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Lake Lappajärvi, a meteorite impact site in western Finland

by Martti Lehtinen

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Errata

p. 7 l. 29	for (1917a)	read (1971a)	
p. 26 l. 22	for harp-edged	read sharp-edged	
p. 44 l. 4	for Douple	read Double	
p. 51 l. 9	for 204	read $\bar{2}04$	
p. 74 Table 9a.	to be added:		
	P_2O_5	0.93	0.49	0.08	0.38	0.13
p. 79 l. 10	for $^{87}Sr/^{68}Sr$	read $^{87}Sr/^{86}Sr$	
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Jones, J.B. (1966) Order in alkali feldspars. Nature 210, 1352-1353.

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LAKE LAPPAJÄRVI, A METEORITE IMPACT SITE IN WESTERN FINLAND

BY

MARTTI LEHTINEN

WITH 23 FIGURES AND 11 TABLES

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Lappajärvi (lat. 63°09′ N, long. 23°42′ E) is a roughly elliptical lake covering an area of 160 km^2. The surrounding bedrock consists of Precambrian granite, granodiorite, granite pegmatites and mica gneisses. The lake occupies an area of low magnetic relief, and the preliminary gravity survey indicates an almost circular (diameter 17 km) Bouguer anomaly of —10 mgal.

A series of rocks of shock-metamorphic origin is present: impact breccia (a breccia of bedrock and mineral fragments; shock effects from weak to moderate), suevite (a mixture of glassy bodies with fluidal texture, and rock and mineral fragments at all stages of shock metamorphism), and kärnäite (a breccia with lava-like groundmass). These rocks are the most important constituents of the glacial drift within the area, but only kärnäite crops out. A detailed petrographic description of each rock type is given.

X-ray, infrared absorption and optical properties of the minerals, especially of coesite, alkali feldspars and quartz, are presented and discussed in detail. Orientations of planar elements in quartz, feldspars and several accessory minerals are examined.

Chemical analyses and Co-Ni determinations of the shock-metamorphosed rocks and the whole-rock Rb/Sr isotope ratios of a kärnäite specimen are presented and discussed.

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INTRODUCTION

The oldest geological observations in the Lake Lappajärvi area were made during the general geological mapping, 1: 400 000, in the 1910's (Berghell, 1921; Saksela, 1934, 1935; Laitakari, 1942), but already in 1858 H. J. Holmberg mentioned the lava-like rock (porphyry) that crops out near the center of the lake. This rock was named kärnäite by Dr. Hugo Berghell of the Geological Survey of Finland. Later, Eskola (1921, pp. 11—12, 1927), Kaikko (1921?), Laitakari (1942, pp. 38—40), and Saksela (1949, pp. 22—23) added to the knowledge of the petrological and chemical properties of the kärnäite. They regarded the kärnäite as a volcanic rock consisting mainly of agglomerate and tuff with a large amount of bedrock fragments.

The possibility that Lake Lappajärvi might be produced by a meteorite (asteroid) impact was first suggested by Professor Frans E. Wickman of Stockholm in the 1950's and by Professor Wolf von Engelhardt of Tübingen in the spring of 1967 (Th. G. Sahama, personal communication, 1967). Svensson (1968, 1971) published mineralogical and morphological evidence in favor of meteorite impact origin.

In order to discover signs of shock metamorphism in the Lappajärvi rocks, the late Professor Vladi Marmo, Professor Th. G. Sahama, and the author visited the area in July, 1967. This preliminary survey did not produce promising results, but in August, 1967, the author found numerous boulders and cobbles of shock-metamorphosed rocks in gravel pits at Hietakangas, about 5 km southeast of the lake. Most of the material for the present investigation was, however, collected in the summer of 1968. Preliminary results of laboratory investigations on this material were published in 1970 by the author.

The purpose of the present investigation is (1) to give petrological, mineralogical and chemical data concerning the Lappajärvi rocks, (2) to prove their shock-metamorphic origin, and (3) to give a brief review of the principles and nomenclature of shock metamorphism, a new type of metamorphism with which many geologists may be rather unfamiliar.

SHOCK METAMORPHISM: A GENERAL ACCOUNT

During the past ten or fifteen years the conception of hypervelocity meteorite impacts producing large circular features on the Earth's surface has grown from

a speculation into a conclusive theory explaining many formerly unsolved "cryptovolcanic" or "cryptoexplosion" structures. Owing partly to the growth in knowledge of the craters and rocks of the moon, this theory has now become one of the most fruitful and popular in the earth sciences and has produced a wealth of papers, and reviews, and also a number of conferences. This does not mean, however, that it has been proved that all enigmatic circular features are of extraterrestrial origin, and even though the number of known and suspected meteorite impact craters and astroblemes (Dietz, 1961) has grown rapidly since 1960, only about 50 to 60 of them are considered to be of irrefutable impact origin (French, 1968, p. 5).

The evidence for meteorite impact as a geologic agent has accumulated as a result of progress in various sciences. Investigations dealing with shock waves, cratering processes, high-pressure phases, underground detonations of nuclear and chemical devices, and detailed geologic and mineralogical studies of known and suspected meteorite impact sites, etc. have revealed changes in, and transformations of, minerals and rocks, which are beyond the scale of any type of "normal" metamorphism or geologic agