbers and one-cation molecular norms are calculated. Oxygen is calculated from the one-cation oxide percentages.

The chemical composition of the central part is nearly primary, only the free quartz and the small amount of biotite show that the middle part is not wholly unchanged. Commonly the gabbros in the neighborhood do not contain free quartz and their potassium content is very low.

Table 2a. Chemical analyses of a fragment. Karlholm, Snappertuna. Analyst, Pentti Ojanperä

	1		2		3		Niggli numbers				Weight norm			
	Weight %	Mol. prop.	Weight %	Mol. prop.	Weight %	Mol. prop.		1	2	3		1	2	3
SiO_2	49.24	8194	47.21	7857	50.69	8436	si	131.7	136.1	146.3	qu	9.99	9.09	10.54
TiO_2	2.42	303	2.67	334	2.62	328	ti	4.9	5.8	5.7	or	9.91	16.98	24.82
Al_2O_3	16.38	1607	16.64	1632	15.77	1547	al	25.8	28.3	26.8	ab	4.72	3.62	3.15
Fe_2O_3	1.23	77	1.32	83	1.41	88	fm	49.1	46.3	52.5	an	33.41	27.95	16.65
FeO	10.85	1510	8.96	1247	10.85	1510	c	20.8	18.9	11.9	en	13.72	12.35	13.40
MnO	0.18	25	0.22	31	0.15	21	alk	4.3	6.5	8.8	fs	15.24	11.36	14.71
MgO	5.51	1367	4.96	1230	5.38	1334	qz	14.4	10.2	11.1	cor	1.40	2.58	4.51
CaO	7.24	1291	6.12	1091	3.84	685	k	0.66	0.81	0.88	mt	1.79	1.92	2.04
Na_2O	0.56	90	0.43	69	0.37	60	mg	0.45	0.46	0.44	il	4.60	5.07	4.98
K_2O	1.68	178	2.87	305	4.20	446	o	0.05	0.06	0.06	ap	0.89	0.85	0.85
P_2O_5	0.38	27	0.37	26	0.37	26	c/fm	0.42	0.41	0.23		95.67	91.77	95.65
CO_2	1.06	241	2.81	639	0.87	194					CO_2	1.06	2.81	0.87
$H_2O +$	2.59	1438	4.55	2526	2.86	1587		—	—	—	$H_2O +$	2.59	4.55	2.86
$H_2O —$	0.49	272	1.08	600	0.40	217		—	—	—	$H_2O —$	0.49	1.08	0.40
Total	99.81	—	100.21	—	99.78	—		—	—	—		99.81	100.21	99.78

1. Gabbro from the center of the fragment.
2. Rock from the boundry line between the central part and the darker outer zone.
3. Garnet-bearing biotite-plagioclase-quartz rock (the dark rim around the fragment).

Table 2b

	1			2			3			One-cation molecular norm			
	Cation number	Cation %	O in oxides	Cation number	Cation %	O in oxides	Cation number	Cation %	O in oxides		1	2	3
SiO_2	8194	48.52	97.04	7857	47.17	94.34	8436	50.09	100.18	Q	9.825	9.115	11.465
TiO_2	303	1.79	3.58	334	2.01	4.02	328	1.95	3.90	Or	10.55	18.25	26.50
$AlO_{3/2}$	3213	19.02	28.53	3264	19.60	29.40	3093	18.36	27.54	Ab	5.35	4.15	3.55
$FeO_{3/2}$	154	0.91	1.37	165	0.99	1.49	177	1.05	1.58	An	35.525	30.175	17.775
FeO	1510	8.94	8.94	1247	7.48	7.48	1510	8.97	8.97	En	16.20	14.76	15.84
MnO	25	0.15	0.15	31	0.19	0.19	21	0.12	0.12	Fs	13.69	10.33	11.13
MgO	1367	8.10	8.10	1230	7.38	7.38	1334	7.92	7.92	Cor	1.63	3.05	5.24
CaO	1291	7.64	7.64	1091	6.55	6.55	685	4.07	4.07	Mt	1.365	1.485	2.625
$NaO_{1/2}$	181	1.07	0.54	139	0.83	0.42	119	0.71	0.36	Il	3.58	4.02	3.90
$KO_{1/2}$	357	2.11	1.06	609	3.65	1.83	892	5.30	2.65	Ap	0.855	0.825	0.825
$PO_{5/2}$	54	0.32	0.80	52	0.31	0.78	52	0.31	0.78		98.57	96.16	98.85
CO_2	241	1.43	2.86	639	3.84	7.68	194	1.15	2.30	CO_2	1.43	3.84	1.15
H_2O+	(1438)	(8.51)		(2526)	(15.16)		(1587)	(9.42)			100.00	100.00	100.00
$H_2O—$	(272)	(1.61)		(600)	(3.60)		(217)	(1.92)					
Total	16.890	100.00	160.61	16.658	100.00	161.56	16.841	100.00	160.37				
	—O for (OH)		8.51	—O for (OH)		15.16	—O for (OH)		9.42				
			152.10			146.40			150.95				
	(OH)		17.02	(OH)		30.32	(OH)		18.84				
Anions pro 100.00 cations			169.12			176.72			169.79				

Considering the series of analyses, it is interesting to note that the amount of calcium and to some degree also that of sodium decrease outwards from the center of the fragment, whereas the amount of potassium increases outwards. The content of ferric iron likewise increases slightly towards the margin. Other constituents occur in nearly the same amounts in the central part and in the marginal zone of the fragment, or at least the differences are not appreciable. The ocmparatively small changes in the chemical composition between the center and the marginal zone indicate, however, the possibility of greater differences in mineralogical composition. In the marginal zone magnesium and ferrous iron are included in the biotite and garnet. The aluminum content has not increased during the metasomatism; the aluminum has only become relatively »enriched» during the metasomatic reduction of the calcium and sodium content. Hence, the occurrence of garnet does not, at least here, indicate any primary »excess» of aluminum. The weight norm and one-cation molecular norm of Analysis 1 show in the values of cor and Cor a slight excess of aluminum, but it is to be noted that the middle part of the fragment is not wholly unchanged either.

Free quartz increases in the fragment towards the margin and the value of si in Niggli numbers reveals the same phenomenon. The values of cor in weight norms and those of Cor in one-cation molecular norms reveal the relative »enrichment» of aluminum in this metasomatic process. The values of mt and Mt show the slight growth of ferric iron towards the margin of the fragment. Obviously a weak oxidation of ferrous iron has taken place in the metasomatic process. The norms and Niggli numbers plainly reveal the changes in the amounts of alkalies and calcium.

Analysis 2 is made from the »front line» of the advanced metasomatism. It is interesting to note its high content of CO_2, which is for the greater part associated with calcium to form calcite. In this analysis the amount of H_2O is comparatively high, too. So it is obvious that volatile constituents were involved in the process. The present authors deem it evident that the original rock was gabbro and no facts indicate that any »basic front» is here in question.

On the whole, the metasomatic process described in the foregoing is an example of the initial phase of granitization. Here in the marginal zone of the fragment the plagioclase is still comparatively rich in anorthite, whereas in the far-advanced process the plagioclase becomes albitic. Here the potassium has been used to produce biotite but it is not yet sufficient in amount to form potassium feldspar.

Processes similar to those described in the foregoing also indicate the possibilities of cordierite production. The formation of garnet during metasomatism is not the only example. The quartzdiorites and granodiorites in southwestern Finland contain almandite when they occur in the granitization zones, i.e. in the zones where they have been affected by the metasomatic granitization process caused by the late-kinematic microcline granite. The senior author (M.H.) will deal with these processes in another paper.

REFERENCES

Eskola, Pentti (1954) A proposal for the presentation of rock analyses in ionic percentage. Ann. Acad. Sci. Fennicae, Ser. A, III, 38.

Laitala, Matti (1954) Havaintoja Snappertunan saaristoalueen kallioperästä. Manuscript in the Archives of the Institute of Geology, University of Helsinki.

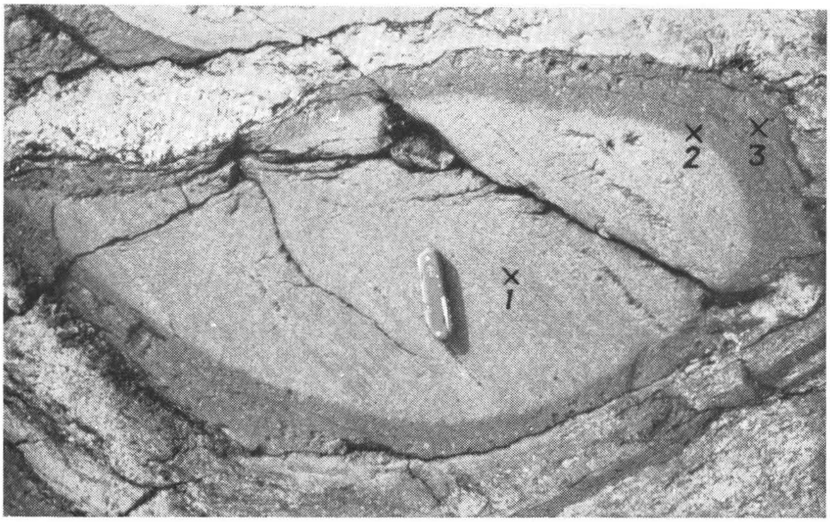


Fig. 1. A fragment of gabbro in migmatite. The fragment is surrounded by a darker garnet-bearing reaction rim. Karlholm, Snappertuna. 1/15 natural size.

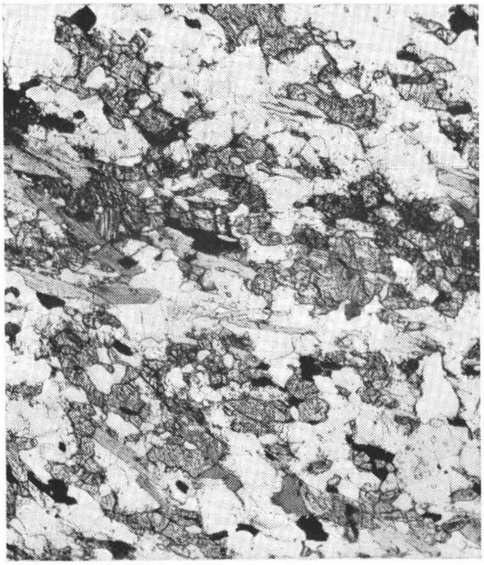


Fig. 2. Gabbro from the center of the fragment. Karlholm, Snappertuna. Nic. ||. 20 ×.

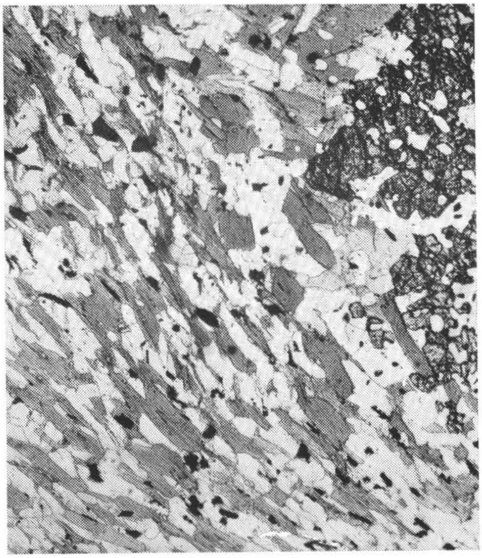


Fig. 3. Garnet-bearing biotite-plagioclase-quartz rock. Karlholm, Snappertuna. Nic. ||. 20 ×.

Maunu Härme and *Matti Laitala:* An example of granitization

10

STRATIGRAPHISCHE UNTERSUCHUNG ÜBER DEN BISHER NÖRDLICHSTEN REZENTEN *CAREX PSEUDOCYPERUS*-FUNDORT IN FENNOSKANDIEN, DEN SEE PITKÄLAMPI IN LIPERI, OSTFINNLAND [1]

VON

R. REPO

Geologische Forschungsanstalt in Finnland, Helsinki

AUSZUG

In der vorliegenden Arbeit wird eine botanische und stratigraphische Beschreibung des vom Verfasser aufgefundenen nördlichsten rezenten Standortes von *Carex pseudocyperus* in Fennoskandien gegeben. Auf der Sondernatur des Standortes beruht es, dass sich die Art seit der postglazialen Wärmeperiode an dieser Stelle hat erhalten können.

INHALT

EINLEITUNG

Bei der Bodenkartierung in dem Gebiet zwischen Joensuu und Outokumpu fand Verfasser am 13. IX. 1954 *Carex pseudocyperus* an dem Kleinsee Pitkälampi (ca. 62°40′ N) neben der Randmoräne Jaamankangas im Kirchspiel Liperi. Der Fundort ist der nördlichste der rezenten, mit Sicherheit bekannten fennoskandischen Funde dieser quartärgeologisch wie auch botanisch bedeutsamen Pflanzenart. Der nächstnördlichste Fundort der Art in Finnland liegt im Kirchspiel Joroinen, ca. 90 km SW von Liperi (Hintikka, 1917). Das frühere postglaziale Verbreitungsgebiet von *Carex pseudocyperus*

---

[1] Eingegangen am 12. März 1955.

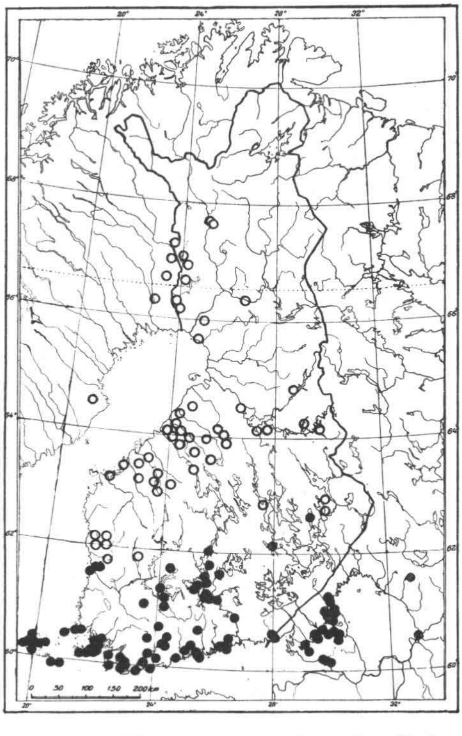


Abb. 1. Die rezenten ostfennoskandischen Vorkommen (Punkte) von *Carex pseudocyperus* und die wichtigsten subfossilen Funde (Kreise) nördlich des gegenwärtigen Wohnbereichs.

hat nach subfossilen Funden eine viel grössere Ausdehnung als heutzutags gehabt und während der boreal-atlantischen Wärmeperiode bis nach Lappland gereicht. Der bisher nördlichste Subfossilfund der Art ist nach freundlicher Mitteilung von Dr. A. L. Backman in Kittilä, 130 km nördlich vom Polarkreis gemacht worden. In den mittleren Teilen Finnlands ist *C. pseudocyperus* während der genannten Wärmeperiode in manchen Gebieten nördlich vom heutigen Wohnbereich, besonders in Ostbottnien, sogar ziemlich häufig gewesen (Backman, 1919, 1935, 1941, 1943, 1948, 1949, 1950; Auer, 1924, 1925; Aario, 1932; Valovirta, 1954; nach Forschungen von Valovirta auch: Virkkala, 1948; Okko, 1949; Mölder & Salmi, 1954). In den östlichen Teilen Finnlands gibt es jedoch deutlich weniger Funde.

Die beigelegene Karte (Abb. 1) zeigt bisher bekannte Funde von *Carex pseudocyperus* in Ostfennoskandien. Für den Anteil der rezenten Funde hat Mag. phil. V. Erkamo (Botanisches Institut der Universität Helsinki) mir die von ihm aufgenommene Karte gütigst zur Verfügung gestellt. An Subfossilienvorkommen liegen die von Backman im Atlas von Hultén (1950, S. 104) gemachten Fundangaben zugrunde, für die ich betreffs einiger nord- und ostfinnischen Funde mündlich Ergänzung von Backman bekommen habe. Der Fund von Suomussalmi ist von Valovirta (1954). Dazu habe ich auch einige Angaben von Aario (1932) mitaufgenommen.

BESCHREIBUNG DES STANDORTES

Der Kleinsee Pitkälampi liegt am Proximal(Nord-)rand der langen ostfinnischen Randmoräne Jaamankangas (vgl. Frosterus & Wilkman 1917), an der Ostseite des vom Jaamankangas nach Norden abzweigenden Oses Käsämänharju. Die Richtung des Pitkälampi ist etwa N—S, seine Länge 1 100 m, Breite 100—150 m und sein Wasserspiegel ca. 85 m ü.d.M. Auch ostseits des Sees erstreckt sich eine Osbildung, jedoch niedriger und undeutlicher als der Käsämänharju. Aus glazifluvialem Material besteht in der Hauptsache die Umgebungen des Sees, weil auch die Jaamankangas-Ablageung grösstenteils aus glazifluvialem Material aufgebaut ist (vgl. Frosterus & Wilkman, 1917). Nordöstlich vom Pitkälampi bildet warviger Ton ein Feldchen, und dieses Gebiet erreicht in kurzer Erstreckung das NE-Ufer des Sees. Der Boden ostseits des Pitkälampi ist bis in die Nähe des Ufers grösstenteils angebaut.

Im allgemeinen macht der See Pitkälampi einen ziemlich oligotrophen Eindruck (Abb. 2), wie die meisten Osseen und -weiher auch anderswo in Finnland. Doch gehören zu der Wasser- und Uferflora manche eutrophe Arten. *Carex pseudocyperus* wächst am Pitkälampi reichlich an den üppigeren, gyttjigen N- und S-Enden (Abb. 3) und verstreut um alle Ufer des Sees, auch an kiesigen und steinigen Stellen. Als Begleitarten von *Carex pseudocyperus* am Uferwasser des Pitkälampi (vgl. Repo, 1955) seien folgende erwähnt: *Calla palustris* (häufig und stellenweise sehr reichlich), *Comarum palustre* (auch häufig), *Alisma plantago-aquatica, Cicuta virosa, Lysimachia thyrsiflora, Menyanthes trifoliata, Carex rostrata, C. canescens, Peucedanum palustre, Lemna minor, Utricularia vulgaris, Phragmites communis* (nur am Ostufer, spärlich), *Typha latifolia* (nur 3 kleine Individuen).

Im Wasser, ausserhalb der Uferzone, wachsen u.a. *Potamogeton pusillus* (häufig), *P. natans, P. perfoliatus, P. obtusifolius, Sparganium simplex* f. *longissimum, S. Friesii, Nuphar, Nymphaea, Subularia aquatica, Elatine hydropiper, Utricularia vulgaris* und *Myriophyllum spicatum* (sehr spärlich, steril).

Die Ufer des Pitkälampi sind allgemein von *Alnus glutinosa, A. incana, Betula pubescens, B. verrucosa,* stellenweise auch von *Rhamnus frangula* und *Prunus padus* gesäumt. Höher hinauf, auf den trockenen Oshängen, ist *Pinus silvestris* die vorherrschende Holzart. Unter den Laubbäumen wächst vielenorts *Oxalis acetosella,* dazu erscheint hier und da u.a. auch *Lysimachia vulgaris* und spärlich in den Ackergräben nahe dem SSE-Ende des Sees *Bidens tripartitus.*

STRATIGRAPHISCHE BESCHREIBUNG

Zur Zeit des spätglazialen Yoldiameeres hat es schon kleinere Gebiete supra-aquatischen Landes in der nahen Umgebung des Pitkälampi gegeben. Nach Sauramo (1928) verläuft in der Nähe die 110 m-Isobase des Yoldiameeres in der Richtung NE—SW südostseits des Pitkälampi, demnach steigen u.a. die sogar bis über 120 m aufragenden Teile des Jaamankangas südseits des Pitkälampi viele meter über das genannte Niveau. Nach dem Yoldia hat das Becken des Pitkälampi zu dem Mittelfinnischen Grosssee und zu dem in der Litorinazeit von diesem abgeschnürten Vor-Satanen (Hellaakoski, 1922, 1936; Sauramo, 1940) gehört. Nach der Abzapfung des letztgenannten Vorsees in der Mitte der Litorinazeit beim Durchbruch der

Abflussrinne in Vuoksenniska (Hellaakoski, 1936, S. 104) wurde der weite See Viinijärvi (gegenwärtige Wasserfläche 150 km², ca. 78 m ü. M.) nördlich vom Jaamankangas abgetrennt. Der Pitkälampi scheint einige Zeit mit dem Viinijärvi vereinigt gewesen zu sein, aber bei späterer Abzapfung des Viinijärvi ist das Pitkälampi-Becken als getrennter Kleinsee bestehen geblieben. Der Höhenunterschied zwischen Viinijärvi und Pitkälampi beträgt nach der topogr. Karte (1: 100 000) etwa 7—8 m. Die Abtrennung von Viinijärvi scheint schon gegen Ende der Litorinazeit von sich gegangen zu sein.

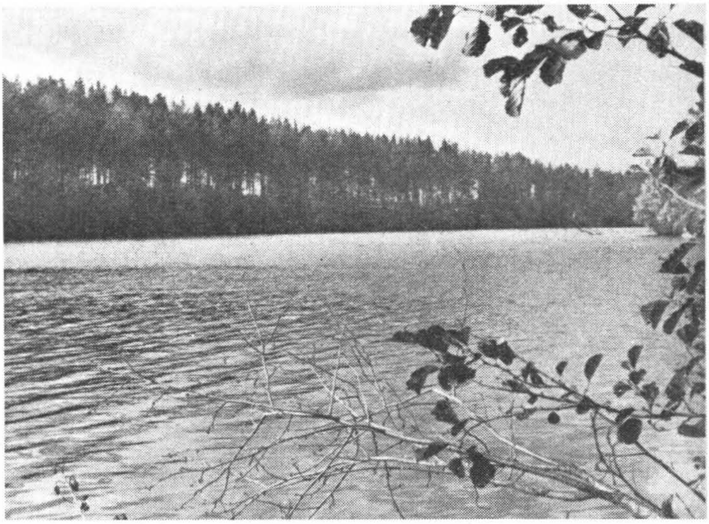


Abb. 2. Blick auf den See Pitkälampi. Im Hintergrunde der Os Käsämänharju.

Im einem ziemlich frühen Zeitabschnitt nach dem Verschwinden der Eisdecke hat *Carex pseudocyperus* nach dem oben Gesagten eine theoretische Möglichkeit gehabt, unter günstigen Verhältnissen seinen heutigen Standort in der Umgebung des Pitkälampi zu erreichen. Nach Valovirta (1954, S. 40) ist die Art nördlicher in Ostfinnland, im Gebiet Suomussalmi, subfossil in ancyluszeitlichen Ablagerungen angetroffen worden. Ferner sind einige rezente nord- und ostfinnische Vorkommen gewisser Wasserpflanzen am besten als alte spätglaziale (cfr. präboreale) Relikte zu verstehen (Kotilainen, 1954 a und b; vgl. Kalela, 1949; Sauramo, 1954 b). Anderseits ist jedoch zu bemerken, dass die seit den 1920er Jahren in Finnland festgestellte Klimaerwärmung (siehe: Hustich, 1952, A Symposium) für ihren Teil eine bedeutende nördliche Erweiterung des rezenten Areals vieler Pflanzenarten verursacht hat (Erkamo, 1952). Zu solchen rezent expansiven

Wärmearten gehören nach Erkamo (op. c.) auch manche Wasser- und Uferpflanzen. Die unweit des Pitkälampi in demselben Kirchspiel am See Särkijärvi rezent angetroffenen südlichen Wasserpflanzenraritäten *Najas tenuissima* und *N. flexilis* (Maristo, 1941; Backman, 1950; Kotilainen, 1951, 1954 b) z. B. sind nach Backman (1951) an diesem Standort späte Neuankömmlinge. Mit Rücksicht auf *Carex pseudocyperus* ist ferner zu bemerken, dass diese art gewissermassen auch kulturphil ist und sich leicht von ihren ursprünglichen Standorten auch nach Ackergräben und -weihern usw. ausbreiten kann (Lagerberg-Linkola-Väänänen, 1938, S. 229—230). Ohne eine auf Makrofossilien begründete Untersuchung ist es somit nicht möglich gewesen zu entscheiden, ob *Carex pseudocyperus* am Pitkälampi eine alte Reliktart oder ein junger, rezenter Einwanderer ist.

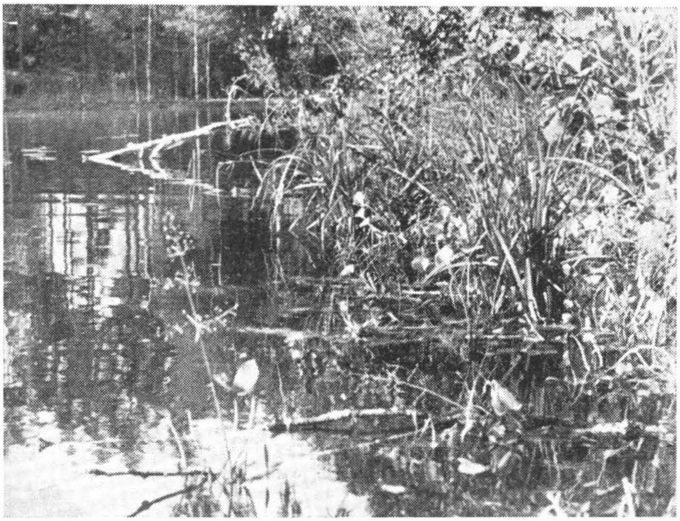


Abb. 3. *Carex pseudocyperus* an seinem Standort (rechts) am Nordende des Pitkälampi. Im Vordergrunde *Alisma plantago-aquatica.*

Für die stratigraphische Untersuchung entnahm ich den Ufersedimenten des Sees zwei Probereihen mit dem Moorbohrer, die eine am NE-Ufer, nahe dem Nordende des Sees, die andere am S(SE)-Ufer. In dem tieferen, dem Nordteil des Sees entnommenen Profil I (Abb. 4; — Mag. phil. Kyllikki Salminen verdanke ich die Pollenanalysen —) hat die Sedimentation organischen Materials am Anfang der atlantischen Klimaperiode (= Litorinazeit)[1] begonnen. Das betreffende Sediment besteht in den unteren Teilen

---

[1] Die Einteilung der Postglazialzeit in Finnland ist nach Hyyppä (1937) und Sauramo (1940, 1941, 1954, 1955). — Die Namen der Ostseephasen Ancylus, Litorina und Postlitorina hat Verfasser als Periodennamen für entsprechende waldgeschichtliche und klimatische Zeitabschnitte angewandt, obgleich Nordkarelien schon oberhalb der genannten baltischen Niveaus gelegen ist.

des Profils aus Grobdetritusgyttja, geht aber gegen Ende der Litorinazeit in ein auf seichtes Uferwasser hinweisendes Sediment mit Holzabfällen, limnischem Torf und Grobdetritus über, in den obersten Teilen in einen auf Zuwachs deutenden *Carex-Sphagnum*-Torf. Das kürzere Profil (II) besteht hauptsächlich aus grobdetritushaltigem, limnischem Torf.

Die Entwicklungsstadien des Pitkälampi-Beckens spiegeln sich deutlich in den *Diatomaceae*-Analysen wider, die Dr. Karl Mölder freundlicherweise für einige Proben von Profil I ausgeführt hat. Die Analyse über den Ton vom Grunde in 3 m Tiefe enthält grösstenteils (71 %) Formen von Süsswasserplankton. Sie weisen auf eine recht weite Wasserdecke hin, eine solche, wie sie in dem Gebiet während des ancyluszeitlichen Mittelfinnischen Grosssees bestanden hat. Die Analyse enthält folgende Formen:

Amphora ovalis 1 %, *Caloneis Schumanniana* var. *biconstricta* 1, *Cymbella prostrata* 1, *Epithemia zebra* 1, *Fragilaria virescens* 1, *Melosira ambigua* 14, *M. distans* 18, *M. granulata* 4, *M. granulata* var. *angustissima* 1, *M. islandica* ssp. *helvetica* 41, *Navicula Hustedtii* 3, *Pinnularia mesolepta* 1, *P. borealis* 1, *Stephanodiscus Hantzschii* 9, *Surirella* sp. fragm. 1, *Synedra ulna* 1, *Tetracyclus lacustris* 1, zusammen 100 %.

Die Analyse der litorinazeitlichen Grobdetritusgyttja von 2,5 m Tiefe, oberhalb der vorigen, enthält eine typische Binnenseeflora. Darunter gibt es 11 % Grossseeformen (*Amphora ovalis*, *Campylodiscus noricus* var. *hibernica*, *Cymbella aspera*, *Melosira arenaria*, *M. undulata*). Der Anteil der Planktonformen hat sich auf 31 % vermindert, die Menge der Litoralformen dagegen zugenommen, wie das folgende Artenverzeichnis erkennen lässt:

Amphora ovalis 2 %, *Campylodiscus noricus* var. *hibernica* 1, *Cyclotella comta* 4, *C. Kützingiana* 2, *C. stelligera* 1, *Cymbella aspera* 2, *Didymosphenia geminata* 1, *Diploneis puella* 1, *Eunotia faba* 1, *E. formica* 1, *E, parallela* 1, *E. veneris* 2, *Fragilaria atomus* 1, *F. brevistriata* 1, *F. construens* 1, *F. lapponica* 14, *F. pinnata* 6, *F. virescens* 4, *Gomphonema acuminatum* 1, *Gyrosigma acuminatum* 1, *Melosira arenaria* 3, *M. distans* 5, *M. granulata* 15, *M. islandica* ssp. *helvetica* 4, *M. italica* 2, *M. italica* var. *valida* 2, *M. undulata* 3, *Navicula pupula* 1, *N. radiosa* 1, *Pinnularia gibba* 2, *P. hemiptera* 2, *P. nobilis* 5, *P. viridis* 1, *Synedra nana* 1, *Tabellaria fenestrata* 3, *T. flocculosa* 1, *Tetracyclus lacustris* var. *strumosa* 1, zusammen 100 %.

Gegen Ende der Litorinazeit hat die Wasserdecke offenbar seicht zu werden begonnen. Das Sediment wird feinsandhaltig, und die *Diatomaceae*-Flora zeigt eine seichte Litoralzone des Kleinsees an. Der folgende Artenbestand ist für 2 m Tiefe festgestellt worden:

Achnanthes Biasolettina 1 %, *A. linearis* 2, *A. microcephala* 1, *A. Peragalli* 1, *Amphora ovalis* var. *pediculus* 2, *Cyclotella comta* 2, *C. Kützingiana* 1, *Cymbella affinis* 1, *C. aspera* 1, *Diploneis didyma* 1, *Eunotia veneris* 2, *Fragilaria atomus* 2, *F. brevistriata* 5, *F. construens* 3, *F. lapponica* 10, *F. pinnata* 6, *F. virescens* 1, *Gomphonema acuminatum* 2, *Melosira ambigua* 4, *M. distans* 8, *M. islandica* ssp. *helvetica* 1, *M. italica* 4, *Navicula anglica* 1, *N. bacilliformis* 2, *N. bacillum* var. *Gregoryana* 1, *N. cincta* 1,

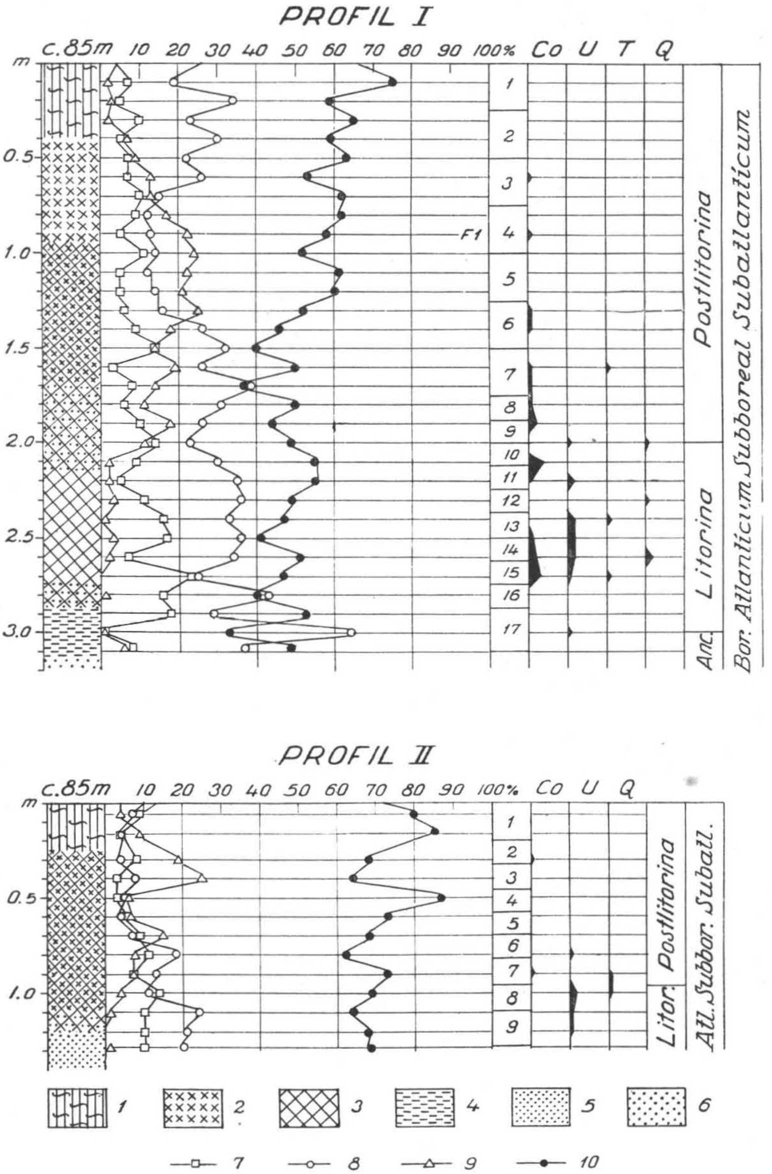


Abb. 4. Profile und Pollendiagramme aus dem See Pitkälampi. 1, *Carex-Sphagnum*-Torf; 2, limnischer Torf; 3, Grobdetritusgyttja; 4, Ton; 5, Feinsand; 6, Sand; 7, *Alnus;* 8, *Betula;* 9, *Picea;* 10, *Pinus;* Co, *Corylus;* U, *Ulmus;* T, *Tilia;* Q, *Quercus;* F, *Fagus.* Die Nummerkolumne gibt die Proben der Makrofossilienanalysen wieder.

N. cuspidata 1, *N. laterostrata* 1, *N. menisculus* 1, *N. muralis* 1, *N. pseudoscutiformis* 1, *N. radiosa* 1, *Nitzschia palea* 1, *Pinnularia gibba* 1, *P. hemiptera* 1, *P. interrupta* 1, *P. maior* 1, *P. mesolepta* 2, *P. microstauron* 2, *P. nobilis* 1, *P. polyonca* 2, *P. stromatophora* 1, *P. subsolaris* 2, *Stauroneis anceps* 1, *S. anceps* var. *hyaliana* 1, *S. phoenicenteron* 1, *Synedra ulna* 1, *Tabellaria fenestrata* 7, *T. flocculosa* 2, zusammen 100 %.

Die *Diatomaceae*-Analyse des postlitorinazeitlichen Teils des Profils in 1 m Tiefe, oberhalb des vorigen, zeigt, dass die offene Wasserfläche sich vermindert und der Zuwachs an der Stelle begonnen hat:

Achnanthes linearis 2 %, *Cyclotella Kützingiana* 1, *Cymbella aspera* 1, *Epithemia zebra* 1, *Eunotia exigua* 1, *E. parallela* 1, *E. pectinalis* 4, *E. pectinalis* var. *minor* 1, *E. tenella* 2, *E. veneris* 1, *Fragilaria lapponica* 3, *F. pinnata* 5, *Gomphonema acuminatum* 1, *G. constrictum* 1, *Melosira distans* 6, *Navicula bacilliformis* 1, *N. cryptocephala* 1, *N. radiosa* 10, *N. subtilissima* 1, *Pinnularia maior* 5, *P. molaris* 1, *P. nobilis* 36, *P. subsolaris* 1, *Surirella ovata* 1, *Synedra ulna* 2, *Tabellaris fenestrata* 3, *T. flocculosa* 4, *Tetracyclus lacustris* 3, zusammen 100 %.

Auf Makrofossilienanalysen gründet sich das genauere Verfolgen des Auftretens von *Carex pseudocyperus* sowie der anderen Wasser- und Uferarten am Pitkälampi, gewissermassen auch das Verfolgen der postglazialen Entwicklung der Vegetation in der Umgebung. Für diese Analysen habe ich beiden Bohrstellen zwei Probereihen entnommen, die

Tabelle 1. Makrofossilienverzeichnis für Profil I, Pitkälampi. Die Ziffern geben die Anzahl der Samen oder Früchte in den Proben an

	Postlitorina									Litorina							
Profil I Tiefe cm		50		100		150			200				250				300
Probe Nr.	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17
Alisma plantago-aquatica			1														
Alnus glutinosa.	1	1	8	10	11	8	4	2	1	1		2	2	5	15		
A. incana	3	1	1	2								1		2			
Andromeda polifolia												2	2				
Betula nana Fruchtschuppe/Samen..																-/1	-/1
B. pubescens —»—	5/2	5/3	5/6	1/7	3/5	-/2	4/3		-/2	-/3		-/2		1/1			
B. verrucosa —»—	14/8	5/4	2/8	-/3	4/3	10/4	18/8	4/2	1/1	3/4	3/3	3/5	2/5	-/2	3/5	3/4	
Calla palustris	2	1				1						1					
Carex canescens	52	30	13	3	13	12	5	2				1		1			1
C. diandra		1	1	1		1						1	1			1	
C. lasiocarpa							1					1	3	1			
C. pseudocyperus	1			1									1		3		
C. rostrata		3	5	5	1	1	1	1		1	1	2	1	1			
Cicuta virosa						1											
Comarum palustre			2	2		1					1		1	1	2		
Empetrum nigrum			2														
Lysimachia thyrsiflora	1			1		1											
Menyanthes trifoliata	4	6	1			3							1		1		
Peucedanum palustre	1			1	1												
Nuphar luteum			1	2			1	1									
Nymphaea alba													1	1	1		
Picea excelsa Nadel/Samen			7/-	10/2	15/1	8/1	7/-										
Pinus silvestris —»—	3/1	5/2	9/1	8/4	5/5	5/6	5/2	1/1	1/-	4/-	2/-	3/-	1/1	6/1	6/-	4/-	2/-
Potamogeton alpinus													1				
P. natans			1			3	1										
Rhamnus frangula		1															
Ranunculus sp.				1											1		
Scirpus lacuster							1	1			1						
S. paluster		2															

danach vereinigt worden sind. Die Analysen, für deren Ausführung ich Oberförster V. E. Valovirta bestens danke, gehen aus den Tabellen 1 und 2 hervor.

Die allgemeinen Züge der in der Umgebung vorherrschend gewesenen Vegetation spiegeln sich einigermassen in den Makrofossilien der Holzarten wider. Wie auch heutzutage, scheinen die Laubbäume *Betula pubescens*, *B. verrucosa*, *Alnus glutinosa* und *A. incana* (spärlicher) in dem ganzen Zeitabschnitt nach der Borealperiode die vorherrschenden Holzarten in der Uferzone, *Pinus silvestris* ähnlich wie obige, auf den trockenen Oshängen, gewesen zu sein. Dagegen kommt *Picea excelsa*, die auch heute in der nahen Umgebung des Pitkälampi viel seltener als die obengenannten Holzarten ist, als Makrofossil nur in den Ablagerungen der kühlen subatlantischen Klimaperiode vor, in dem Zeitabschnitt, wo *Picea*-Pollen auch am reichlichsten auftritt. Ziemlich gleichzeitig mit dem reichlichsten Vorkommen von *Picea* sind in beiden Profilen auch einzelne Samen von *Empetrum nigrum* angetroffen worden. Zwar ist die Art auch gegenwärtig in ganz

Tabelle 2. Makrofossilienverzeichnis für Profil II, Pitkälampi

	Postlitorina							Litor.	
Profil II Tiefe cm				50				100	
Probe Nr.	1	2	3	4	5	6	7	8	9
Alisma plantago-aquatica	4	2	1		1		3		1
Alnus glutinosa	6	10	14	20	16	2	20	34	7
Arctostaphylos uva-ursi			1		1				
Betula pubescens Fruchtschuppe/Samen	3/5	-/4	3/6	3/5	-/6	-/1	5/5	8/6	4/-
B. verrucosa —»—	1/1	-/2	-/3	1/3	1/7	3/5	14/12	14/5	9/-
Bidens tripartitus			1					1	
Carex canescens		2				2		1	
C. pseudocyperus			1					1	
C. rostrata	3	2	1	1	2	2	1		
Cicuta virosa		1	5	3		1	1	1	
Empetrum nigrum					1				
Myriophyllum alterniflorum					1				
M. spicatum									1
Nuphar luteum			1		1				
Nymphaea alba	1	1	2			1	5		1[1]
Picea excelsa Nadel/Samen.........			1/1	1/-	2/1				
Pinus silvestris —»—	-/1	6/-	8/3	7/3	3/5	12/3	-/4	2/1	4/1
Potamogeton natans								1	
P. perfoliatus	2	3	1	1		1	1		
P. pusillus		2							
Ranunculus sp.							1		
Rubus idaeus							1	1	
R. saxatilis									1
Sparganium minimum						1			
Stellaria sp.							1		
Thalictrum sp.			1						

---

[1] Gehört zum *N. candida*-Typ.

Finnland allgemein, aber als nördliche, auch in ungünstigen Verhältnissen fortkommende Art dürfte *Empetrum* durch sein Auftreten teilweise den kühlsten Abschnitt der subatlantischen Periode widerspiegeln.

Im allgemeinen kann man in der subfossilen Flora des Pitkälampi keinen bedeutenden Unterschied zwischen Litorina und Postlitorina feststellen. Alle als Makrofossilien vorkommenden Arten sind auch rezent, die meisten direkt am Pitkälampi, die übrigen in der nahen Umgebung anderswo im Kirchspiel Liperi, angetroffen worden. Die Gesamtmenge der Pflanzenarten der Profile wiederum ist in der kühlen Postlitorinaperiode ein wenig grösser als in der warmen Litorinazeit. Die Ursache dieser Erscheinung ist offenbar die Tatsache, dass die postlitorinazeitlichen Teile der Profile bedeutend länger sind und daburch mehr Probenmaterial enthalten haben. Kaum wäre z.B. anzunehmen, dass die in Finnland gegenwärtig ziemlich südlichen Arten (vgl. Hultén, 1950) *Lysimachia thyrsiflora, Peucedanum palustre* und *Rhamnus frangula*, die nur in postlitorinazeitlichen Teilen von Profil I und dazu rezent vorkommen, in der klimatisch güngstigen Litorinazeit nicht in der Gegend Pitkälampi aufgetreten wären, -obschon Reste dieser Arten zufällig nicht in dem kleinen Probenmaterial vorkommen. Anders als die vorigen ist *Betula nana* nur im Litorina (Profil I) angetroffen worden, obgleich sie vor allem in Ablagerungen der subatlantischen Periode zu erwarten wäre. Im ganzen scheint die Flora vom Pitkälampi also in den Hauptzügen schon seit der Litorinazeit gleich der gegenwärtigen gewesen zu sein.

Carex pseudocyperus tritt in den beiden Probenreihen schon in den Sedimenten der atlantischen Klimaperiode hervor. In dem längeren Profil wird die Art schon zu Beginn der genannten Periode angetroffen, sobald, nach Sediment und Fossilien, die Umstände auch für das Auftreten anderer Seichtwasser- und Uferpflanzen günstiger geworden sind. Schläuche der fraglichen Seggenart sind dazu (obschon spärlich) auch in den postlitorinazeitlichen Ablagerungen der beiden Profile gefunden worden. Weil sich auch Begleitarten von *Carex pseudocyperus* während der letztgenannten Zeit (im Profil I z.B. *Carex diandra*, im Profil II *Alisma plantago-aquatica)* zeigen, scheint es wahrscheinlich, dass *Carex pseudocyperus* nach seiner Ankunft am Pitkälampi seine Standorte zu behaupten vermocht und auch noch zur Zeit der subatlantischen Klimaverschlechterung ausreichende Bedingungen gefunden hat, sich bis zu unseren Tagen an seinem Wuchsort zu erhalten. An dem nachwärmezeitlichen Abnehmen des Fossilvorkommens dieser Art dürfte der auf klimatische Ursachen zurückzuführende Rückgang der generativen Fortpflanzungsfähigkeit schuld gewesen sein. Offenbar hat aber die in den letzten Jahrzehnten herrschende warme Klimaperiode sehr günstig auf die Verbesserung der Lebensverhältnisse dieser Art eingewirkt. Dies lässt sich an der reichlichen Fruchtbildung und dem Auftreten von Samenkeimlingen sowie daran erkennen, dass die Art heute überall an den Ufern des Sees verbreitet ist, auch auf dürftigeren Standorten. Diese Beobachtungen stimmen mit denen überein, die Erkamo (1952) in den letzten Jahren bei vielen anderen Wasser- und Uferpflanzen gemacht hat.

Weil die organischen Sedimente an den ufernahen Bohrstellen nicht tiefer als bis zur Litorinazeit reichen, bleibt es vorläufig unentschieden, ob *Carex pseudocyperus* schon vor der atlantischen Klimaperiode in dem Gebiet vorgekommen ist.

Sowohl in der rezenten als in der subfossilen (Prof. II) Flora vom Pitkälampi erweckt das Auftreten von *Myriophyllum spicatum* besonderes Interesse. Dieses in Finnland hauptsächlich nur im Brackwasser angetroffene Wasserblattkraut ist in mittelfinnischen Seen sehr selten (vgl. Hultén, 1950, Karte 1927), und diese Süsswasservorkommen scheinen Relikte (vgl. Kotilainen 1954 a und b) eines früheren, marinen Areals zu sein. Auer hat in Nordkarelien konstatiert (Sauramo & Auer, 1928, S. 37), dass die Art im See Höytiäinen (ca. 20 km NE vom Pitkälampi) schon zu Beginn der Ancyluszeit (oder des Boreals) vorgekommen ist, heutzutage wird aber die Art nicht im Höytiäinen angetroffen. Die bisher einzigen anderen rezenten Fundorte von *Myriophyllum spicatum* in den östlichen Teilen Mittelfinnlands (Särkijärvi in Liperi und Kotkatjärvi in Joroinen) sind Seen und Weiher an glazifluvialen Ablagerungen. Ein solcher See ist ja auch der Pitkälampi. Es ist deutlich beobachtet worden, dass das in glazifluvialen Ablagerungen sich reichlich ansammelnde und in den Randteilen dieser Sedimente hervorquellende Grundwasser durch seine edaphischen und thermischen Eigenschaften vielen Pflanzenarten günstige Sonderverhältnisse bietet. Diese Erscheinung ist sowohl in Nebenseen und -weihern als in Quellbächern und -mooren an Osen festzustellen (vgl. Kotilainen, 1936, 1942, 1954 b; Ulvinen, 1954; Repo, 1955; u.a.). Auch mikroklimatisch bieten die Ose der südlichen, auf Wärme angewiesenen Flora bessere Lebensverhältnisse als die Moränengebiete. Das rezente und subfossile Vorkommen von *Myriophyllum spicatum* zugleich mit *Carex pseudocyperus* zeugt auch für die Reliktnatur der letzteren Art am Pitkälampi. — Gegen die Möglichkeit, dass *Myriophyllum spicatum* erst spät angekommen wäre, spricht auch der Umstand, dass die Art auch in günstigerer Umgebung, in südfinnischen Brackwasser, nach Luther (1951, S. 249—250) heutzutage eine ziemlich geringe Fähigkeit generativer Fortpflanzung hat.

Der nächstnördlichste rezente (— und wahrscheinlich auch relikte; vgl. Hintikka, 1917, S. 107 —) Fundort von *Carex pseudocyperus*, der Kleinsee Immolanlampi in Joroinen, befindet sich auch neben einem Os. Ferner sei auch erwähnt, dass sich nach Okko (Jalas & Okko, 1951) *Cladium Mariscus* in seinen in Fennoskandien nördlichsten, disjunkten Vorkommen, am See Saarijärvi in Joroinen (Tiitinen, 1950) seit der atlantischen Klimaperiode erhalten hat. Auch dieser Fundort ist ein typischer Nebensee an einem Os.

Wenn man die Makrofossilienfloren der beiden Profile von Pitkälampi miteinander vergleicht, wendet sich die Aufmerksamkeit ausser den gemeinsamen Arten auch den lokalen Verschiedenheiten der beiden Standorte zu. Das nördlichere Profil (I) ist einer flacheren Umgebung, einem bis an das Ufer reichenden Tongebiet entnommen worden, wogegen in der Umgebung der südlicheren Bohrstelle überall Sand- und Kiesboden vorherrschen und ganz in der Nähe die hohen, steilen Oshänge des Jaamankangas aufsteigen. Der Artenbestand nahe der nördlicheren Bohrstelle ist schon von Anfang an bedeutend seggenreicher (u.a. *Carex diandra)* gewesen, und zu der Flora gehören da z.B. *Calla palustris*, *Scirpus paluster* und *Rhamnus frangula*, die sich nicht in den Ablagerungen von Profil II finden. In der letztgenannten Probenreihe dagegen sind Arten trockener Oshänge vertreten, besonders *Arctostaphyllos uva-ursi*, *Rubus saxatilis* und *R. idaeus.*

Allgemeine Veränderungen von Standortsfaktoren am Pitkälampi dürfte das Auftreten von *Scirpus lacuster* bezeugen. Diese Art, die ich am See nicht rezent bemerkt habe, scheint sich nur so lange erhalten zu haben, wie der Pitkälampi einem grösseren See vereinigt gewesen ist, nachher aber, in veränderten Zuwachsverhältnissen, scheint die Art an ihren Standorten nicht mehr fortbestanden zu haben.

Im Profil I wird in ca. 2 m Tiefe ein Übergang in ein sehr feinsandhaltiges Sediment bemerkbar. Diese Vermehrung von Mineralmaterial hat zu ihrer Zeit als bedeutender Faktor auf die floristische Natur des Ufers eingewirkt. Doch scheint sich nur um eine ganz lokale Erscheinung zu handeln (z.B. mag die Stelle Sedimentationsgebiet eines Einlaufbaches geworden sein), weil im Profil II in diesem Zeitabschnitt keine entsprechende Artenabnahme zu erkennen ist.

Auf die Verschiedenheiten zwischen der rezenten und der subfossilen Flora hat teilweise, besonders bei Arten trockener Standorte, neben klimatischen und quartärgeologischen Faktoren auch eingewirkt, dass sich die Natur der ursprünglichen Standorte heute zu einem grossen Teile durch Kultur verändert hat. So gibt es am südlichen Ende des Sees z.B. für das Gedeihen von *Thalictrum* günstige Bachuferwiesen und -gesträuche nicht mehr, obschon die Art (cfr. *T. flavum*, Profil II) bis gegen Ende der Postlitorinazeit in dem Gebiet vorgekommen ist.

ZUSAMMENFASSUNG

Wie die stratigraphische Untersuchung zeigt, ist *Carex pseudocyperus* auf seinem nördlichsten rezenten Standort, am See Pitkälampi in Liperi schon seit dem Anfang der atlantischen Klimaperiode aufgetreten. Für diese typisch wärmezeitliche Art, deren Wohnbereich während der klimatisch günstigen boreal-atlantischen Periode weit gewesen ist, hat der neben Osbildungen gelegene Standort Sonderverhältnisse geboten, und die Art hat sich hier offenbar bis durch die kühle subatlantische Periode, als sie schon anderswo in der Umgebung verschwunden war, erhalten können.

Die aus glazifluvialem Material aufgebauten Ablagerungen, wie sie die finnischen Ose und zum grössten Teile auch die Randmoränen (u.a. Jaamankangas und Salpausselkä) darstellen, können in ihren Schichten grosse Niederschlagsmengen als Grundwasser sammeln. Die grossen Hohlräume der Sand- und Kieslager lassen das Grundwasser leicht durchfliessen, und in den Randteilen der glazifluvialen Ablagerungen kommt das Wasser in reichlichen Quellen hervor. Wie in manchen botanischen Untersuchungen (u.a. Cajander, 1913, S. 128, 1916, S. 133—137; Kotilainen, 1927, 1936, 1942, 1951, 1954; Fagerström, 1941; Tiitinen, 1950; Ulvinen, 1954; Repo, 1955) festgestellt, kann das fortdauernd fliessende Quellwasser durch seine edaphischen und thermischen Eigenschaften eutrophen und südlichen Pflanzenarten günstige Lebensbedingungen bieten. Die Nebenseen und -weiher an Osen u.a. sind Fundorte von manchen anspruchsvollen Hydrophytenraritäten in Finnland (vgl. Kotilainen, 1951, 1954; u.a.). Eben an einem derartigen Nebensee, am Pitkälampi, wächst auch *Carex pseudocyperus*. Besonders günstige edaphische Verhältnisse scheint das Quellwasser zu bewirken, wenn es im Bezirk des Nebensees bei glazifluvialem Material auch Tonboden gibt (vgl. Immolanlampi, Särkijärvi, Pitkälampi u.s.w.). Auch mikroklimatisch dienen die Osgebiete für Wärme fordernde Arten als günstigere Standorte als die Moränengebieten (vgl. Jalas, 1950; u.a.).

Durch die verschiedenen, sowohl nassen als trockenen Standorte, die die langen, einheitlichen Ose in Finnland bieten, haben sie als wichtige Einwanderungswege der Pflanzen gedient. In diesen Hinsicht dürfte die Bedeutung der genannten Ablagerungen besonders wichtig in Ostfinnland gewesen sein, das sich schon in einer frühen spätglazialen Periode der Eisdecke entledigt hatte.

LITERATUR

AARIO, LEO (1932) Pflanzentopographische und paläogeographische Mooruntersuchungen in N-Satakunta. Fennia 55, No. 1.

AUER, VÄINÖ (1924) Die postglaziale Geschichte des Vanajavesisees. Comm. Inst. Quest. Forest. Finl. 8.

—»— (1925) Investigationes of the ancient flora of Häme (Tavastland). Comm. inst. Quest. Forest. Finl. 9.

BACKMAN, A. L. (1919) Torvmarksundersökningar i Mellersta Österbotten. Referat: Moor-Untersuchungen im Mittleren Österbotten. Acta forest. fenn. 12.

—»— (1935) Die nördlichsten Fossilfunde von Najas flexilis und Carex pseudocyperus in Finnland. Soc. Scient. Fenn. Comm. Biol. V. 3.

—»— (1941) Najas marina in Finnland während der Postglazialzeit. Acta Bot. Soc. Fauna et Flora Fenn. 30.

—»— (1943) Ceratophyllum submersum in Nordeuropa während der Postglazialzeit. Acta Bot. Soc. Fauna et Flora Fenn. 31.

—»— (1948) Najas flexilis in Europa während der Quartärzeit. Acta Bot. Soc. Fauna et Flora Fenn. 43.

—»— (1949) Mooruntersuchungen am Lappajärvi-See im Mittleren Ostbottnien. Bull. Comm. géol. Finlande 144, S. 151.

—»— (1950) Najas tenuissima (A. Br.) Magnus einst und jetzt. Soc. Scient. Fenn. Comm. Biol. X. 19.

—»— (1951) Najas flexilis funnen i Libelits. Memor. Soc. Fauna et Flora Fenn. 27, S. 3.

CAJANDER, A. K. (1913) Studien über die Moore Finnlands. Fennia 35, No. 5.

—»— (1916) Metsänhoidon perusteet I. Porvoo.

ERKAMO, VILJO (1952) On plant-biological phenomena accompanying the present climatic change. Fennia 75, S. 25.

FAGERSTRÖM, L. (1941) Anteckningar om floran och vegetationen i Terijoki socken. I. Memor. Soc. Fauna et Flora Fenn. 17, S. 121.

FROSTERUS, BENJ. & WILKMAN, W. W. (1917) Suomen Geologinen Yleiskartta. Lehti D 3. Joensuu. Maalajikartan selitys. Geologinen toimisto. Helsinki.

HELLAAKOSKI, AARO (1922) Suursaimaa. Referat: Gross-Saimaa. Fennia 43, 4.

—»— (1936) Das Alter des Vuoksi. Bull. Comm. géol. Finlande 115, S. 75.

HINTIKKA, T. J. (1917) Carex pseudocyperus L.-löytö Pohjois-Savosta ja havaintoja sen kasvupaikasta. Meddel. Soc. Fauna et Flora Fenn. 43, S. 97.

HULTEN, ERIC (1950) Atlas över växternas utbredning i Norden. Stockholm.

HUSTICH, ILMARI (1952) The recent climatic fluctuation in Finland and its consequenses. A. Symposium. Fennia 75.

HYYPPÄ, ESA (1937) Post-Glacial Changes of Shore-Line in South-Finland. Bull. Comm. géol. Finlande 120.

JALAS, JAAKKO (1950) Zur Kausalanalyse der Verbreitung einiger nordischen Os- und Sandpflanzen. Ann. Bot. Soc. »Vanamo» 24, N:o 1.

JALAS, J. & OKKO, V. (1951) Botanical and geological analysis of the Cladium Mariscus in Joroinen. Arch. Soc. »Vanamo» 5, S. 82.

KALELA, AARNO (1949) Mistä ja milloin Suomi on saanut kasvistonsa? Eräitä ääriviivoja. Suomen Luonto, S. 9.

KOTILAINEN, M. J. (1927) Untersuchungen über die Beziehungen zwischen der Pflanzendecke der Moore und der Beschaffenheit, besonders der Reaktion des Torfbodens. Veröff. des finnischen Moorkulturvereins 7.

—»— (1936) Carex paniculata L. Suomessa. Memor. Soc. Fauna et Flora Fenn. 12, S. 56.

—»— (1942) Carex diandra x paniculata (= C. germanica Richter) ja Carex canescens x paniculata (= C. silesica Figert) Suomessa. Memor. Soc. Fauna et Flora Fenn. 17, S. 193.

—»— (1951) Uusista kvartäärigeologisestikin huomattavista vesikasvilöydöistämme. Luonnon Tutkija 55, S. 37.

—»— (1954 a) The Stratiotes Lakes in Kittilä (Finnish Lapland) as a Floristic and Quaternary Problem. Botanisk Tidskrift 51, S. 141.

—»— (1954 b) Vaateliaiden uposlehtisten vesikasvien alueellisesta levinneisyydestä itäisessä Fennoskandiassa. Luonnon Tutkija 58, S. 136.

LAGERBERG, T., LINKOLA, K. & VÄÄNÄNEN, H. (1938) Pohjolan luonnonkasvit. I. Porvoo—Helsinki.

LUTHER, HANS (1951) Verbreitung und Ökologie der Höheren Wasserpflanzen im Brackwasser der Ekenäs-Gegend in Südfinnland. II. Acta Bot. Soc. Fauna et Flora Fenn. 50.

MARISTO, LAURI (1941) Die Seetypen Finnlands auf floristischer und vegetationsphysiognomischer Grundlage. Ann. Bot. Soc. »Vanamo» 15, N:o 5.

MÖLDER, K. & SALMI, M. (1954) Suomen geologinen yleiskartta. Lehti B 3. Vaasa. Maalajikartan selitys. Geologinen tutkimuslaitos. Helsinki.

OKKO, VEIKKO (1949) Suomen geologinen yleiskartta. Lehti B 4. Kokkola. Maalajikartan selitys. Geologinen tutkimuslaitos. Helsinki.

REPO, R. (1955) Piirteitä Jaamankankaan kasvimaantieteellisestä merkityksestä. Referat: Über die geobotanische Bedeutung der Randmoräne Jaamankangas (PK, Ostfinnland). Arch. Soc. »Vanamo». (im Druck).

SAURAMO, MATTI (1928) Über die spätglazialen Niveauverschiebungen in Nordkarelien, Finnland. Bull. Comm. géol. Finlande 80.

—»— (1940) Suomen luonnon kehitys jääkaudesta nykyaikaan. Porvoo—Helsinki.

—»— (1941) Die Geschichte der Wälder Finnlands. Geol. Rundschau 32, S. 579.

—»— (1954 a) Das Rätsel des Ancylussees. Geol. Rundschau 42, S. 197.

—»— (1954 b) Myöhäisjääkautisesta kasvistosta ja kasvillisuudesta, erityisesti metsänhistoriasta. Luonnon Tutkija 58, S. 130.

—»— (1955) On the Nature of the Quaternary crustal upwarping in Fennoskandia. Acta Geographica 14, S. 334.

SAURAMO, M. & AUER, V. (1928) On the development of Lake Höytiäinen in Carelia and its Ancient Flora. Bull. Comm. géol. Finlande 81.

TIITINEN, OLAVI (1950) Cladium Mariscus RBr. Joroisissa (Sb). Referat: Cladium Mariscus RBr. in Joroinen (Sb). Arch. Soc. »Vanamo» 4, S. 109.

ULVINEN, TAUNO (1954) Cardamine flexuosa With. in Luumäki gefunden. Arch. Soc. »Vanamo» 9, S. 44.

VALOVIRTA, V. E. (1950) Subfossiilinen Najas minor All. löydetty Suomesta. Suo 1950, S. 36.

—»— (1954) Suot. (In Maaperäkartan selitys. Lehti —4422. Suomussalmi.) Geologinen tutkimuslaitos. Helsinki.

VIRKKALA, K. (1948) Suomen Geologinen yleiskartta. Lehti D 4. Nurmes Maalajikartan selitys. Geologinen tutkimuslaitos Helsinki.

11

ABOUT THE GRANITE PROBLEM AND SOME MASTERS OF THE STUDY OF GRANITE [1]

BY

PENTTI ESKOLA

Institute of Geology, University of Helsinki

ABSTRACT

The impulsive to write this piece came from a paper by Walton (1955) on the emplacement of granite. To elucidate the human side of the sharp dispute between the magmatists and transformists, some of the chief protagonists are briefly sketched in a humoristic light. Thereafter the writer's present stand toward these problems is briefly summarized as an interim report and to state the respects in which he has revised his opinion in favor of the transformist school.

CONTENTS

INTRODUCTION

Today, the 4th of March 1955, I received a brilliant paper by Matt Walton (1955) dealing with the granite problem in a spirit which at once aroused in me the desire to express my satisfaction and to add some comments. Now that the granitic minimum has been established, the experimental difficulties in the physical chemistry of granite magma are mastered and, on the other hand, the real existence of metasomatic granites

---

[1] Received March 14, 1955.

is assured, one more round in the long debate on granite, opened by Werner and Hutton, seems to be nearing its end. It may be appropriate, therefore, to recall some of the personalities and episodes in this scholarly controversy during the last quarter of a century. The events of the past can, I hope, now be looked upon through the spectacles of humor, without anybody's feeling offended.

It is to be hoped that the clearing up of the dispute in a way that justifies both fractions to a certain extent will put an end to the turbulent state of affairs that has prevailed for more than two decades, reminding one of the poem by Gustaf Fröding:

What's settled truth in Leipzig and in Jena,
is just a stupid joke in Heidelberg.

The magmatists and the transformists have isolated themselves in their respective camps. It can be predicted that a pupil of Tilley and Nockolds will become a magmatist as surely as one of Holmes and Doris Reynolds will prove a transformist. This may be human; yet it is unfortunate. A recruit who has got an orthodox training in one or the other of the schools will become deaf and blind to the way of thinking of the opposite school. It may relieve the controversy if the students learn to know some human characteristics and perhaps some human faults in the influential representatives of both schools.

I have myself been wrestling with the problems of granite for half a century, during which my conclusions have been steadily forming. At present they appear rather definite, but as I have found it necessary at present to concentrate my energies upon general problems of the world and of life, it will still take a year or two before my results can be published. The present paper may therefore also serve as an interim report on the present stage of my ideas and as a statement of the points in which they have changed since my last papers on these problems (Eskola, 1950 and 1952).

GLIMPSES OF SOME LEADING PARTICIPANTS IN THE CONTROVERSY ON GRANITE

My recollections may be extended back to the time before the opening of the last dispute between the magmatists, or the »pontiffs», and the transformists, or the »soaks». In 1927 C. E. Wegmann (Fig. 2) came to Finland. In the prime of his youth, he was eager to introduce Alpine tectonics into this country, and we Finnish geologists, well conscious of our shortages in tectonics, were grateful to learn. J. J. Sederholm (Fig. 1), though more than twice as old as Wegmann, became perhaps his most enthusiastic pupil and gave proof of admirable vitality and susceptibility, while Wegmann certainly received impulses from Sederholm in treating the problem of granitization, a term which the latter had been using for a score of years earlier, though in a sense different from that in which it was later used by the transformist school. Sederholm spoke of anatexis and palingenesis, i.e. remelting and rebirth of granite, but he was a magmatist up to the end of his life in 1934. The great international excursion and conference which he organized in 1931 still fell into the pre-transformist period; H. G. Backlund and A. Holmes were good magmatists at that time. But it was during his work with E. H. Kranck in the archipelago east of Helsinki that Wegmann (1931) gradually began to nourish more nonmagmatic ideas.

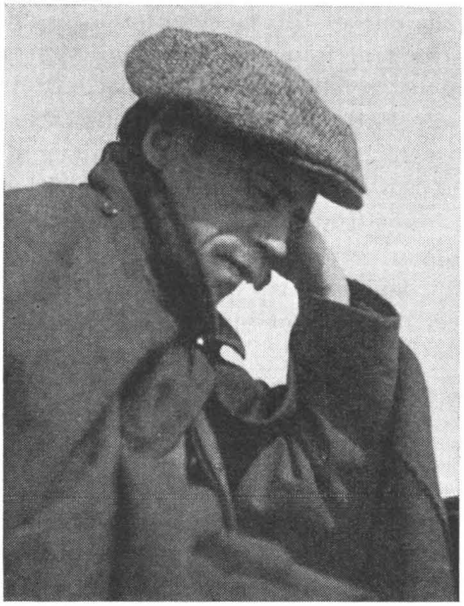


Fig. 1. Johannes Jakob Sederholm relaxing in a motorboat in the Pellinge archipelago 1914.

Fig. 2. Césare Eugen Wegmann at the Geological Survey of Finland about 1928.

During his stay in Finland Wegmann became best known as a humorist. His kindness and right-mindedness won him many friends, but sometimes his sense of humor was felt rather too strong for our latitudes. The elder generation still remembers the boyish tricks he played on some of our people who did not appeal to him. When I read in Wegmann's ingenious article »Zur Deutung der Migmatite» in Geologische Rundschau 1936 those winged words that the assumption of magma »verschleiert den Tatsachenbestand und hemmt die Forschung», I cannot avoid the suspicion that Wegmann here has, in the same manner, made fun of the world's wise geologists. For this assertion is perfectly unjustified[1].

---

[1] Perhaps some confusion has arisen from the current definition of magma as a porridge of crystals and liquid (Shand e.a.) in accordance with the meaning of the Greek word *magma*. Of course greater or smaller amounts of crystals may be suspended in magma, but the liquid phase alone is essential and subject to exact physicochemical theoretical and experimental treatment. It would be better to define the liquid phase alone as magma.

In spite of this, the said article has, certainly with full reason, had a revolutionary influence upon contemporary petrological thought. With the concepts of the migmatite front (though often abused) and intergranular film Wegmann has introduced into petrology two weapons that will nevermore be set aside. Although Wegmann is one of the most extreme representatives of the soak school, perhaps its real founder, as would appear from his early condemnation of magma just referred to, the said concepts are adapted for use by the pontiffs as well. Furthermore, his demonstration of the occurrence of relict stratification, or the old design (alte Zeichnungen), in granites has afforded conclusive evidence of metasomatic granitization of sediments. Being married and in his late fifties Wegmann appears more stabilized nowadays, but still, in congenial company, the old rascal frequently reappears from under his ever-widening tonsure and from behind his Mephistophelian mask. We may still await from him many a gallant idea and many a refreshing joke!

P. Niggli (Fig. 3) got from Wegmann the nickname of »Pope» and from Read (Fig. 4) of »Pontiff» (soon extended to all magmatists) because he evaluated his own stout person very highly and did not appear to have seriously doubted his infallibility. Another serious trouble with Niggli was his love of learned terms of Greek derivation. But his judgment was sharp and his conceptions of magma and crystallization were clear, and his physicochemical foundation was of a different class from that of the whole soak school. His really unpardonable fault was his lack of humor. Hans Cloos (Fig. 5) once said it was tragic that the two foremost geologists of Switzerland, Niggli and Wegmann, did not understand each other's language. Niggli's paper of 1942 on the granite problem has been very repulsive to Read, as appears from his »Meditations». This may be just a reaction incited by Niggli's unsmiling gravity, or also because Read could not fully follow Niggli's stern logic. Anyhow, the papers just cited may be highly recommended for comparison, and both give a fair historical outline of the undulations of the debate around the granite problem throughout a century and more.

In the heart of the City of London is the old club of the »Worshipful Company of Tallow Chandlers», where the Geological Society of London has had its anniversary dinners since the days of Lyell. There I saw, in 1948, H. H. Read's crooked-nosed Anglo-Saxon figure presiding in great splendor. His humor-filled devotion in performing the traditional rituals conformed admirably to the style of the tallow chandlers. I cannot remember having ever heard this man utter a sentence without a witticism, hidden or manifest, but with every sentence he also cut deeply into the subject matter. Read's two presidential addresses, entitled »Meditations on granite», have afforded the best possible defense to the soak cause and, although in my opinion he is wrong in his chief argument, I have learned

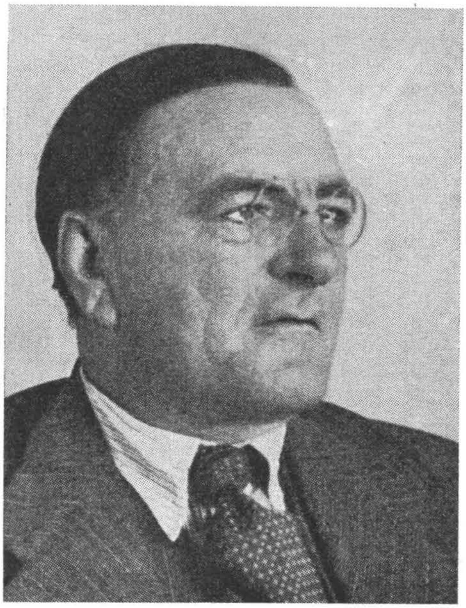


Fig. 3. Paul Niggli, picture published in a memorial address 1953.

Fig. 4. Herbert Harold Read on Sederholm's excursion in Finland 1931.

much of great value from these papers. Later, in his South-African address (Geol Soc. of S. Africa, Annexure to vol. L) Read appears strengthened in his assertion that there are »granites and granites». At present now that I find metasomatic granitization somewhat more important than I did before, there remains very little difference between our views.

The Geological Society of America edited in 1948 a symposium on the »Origin of Granite». One's curiosity is aroused by the chief pontiff's, N. L. Bowen's paper »The granite problem and the method of multiple prejudices». Bowen is a first-class stylist and has often proved rather sharp in his criticism. He was the man to coin the nick-name »soak» in retaliation to Read's »pontiff». But here he is unexpectedly gentle, and the irony in the title is not directed against the soaks alone. The Frenchmen M. Roubault and R. Perrin, the first ones to let dry atoms jump headlong into the granites, are knocked out an with easy hand, and H. Ramberg's reasonings on diffusion are dismissed as light-minded. The soaks are confronted with some questions they have never been able to answer: How is the relative enrichment of granites in iron to be accounted for, if it is not a consequence of crystallization differentiation? How is the zoning in crystals to be understood? One wonders why Bowen did not apply his heavy artillery, which would have been close at hand. Perhaps he regards sharp polemics unnecessary at the present stage of the debate. In collaboration with O. F.

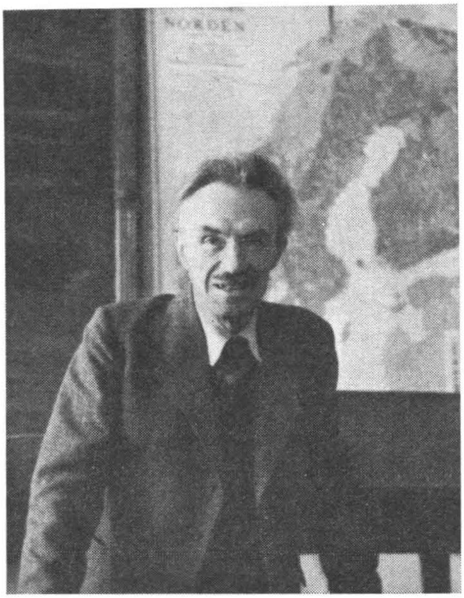


Fig. 5. Hans Cloos lecturing in Helsinki 1941.

Fig. 6. Norman Levi Bowen in Maryland 1921

Tuttle and others, Bowen has performed experimental work whose results, when published, will speak a more eloquent language than long sermons. Furthermore, he has also learned that there may be a certain amount of truth in the argument of the soaks as well.

Hans Cloos, the master of granite tectonics, was one of the first to show that the controversy is not irreconcilable. This is ably demonstrated in his dialogue with A. Rittmann »Zur Einteilung and Benennung der Plutone» in Geologische Rundschau 1941. Granites and basalts are shown to be radically different, as later proclaimed with great pondus by Read. The greatness of Cloos appeared in his sympathy for his fellow men, allies and antagonists alike. Brotherhood and co-operation between men and nations was the goal toward which he strove. A special chapter, worth once to be told in public, was his intimate friendship with C. E. Wegmann. I feel happy over the privilidge of having spent many a long period in the company of Hans Cloos, in his home at Reuterstrasse in Bonn, in Helsinki and, for the last time, in London during the Geological Congress. Of course we talked much about the problems of granite as well as other geological questions, but a still greater part of the time we spent together was devoted to discussing general human problems.

I put my head into the wasp swarm of a most heated debate with a lecture at the Geological Congress in London in 1948 on the nature of metasomatism in the processes of granitization (Eskola, 1950). My assertion was that many granites have resulted from a primary crystallization of granite magma, whereas others are products of metasomatic replacement of different rocks, and that the metasomatism was also effected by granitic magma soaking into the older rocks. Most of my arguments presented to support these theses still seem valid to me today, while some of them call for revision, or restriction. Walton has brought forth new evidence to support some of them.

THE IMPORTANCE OF ACID VOLCANICS

One of my arguments referred to the common occurrence of volcanic lavas of granitic composition. Under the heading »The existence of granite magma» Walton corrects the oft-repeated assertion that rhyolite volcanism, the equivalent of granite plutonism, is trivial in quantity compared to basaltic volcanism. On the contrary, as detailed knowledge of the cordilleran belts grows, we are constantly revising upward our estimates of the quantity of rhyolite volcanics, commonly disguised as ash and welded tuffs, etc., while in nonorogenic regions clays often have large contributions from primary acid pyroclastics. All these volcanics undoubtedly once existed in the form of molten lava, or magma. In the so-called granite batholiths intermediate types generally exceed in amount the granites in the strict sense of the word and so we must take into account the corresponding intermediate volcanics in addition to the rhyolitic end member.

As to the Archean of Finland, the recent development of ideas has been inclined to minimize the amount of acid volcanics and instead to interpret the leptites as sediments. Although sediments may indeed be more common in the Archean than was believed a score of years ago, I am convinced that *e.g.* Tuominen and Mikkola (1950 and Mikkola, 1955) greatly exaggerate the part played by true sediments. Apart from the obvious fact that acid and intermediate volcanics are fairly common, volcanic material is probably present in great amounts among the so-called micaschists.

Excessive alumina has been commonly regarded as a sure proof of the sedimentogeneous nature of metamorphics, especially when the excess appears in the presence of cordierite or almandite. In formations subjected to incipient granitization, however, this criterion may be fallacious, as recently proved by Härme and Laitala (1955). Enclosed in granite they found a fragment of amphibole gabbro composed of bytownite and cummingtonite. A contact zone of this fragment has been potash-metasomatized and contains very little plagioclase and no amphibole but instead much biotite and some almandite, possibly also microcline. Thus calcium has been replaced by potash with the result that the rock now contains some excessive alumina:

$$CaAl_2Si_2O_8 + K_2 + 4\ SiO_2 = 2KAlSi_3O_8 + Ca + Al,$$

giving rise to almandite, or cordierite. Thus an appearance of argillitic composition may arise from potash metasomatism. The result would be the same if the metasomatizing agent were granitic magma. As andesites and dacites are fairly common among orogenic volcanics, it is quite possible that some cordierite- and almandite-bearing rocks, such as the so-called kinzigites, may have originated through granitization from intermediate volcanics. Truly sedimentogenous kinzigites, etc., undoubtedly exist as well. Their mode of origin is conclusively proven in cases where they show a varved structure which is rather common. In the Swedish part of the Svionian the leptites are generally supposed to be of volcanic origin. This is no doubt generally a correct interpretation. Thus the acid volcanics are probably no less common than the basic ones.

Proterozoic acid volcanics underlie extensive areas in Dalecarlia in Sweden and in the bottom of the Gulf of Bothnia and the Baltic near Gotland (Eskola, 1934), and on Suursaari. These porphyries are assuredly products of crystallization of melts, or magmas, and so are also the graphophyres (Eskola, 1928), which also underlie vast areas in the bottom of Golf of Bothnia. I called the graphophyres rapakivi rocks because they are structurally similar to many varieties of rapakivi.

THE MAGMATIC CHARACTER OF RAPAKIVI

For Finnish geologists rapakivi is a rock type whose primarily magmatic character is most assured. Therefore Backlund's interpretation of rapakivi as metasomatized arkose was of the nature to make the transformist views in general unpopular in Finland. To the abundant evidence of the magmatic nature of rapakivi (see Eskola, 1950), Walton's paper now adds a new and very strong argument. Under the headline »The significance of 'disharmonious' granites» Walton writes (p. 8):

> »Now it seems to me that there are certain geologic environments in which granitic bodies up to major bathylithic size are occasionally emplaced that present insuperable difficulties to granitization hypotheses. These are the terrains of entirely unmetamorphosed sediments invaded by granite masses that are surrounded by well-defined, high temperature contact metamorphic aureoles. — — — — — — —
> The contact metamorphic aureole in which minerals such as clays, stable at surface temperatures, are crystallized into minerals stable at higher temperature proves that the disharmonious granite mass represents a highly localized source of heat in a low temperature environment.»

It is easy to recognize such disharmonious granites *e.g.* in the plutons of the Oslo area emplaced into unmetamorphosed sediments. But the rapakivi masses apparently also belong to this group, although in Finland they are all emplaced in a crystalline, old pre-Cambrian environment where the rocks have been highly metamorphic before the emplacement.

It has been generally assumed that the rapakivi would be older than the Jotnian sandstone. The general opinion was that the potash feldspar of the arkosic sandstone would be rapakivi feldspar, but just recently Simonen (Simonen and Kouvo, 1955) has ascertained that they are invariably microcline; whereas orthoclase, the typical feldspar of rapakivi, was nowhere found, although the sandstone along the southwestern margin of the Jotnian area of Satakunta is bounded by the Laitila rapakivi, the immediate contacts being covered by Pleistocene deposits. It seems therefore that the rapakivi is younger than the unmetamorphosed sandstone and intrusive in it. In Sweden a rapakivi-like granite is, according to von Eckermann, in fact found cutting the sandstone in the Loos-Hamra region. Thus the rapakivi is a typically disharmonious granite. We can say this also without knowing its relation to the sandstone, and even if the rapakivi were, as commonly believed, subjacent to the sandstone, on the basis of the contact phenomena seen in the Archean country rocks.

One of the best outcrops of the Laitila rapakivi contact is situated about 0.5 km south of the church of Kodisjoki (Fig. 7). In a flat rock surface the rapakivi is typical, with rounded ovoids. Toward the contact first smaller and gradually larger fragments of hornblende-gneiss appear. They are angular and have been turned into different positions, but otherwise they have not been changed appreciably. Finally, big fragments are almost in situ. The rapakivi here loses its porphyritic structure but is coarse-grained

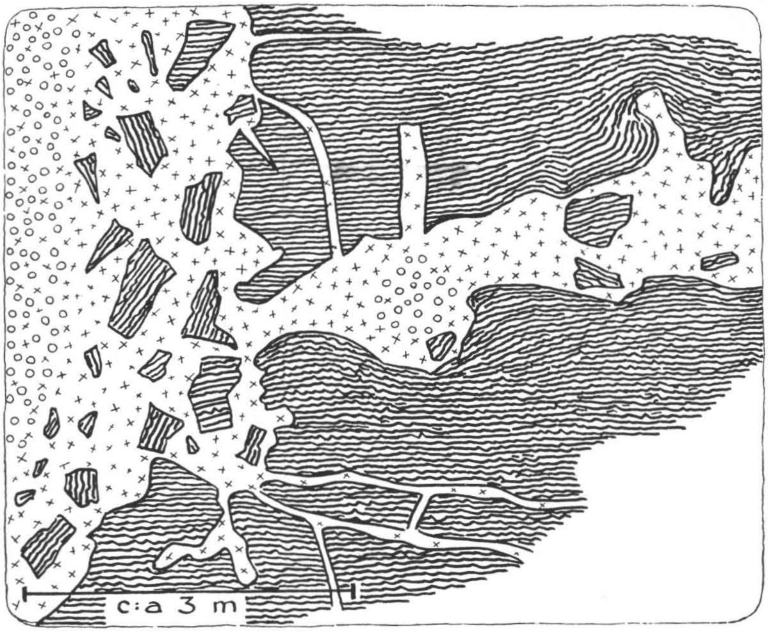


Fig. 7. Intrusive contact between rapakivi and hornblende gneiss sketched by the writer during his first attempts to study granites in 1905.

and so are the apophyses, excepting widened portions in which the rapakivi structure still reappears 20 m from the contact. Farther away the apophyses fade away in the migmatic gneiss. As I have said (Eskola 1950, p. 7):

> »The rapakivi retains its typical structure up to the very contact and even in the breccia, showing that tranquility of structure which is characteristic of the rapakivi, with no signs of the movement that is so strikingly manifested in the breccia. The rapakivi must have crystallized exactly at the place it now occupies. It is a post-kinematic pluton.»

This type of contact is dominant around all rapakivi plutons. Less common is that type referred to by Read in his »Meditations» II p. 76, i.e. a gneiss in which rapakivi ovoids occur in the schistose country rock. A single example of this type is known to me from the Laitila pluton. Read regarded it as a proof of the nonmagmatic nature of the rapakivi itself, but considering that the phenomenon occurs in a contact zone only a few meters thick my interpretation *(loc. cit)* that the »porphyroblasts» have »grown out from the pore magma permeating some way into the country rock», is quite natural. This, I gather, is also in accordance with Walton's concept of disharmonious plutons.

GRANITIZATION AND GRANITE MAGMA

By producing several examples Walton is able to account for the room problem, so often used as an argument for transformist views, and he even refers to a »negative room problem». In Finland metasomatically granitized masses frequently display a swelling up. This is most apparent in mantled domes, but also elsewhere in migmatite areas where granitization has been particularly strong. This is one evidence that granitic magma has been intruded at metasomatic granitization as well.

The arguments that I have brought forth for my thesis that metasomatic granitization also has been effected by granitic magma have not been disproved, as far as I know. Walton also seems to admit that some melting can take place at the levels where granitization in a metamorphic environment is in operation, though apparently he thinks that the process mainly occurs without the mediation of magma.

One of my arguments was that so many different rocks, such as clayey sediments, intermediate plutonics and volcanics, quartzites, and limestones may be converted into granites, eventually tending towards a rather constant ideal-granitic composition. As they need very different additions of substances to attain such a composition, yet all get what they need, it is hard to see why it should not be assumed that the granitizing agent has been granite magma. This is most apparent in the case of quartzites, which really must get all the granite elements except silicon and oxygen. If all these elements are introduced simultaneously, they will make a magma. One might imagine them wandering one after the other, yet there is no evidence that intermediate products exist. In the case of limestone, again, this is the case, the intermediate products being represented by the skarns. These are very resistent to granitization and, eventually to become granitized, they must get potash, soda, and alumina, which hardly can move in rocks except as granite magma.

In the discussion after my congress lecture, Doris L. Reynolds remarked to this argument of mine that when pyroxene was converted to amphibole only those elements necessary for the change were fixed, and in that small-scale example of transformation there could be no doubt that ions and not magma entered the pyroxene. To this it can be responded, first, that uralitization of pyroxene signifies a change of facies and occurs only when the compounds needed already exist in the rocks, and second, that different kinds of metasomatism involve migration of one or other kind of ions, but when all ions existing in granites are present, then there is magma, or a concentrated silicate solution. The assumption of very diluted silicate solutions is contradicted by the hiatus between the crystallization of pegmatite and residual quartz, and by the experimental results that the residual solution after the crystallization of silicates contains almost nothing but silica.

I must, however, admit that the case is different with rocks like clay, or diorite, which in themselves contain most of the granite elements. In many cases potash metasomatism alone is sufficient to convert them into granites. This seems actually to be the case, and my earlier standpoint in this respect must be revised.

THE GRANITE MAGMA AS A EUTECTOID

Now we come to my third important argument for the magmatic nature of most granites, and that is the fact that the composition of granite is very nearly constant and, moreover, as now appears from the experimental studies on the granite system carried out in the Geophysical Laboratory by Bowen, Tuttle *et al.*, that this is precisely the composition of mixtures next to the minimum in the solidus as well as liquidus surfaces of the haplogranitic system. This alone is conclusive evidence, for it is inconceivable that metasomatism of any kind could possibly bring forth, or even tend toward, such a composition as is the natural end product of magmatic crystallization differentiation.

Of course the minimum point (»the granitic eutectic») has real significance only if the crystallization of actual granite magmas ends with the crystallization of the granite minerals and the remaining aqueous solution is very diluted. In other words, there must be no continuity between granite magmas and hydrothermal solutions, and it is not possible that a gradually increasing concentration of water can depress the crystallization temperature much under the minimum point in dry melts. As well known long ago, such a break in crystallization does not occur if the magma contains appreciable amounts of »carriers», *i.e.* such elements or compounds as carbon dioxide, fluorine, sulfur. In such cases the differentiation results in lamprophyres, carbonatites, fluorite veins, and sulfidic ores. But an excessive content of alkalies may have a similar result. On this point Dr. O. F. Tuttle kindly communicated to me some of his experimental results, indicating that in melts containing potash and soda in excess over the ratio of these in the feldspars the path of crystallization finally departs from the quartz-feldspar join and the residual liquid will be continually enriched in water, the crystallization finally ending within the hydrothermal field with alkali silicates, or complex silicates of alkali and other cations, such as aegirite or riebeckite. As compositions with excessive alkalies are extremely rare, especially in the Archean territories, the departure from magmatic crystallization in the course of crystallization differentiation is of little practical concern.

Nevertheless, alkali silicates may be brought by means of diffusion into rocks already crystallized and effect changes in the minerals. At proper proportions of the substances added, this may lead to granitization, e.g. when alkali silicates are added. This is another case where my earlier conclusion must be modified and the possibility of hydrothermal granitization admitted.

Normally the introduction of alkalies into normal granites with almost equal alkali/alumina ratios would lead to agpaitic alkali rocks, such as fenite, but if there is an ever so small excess of alumina contained in aluminium silicate minerals, the first effect will be the conversion of these into alkali-bearing minerals, mostly micas. The very common occurrence of micaceous pseudomorphs after cordierite, almandite, or andalusite in our granites proves that such a process is in fact fairly common. The change, then, signifies potash metasomatism.

Real granitization by means of potash metasomatism may take place in intermediate or basic magmatic rocks, as stated in the foregoing (p. 124) This kind of granitization is exceedingly common in the Finnish Archean, but it can well have and probably often has taken place by simple injection of granite magma.

Recent geological mapping has shown that metasomatic granitization has been fairly common in the Archean of Finland. In 1948 I still believed that the two tectonic groups most important within our Archean (pre-rapakivi) rocks, *viz* the synkinematic and late-kinematic granites, would both comprise granites resulting from crystallization differentiation as well as such formed by metasomatism of original sediments and intermediate or basic magmatic rocks. In my discussion of 1952 I had learned that most synkinematic granites of ideal-granitic composition are metasomatic. Today I must, relying upon the mapping and research of the Survey geologists, admit that there may be no synkinematic ideal granites of original bulk composition in Finland.

Among the granites of the late-kinematic group metasomatic types with preserved »ghostly relics» or »old design» of stratification are fairly common. Walton says (p. 11):

> »As far as present knowledge is concerned it seems obvious to me that both granitization and granite magma can be shown in certain cases to have operated on a sufficiently large scale to be regarded as major factors in the emplacement of granite, but a broad shadow zone of uncertainty exists between the definable extremes, probably encompassing a major part of the granites of the world, in which we are unable to define the contribution of the two processes with any certainty.»

I would class with the metasomatic granites all those that show relict features indicating that they had earlier been something else than they are at present, and all the others as magmatic. There are probably always portions, mixed with rocks of the metasomatic mode of origin by replacement, that have had a liquid phase from which the crystallization has taken place in a truly magmatic way. But I still believe that also a considerable part of the granites that, according to the said definition, are classed with metasomatic rocks, have been metasomatized by granite magma, which probably is for the greater part anatectic, but which can also include truly juvenile material. This belief of mine is based upon the fact that these granites commonly have a typical eutectoid, ideal-granitic bulk composition. As far as I know, nobody has been able to explain how such a composition could possibly originate from metasomatism effected by hydrothermal solutions, or free cations, as all sure examples of nonmagmatic metasomatic rocks show variable compositions which do not follow the rules of crystallization differentiation, as the granites in question actually do.

Rocks of variable noneutectoid composition are also found among the granitic-looking rocks. They may deviate from the eutectoid composition either in their lesser content of alkalies, in an abnormal proportion of potash and soda, or in an amount of silica in excess over the eutectoid ratio. Such rocks may have resulted from a truly hydrothermal metasomatism. In extensive granite areas, however, such rocks are hardly found.

The granites of eutectoid composition can thus be classed into those showing evidence of metasomatic emplacement, and homogeneous granites with no such evidence. The latter usually show a more cross-cutting relation to their country rocks. Such granites, as a rule, can be regarded as products of direct crystallization from magma, either juvenile or anatectic, or mixtures of both.

The rapakivi plutons are the purest magmatic granites in our country. From them we can follow all gradations to metasomatic granites. Walton is to be credited for having elucidated the magmatic emplacement, making it clear that the purest magmatic rocks are commonly found in disharmonious plutons.

REFERENCES

(The papers by Wegmann, Read, Niggli, Bowen and Hans Cloos mentioned in the text are not taken up in the list below, as they are supposed to be known to every petrologist.)

ESKOLA, PENTTI (1928) On rapakivi rock from the bottom of the Gulf of Bothnia. Fennia 50, No. 27.

—»— (1934) Über die Bottenmeerporphyre. Bull. Comm. géol. Finlande. 104, p. 111.

—»— (1950) The nature of metasomatism in the processes of granitization. Intern. Geol. Congr., Report 18th Session, Great Britain 1948, III, p.l.

—»— (1952) A discussion of domes and granites and ores. Bull. Comm. géol. Finlande 157, p. 125.

HÄRME, MAUNU, and MATTI LAITALA (1955) An example of granitization. Bull. Comm. géol. Finlande 168, p. 95.

MIKKOLA, TOIVO (1955) Origin of ultrabasics in the Orijärvi region. Bull. Comm. géol. Finlande 168, p. 39.

SIMONEN, AHTI and OLAVI KOUVO (1955) Sandstones in Finland. Bull. Comm. géol. Finlande 168, p. 57.

TUOMINEN, HEIKKI, and TOIVO MIKKOLA (1950) Metamorphic Mg-Fe enrichment in the Orijärvi region as related to folding. Bull. Comm. géol. Finlande 150, p. 67.

WALTON, MATT (1955) The emplacement of »granite». Am. J. Sci. 253, p. 1.

12

SCAPOLITE FROM LÅNGÖ, ÅVA AREA, ÅLAND ISLANDS [1]

BY

SIMO KAITARO

Geological Survey of Finland, Helsinki

ABSTRACT

A scapolite pegmatite in the Åva area is described. The chemical analysis, optical properties and x-ray powder diffraction data for the scapolite, containing 36 % meionite, are given.

Although scapolite is relatively common as a minor constituent, especially in sedimentogeneous rocks, scapolite occurrences proper are rare. Therefore every new occurrence arouses interest. When mapping the Åva area, the present author found scapolite on the southern shore of the island of Långö in the parish of Brändö, Åland Islands. There it occurs immediately on the shore in an area of about 1 m^2, but apparently only as a layer some cm thick. The general dip of the rock surface as well as of the lower contact of the scapolite dike seems to be a gentle one towards the center of the area, similar to that of many pegmatites, especially in the eastern border zone of the Åva granite (Kaitaro, 1953). The country rock is a heterogeneous monzonite.

The pegmatite dike, the predominant mineral of which is scapolite, is relatively sharply bordered against the wall rock. The contact is very sinuous, especially on a small scale. Its breadth cannot be determined, because the upper contact is not visible. Scapolite is pale greenish yellow with a characteristic greasy luster and forms radiating groups some cm in length. It has no definite crystal forms. The outlines are rather rounded, but a distinct prismatic cleavage gives it the characteristic appearance.

The other minerals of the pegmatite, in the approximate order of their abundance, are: biotite, allanite, quartz, fluorite, calcite, magnetite and apatite. Biotite, which also forms greater aggregates, is much altered and generally green in the thin section. Allanite mostly occurs as jetty tabular

---

[1] Received March 4, 1955.

plates 2—3 cm in length. It has altered into metamict material (refraction index = 1.69). Therefore its accurate optical determination has not been possible and the x-ray diffraction analysis shows no crystal structure, but it is similar to a mineral that, occurring in the same area, has been determined by a chemical analysis as allanite (Kaitaro, 1953). Around allanite there commonly occur radiating fractures, apparently caused by an increase in volume through radioactive alteration. Quartz occurs sometimes as separate grains and very often as small inclusions and in narrow fractures in scapolite. Fluorite, partly violet or bluish, is found only in minor amounts, especially in narrow fractures. Apatite occurs occasionally as inclusions in

Table 1. Scapolite. Långö, Brändö, Åland Islands. Analyst, P. Ojanperä

Constituent	%	Atomic prop.	Physical prop.
SiO_2	55.78	$\left.\begin{matrix}\mathrm{Si} & 928\\ \mathrm{Al} & 438\end{matrix}\right\}$ 12 × 114	$\omega = 1.565$
TiO_2	0.01		$\varepsilon = 1.544$
Al_2O_3	22.34	$\left.\begin{matrix}\mathrm{Mg} & 1\\ \mathrm{Ca} & 145\\ \mathrm{Na} & 239\\ \mathrm{K} & 19\end{matrix}\right\}$ 4 × 101	$\omega - \varepsilon = 0.021$
Fe_2O_3	0.46		
FeO	0.21	$\left.\begin{matrix}\mathrm{Cl} & 61\\ SO_3 & 8\\ CO_2 & 47\end{matrix}\right\}$ 116	Sp.gr. = 2.675 ± 0.005
MnO	0.02		(Clerici solution and Westphal balance)
MgO	0.05	$(Na,Ca,K,Mg)_4(Si,Al)_{12}O_{24}(CO_3,SO_4,Cl)$	
CaO	8.17		
Na_2O	7.39		
K_2O	0.89		
Cl	2.16		
F	0.01		
SO_3	0.66		
CO_3	2.06		
H_2O + ...	0.65		
H_2O —	0.04		
	102.30		
Less O for Cl	1.89		
	100.41		

scapolite. Across the dike there go some straight narrow joints, which are visible only because there occur reddish pigments along it. Besides there occur sericitic and chloritic alteration products and, in border zones, remains of altered feldspar (plagioclase).

The impure contact zone has been avoided in taking the scapolite material for further investigations. The material purified by centrifuging in Clerici solution has been analysed and the result, together with its physical data, is given in Table 1. The optical properties have been determined by Mrs. T. Mikkola, M. A. The purity of the analysed material was tested under the microscope. The main impurity was quartz, but it was found to be very small in amount. The analytical result agrees with this observation. Besides, there occur some turbid or reddish pigmented scapolite grains. The calculations yield a slight deficiency of cations, especially with regard to anions other than SiO_2. This may not result from impurities or analytical errors alone. Iron oxides were neglected, because they are definitely in the form of oxide pigment. Some other amounts are too small and apparently included in minor impurities.

The material in question has also been studied by means of the x-ray powder diffraction analysis. The apparatus used was the Philips Geiger-counter x-ray spectrometer, CuK_{α}-radiating with a Ni-filter. The result of this study is given in Table 2.

Table 2. X-ray diffraction data of scapolite from Långö

d Å	I	d Å	I	d Å	I
6.05	20	2.73	30	1.815	10
4.26	10	2.69	60	1.705	10
3.82	70	2.30	20	1.559	10
3.55	30	2.14	20	1.463	10
3.47	100	2.07	10	1.402	10
3.08	100	2.02	10	1.382	20
3.02	40	1.920	20	1.364	20
2.84	20	1.897	10		

In the calculations presented (Table 1) the possibility has not been taken into account that sulfate and carbonate radicals may be contained in the sodic component forming $NaHCO_3$ and $NaHSO_4$ (Winchell, 1951). This scapolite contains these radicals in excess in relation to calcium. All the SO_3, as well as a part of the CO_2, is thus apparently contained in the marialite component and the formula for the scapolite may be presented as follows: $(Na, K)_4$ (Cl, HCO_3, HSO_4) $Si_9Al_3O_{24} \cdot Ca_4CO_3Si_6Al_6O_{24}$. The amount of the meionite component is about 36 % (dipyre). In this respect the scapolite from Långö differs from those of Laurinkari (Borgström, 1914) and Pusunsaari (Laitakari, 1947), which contain more meionite, 66 % and 75 % respectively.

REFERENCES

BORGSTRÖM, L. H. (1914) Die Skapolithlagerstätte von Laurinkari. Bull. Comm. géol. Finlande 41.

KAITARO, S. (1953) Geologic structure of the late pre-Cambrian intrusives in the Åva area, Åland Islands. Bull. Comm. géol. Finlande 162.

LAITAKARI, AARNE (1947) The scapolite occurrence of Pusunsaari. Bull. Comm. géol. Finlande 140, p. 115.

WINCHELL, A. N. (1951) Elements of optical mineralogy. Part II. Fourth Edition. New York.

13

ON PLEOCHROIC HALOS IN SOME RAPAKIVI ROCKS [1]

BY

SIMO KAITARO

Geological Survey of Finland, Helsinki

ABSTRACT

The occurrence of pleochroic halos in rapakivi rocks in general is briefly reported. Two rapakivi rocks, viz. the granites of Ytö and Sinkko, are found to be extremely rich in pleochroic halos. Statistical and quantitative studies have been carried out on the halos in these rocks. Some observations of the radioactivity of the rocks have been made in the field with a Geiger-counter.

It is common knowledge that around some radioactive minerals there may occur pleochroic darkenings or halos in the host mineral, e.g. biotite and hornblende. The alpha-particles, emitted from some radioactive element of the nucleus mineral, have given rise to a halo. Because radioactive elements form families or series in which later disintegration products may emit alpha-particles, we may have halos with several concentric rings. The ring radius depends on the rate of decay of the element in question and the relative stopping power of the host mineral (cf. Rankama, 1954). Thus, in well-developed halos it is possible to determine the radioactive element or elements contained in nuclei, but not the nucleus mineral itself. The existence of halos is not positive proof of zircon constituting the nuclei, as usually presumed. Some other minerals, e.g. xenotime, monazite, and apatite, may contain radioactive elements as well (cf. Hutton, 1947).

It has been found that rapakivi rocks are favorable for such studies. Joly (1923) already discovered, when studying pleochroic halos of various geological ages, that in the rapakivi from Viipuri most pleochroic halos appear to be derived from uranium. Other observations concerning pleochroic halos of Finnish rocks have not, as far as is known, been published. The purpose of the present paper is to study this matter further. Thin sections from the collections of the Geological Survey (the rapakivi area of Viipuri) and of the Geological Institute of the University of Helsinki (the areas of Laitila and Vehmaa) have been studied.

---

[1] Received April 15, 1955.

Rarely have multi-ringed halos, i.e. with several concentric rings, been found. In normal or typical rapakivi containing feldspar ovoids, halos regularly occur, at least to some extent, in biotite. The darkenings occurring in hornblende, on the other hand, are less distinct and in most cases diffuse around the edges. Such diffuse and irregular darkenings, which are likely to be met with in biotite as well, are distinguished in the present study from ordinary halos by being designated as pleochroic zones. Occasionally rosy red halos also occur in fluorite (e.g. Pyterlahti rapakivi). In biotite, however, the fluorite is surrounded by a pleochroic zone. In granites occurring in association with rapakivi, on the other hand, the number of halos proper, in particular, appears to vary considerably, whereas the number of zones does not vary quite so greatly.

Two thin sections of the same type of granite are likely to differ markedly in respect to the occurrence of halos; one may have them in great abundance while the other may have notably fewer. There exist, however, two types of granite in which halos occur invariably in considerable abundance. These porphyritic granites belonging to rapakivi rocks are the Sinkko granite in the Lappeenranta region (Hackman, 1935, pp. 24—25) and the Ytö granite in the Laitila rapakivi area; not only do halos occur in them in abundance, generally speaking, but many of the halos are multi-ringed, usually two-ringed (Plate I). This is unusual in other rapakivi rocks. Dr. A. Kahma (personal report), who has mapped the Laitila rapakivi area, describes the granite of Ytö as follows:

> The granite of Ytö is situated between the villages of Ytö, Kusni and Katinhäntä in the parish of Laitila. In the North-South direction, its breadth is greatest in the West, where it extends approximately 6.5 km, narrowing to the East and ending at the village of Katinhäntä (Fig. 1).
>
> The Ytö granite is gray, medium- or fine-grained, porphyritic, with sparse phenocrysts of orthoclase 2—6 cm in diameter and an abundance of small feldspar laths 1—2 cm long and 0.5 cm wide, which are in many cases subparallel. In the southern part the color near the contact is nearly always reddish, but in the northern part only in a narrow strip of the Ytö granite near the Ytö-Katinhäntä road. In the larger idiomorphic orthoclase phenocrysts, rings formed by inclusions are likely to be observed, which, in spite of the angular appearance of the crystals, are circular. Seldom does one meet with orthoclase ovoids characteristic of normal rapakivi.

A micrometric analysis of the Ytö granite made by the point counter method gave the following result: Quartz 31.3 %, potash feldspar 34.5 %, plagioclase (12—15 % An) 25.6 %, biotite 6.0 %, muscovite 1.2 %, fluorite 0.4 % and other accessories 1.0 %, total 100 %. Biotite, which from the standpoint of the present study is of special interest, occurs as flakes mainly measuring over 0.25 mm^2. Some smaller flakes are altered and greenish. In biotite there are numerous pleochroic halos. Some flakes as a whole are in a certain position quite opaque. It may be estimated that about one third of the biotite surface is covered by pleochroic darkenings.

In comparing the foregoing description with that given by Hackman of the Sinkko granite, we shall note that common to both is above all the gray color of the rock, which is rather exceptional in the rapakivi rocks. Reddish contact varieties are, however, common. Characteristically they are porphyritic, the phenocrysts being angular and of different sizes, though generally small. Of the dark minerals only biotite occurs in the Ytö granite, and it likewise constitutes the most important mafic mineral in the granite of Sinkko, although in places colorless diopside also occurs. Noteworthy is the fact that already in Hackman's report the abundance of pleochroic halos was remarked upon. In both granites, unlike all other rapakivi rocks, traces of deformation or alignment of minerals have been observed. They occur, hovewer, only weakly or locally.

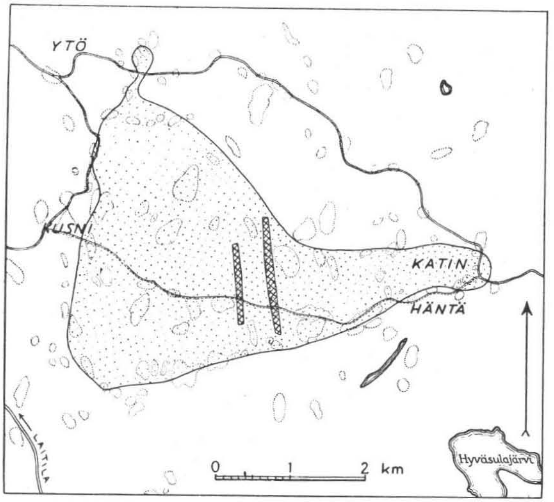


Fig. 1. The area of the Ytö granite (dotted) surrounded by normal rapakivi (left blank). (Olivine diabase dikes cross-hatched.) According to A. Kahma.

In the granite of Ytö the pleochroic halos are generally quite distinct and also sharply limited. All the pleochroic halos, and often also the zones, are equally broad, being either round or, more rarely, ellipsoidal, depending on the form and character of the mineral nucleus. The pleochroism of biotite is very marked (dark brown — light grayish brown or yellow). The halos are quite dark brown or opaque in the former position, while in the latter they are almost as light in color as the host biotite itself, though more brownish. It is to be noted that pleochroic darkenings are by no means to be seen around all the inclusions. Particularly in the larger halos the outer edge is commonly somewhat darker than the center. This is apparently due to the fact that ionization at various distances from the source of radiation gradually increases with the distance and has its maximum near the end of the alpha-particle range. The resulting ionization oxidizes ferrous iron to ferric iron in the host biotite and hence causes coloration (cf. Rankama, 1954).

In order to obtain some kind of statistical data, the number of different halos and darkenings was systematically counted in a certain thin section of the Ytö granite, the various pleochroic phenomena appearing in the field of vision being noted each time. In the count the halos and zones were distinguished. The former were further classified according to their size, as large and small. The limit was 30μ, as the radius corresponding to the outermost ring of the thorium and uranium types is appreciably larger. Fully developed halos of the type mentioned are accordingly to be classified, at least in maximum sections, among the large ones. These halos are taken into account separately. In addition, all the diffuse zones have been counted whose breadth is generally rather small ($<30\mu$). In all the foregoing, moreover, it has been noted whether there is a visible mineral nucleus or not. The results are to be seen in the accompanying table (Table 1). The numbers in brackets show how many have visible nuclei.

Table 1. Pleochroic darkenings in the biotite of the Ytö granite

Pleochroic halos			Diffuse zones	Total
Multi-ringed	Large	Small		
32 (2)	257 (49)	112 (13)	66 (16)	437 (80)

It is interesting to observe that in multi-ringed halos a visible nucleus occurs much more seldom than in other large halos. The explanation probably lies in the fact that, evidently owing to greater radio-activity, an excessively large nucleus easily causes overexposure or general darkening. The occurrence of smaller halos and diffuse darkenings around visible inclusions indicates that they cannot all be minor sections of larger halos. This would, in fact, be quite possible if we think of the matter three-dimensionally. Accordingly, the small halos would only represent sections in which the nucleus proper, whether visible or not, is not included in the thin section (Fig. 2). It is to be observed that many small halos, in the center of which no inclusion can be seen, are discs of quite even color and sometimes even with diffuse edges (comp. 4 a in Fig. 2), whereas in larger ones the halo often darkens appreciably toward the edge (1 a and 3 a in Fig. 2). This is because the radius of the outer rings is of the same magnitude as the thickness of the thin section. The lesser number of small halos with a visible nucleus as compared to the larger ones may be explained as resulting from this. In the zones, again, there are more of them with a nucleus than in any of the other groups.

The fact that the human eye is not able to judge correctly the extreme limit of the darkening makes accurate determinations difficult. The result depends greatly on the equipment used as well as the lighting conditions, and therefore the values given by various authors are not in agreement. The approximate radii of large halos (radius $\geq 30\mu$) have been measured by a micrometer eyepiece, but only for distinguishing the familiar uranium and thorium halos on the basis of the different ring radii. As shown in

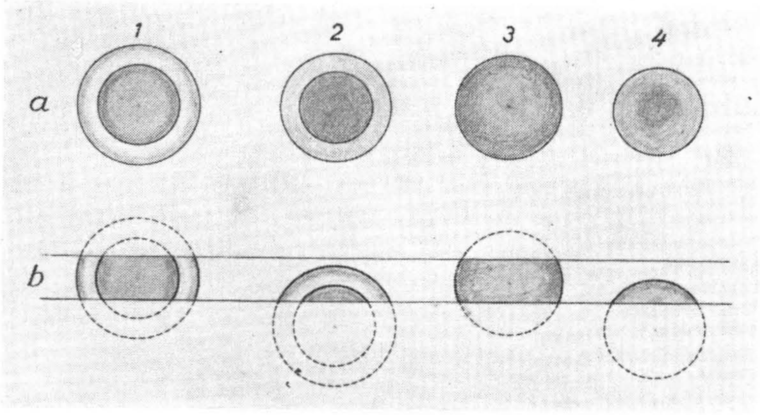


Fig. 2. Pleochroic halos in a thin section. Above (a) view in the thin section, below (b) cross section of the thin section. The outer borders of the rings have been especially drawn out, the broken line showing the course of the rings outside the thin section. 1 and 2, halos with a paler outside ring; 3 and 4, single-ringed overexposed halos.

Table 2, the difference of two outer ring radii in a thorium halo is 14μ, but only half of it in a uranium one. Also the ratio of these two rings is different in both types. Both of these criteria have been used in addition to the actual size of the rings. It has been found that most measured halos with a paler outside ring are definitely of the thorium type. In the dark central disc there very seldom occur distinct rings. The large single-ringed halos are overexposed. Their intensive darkening obliterates the inner structure. The radius of these halos varies greatly, being sometimes even

Table 2. The alpha-emitters of the uranium and thorium families with corresponding halo ring radii in biotite (Henderson and Sparks, 1939)

Uranium family		Thorium family	
Element	Ring radius	Element	Ring radius
RaC'	34.2 μ	ThC'	41.8 μ
An AcC	27.2	ThA	27.8
RaA	23.0	Tn ThC	23.9
Rn RaF	19.2	ThX RdTh	20.0
Ra Io UII	15.3	Th	12.4
UI	12.7		

slightly greater than that of the outer ring in a thorium type. Some of them have ring radii about 35μ and may well be derived from uranium. Uranium halos tend to be overexposed, which makes their accurate study impossible. Some halos, especially small ones with a visible nucleus, may be embryonic or of an unusual type, owing to the alpha-particles from the later members of the radioactive series (Henderson and Sparks, 1939; Henderson, 1939).

The minerals constituting the nuclei of different halos can not be determined in the thin sections but in most cases their general optics points to zircon. Particularly in small grains, however, distinguishing it from xenotime and monazite is not possible. In the heavy fraction of the same rock, the main mineral is zircon. It occurs as slender, clear and colorless prisms, sometimes containing minor inclusions. Some grains are zoned. It could be supposed that thorium present in zircon has given rise to the thorium halos, which generally are less common than the uranium type.

The study of the Sinkko granite gives a result similar to that presented in the foregoing. The two-ringed, full-sized halos are in general distinctly of the thorium type. Another special observation is that in some large, very dark halos there is near the nucleus mineral a narrow bleached zone (Photo 2 in Plate I).

Some Geiger-counter measurements were made in the field for investigating the radioactivity distribution in the area of the Sinkko granite and its vicinity. The determinations were made in road cuttings and exposures along the main road running through the area. The record of gamma counts is as follows: Four determinations made of the Sinkko granite in different localities gave a count range of 43—70 counts a minute and an average count of 55.5 c/m. As for the Lappee granite, over 1 km north of the Sinkko granite, three determinations gave a result of 31—44 c/m and an average of 38 c/m respectively. The radioactivity maximum is found near the southern contact. This difference in counts exceeds the possible observation errors. The relative amount of pleochroic halos may well be considered, at least in this case, to depend on the greater radioactivity of the rock.

Acknowledgements — The author is especially indebted to Prof. K. Rankama, who called his attention to the matter and subsequently helped the study along with much valuable advice. The author is also grateful to Prof. P. Eskola and Dr. A. Kahma for generously putting material at his disposal.

REFERENCES

HACKMAN, V. (1934) Das Rapakiwigebiet der Gegend von Lappeenranta (Willmanstrand). Bull. Comm. géol. Finlande 106.

HENDERSON, G. H. (1939) A quantitative study of pleochroic haloes V. The genesis of haloes. Proc. Roy. Soc. London, Ser. A, 173, p. 250.

—»— and SPARKS, F. W. (1939) A quantitative study of pleochroic haloes IV. New types of haloes. Proc. Roy. Soc. London, Ser. A, 173, p. 238.

HUTTON, C. OSBORNE (1947) The nuclei of pleochroic haloes. Am. J. Sci. 245, p. 154.

JOLY, J. (1923) Pleochroic haloes of various geological ages. Proc. Roy. Soc. London, Ser. A, 102, p. 682.

RANKAMA, KALERVO (1954) Isotope geology. London.

PLATE I.

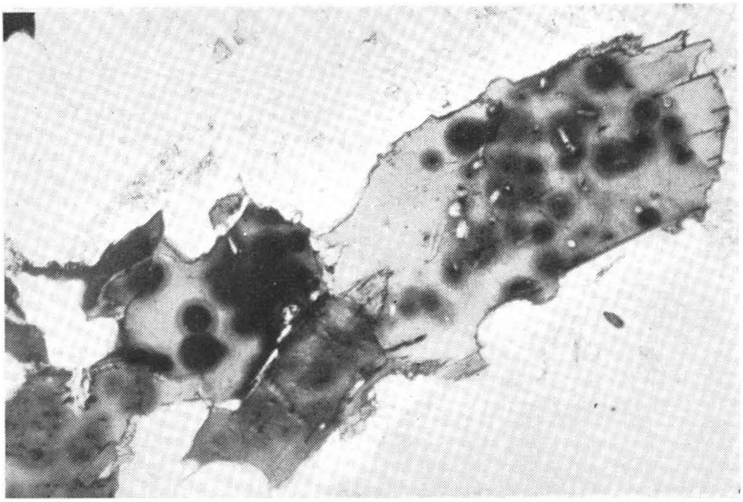


Photo 1. Pleochroic halos in the biotite of the Ytö granite. Nic. ||. Magn. 70 ×. Photo E. Halme.

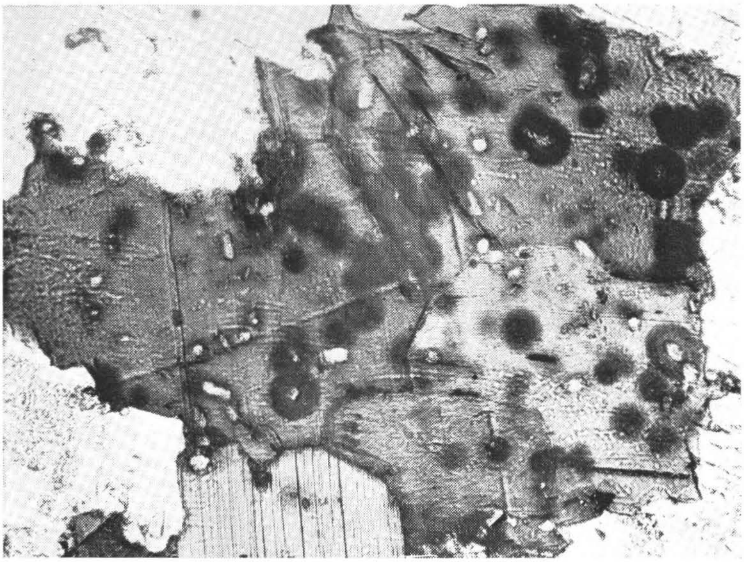


Photo 2. Pleochroic halos in the Sinkko granite, Karkola. N. B. The biotite flake in the middle at the bottom is in such a position that halos are hardly visible. Nic. ||. Magn. 70 ×. Photo E. Halme.

Simo Kaitaro: On pleochroic halos in some rapakivi rocks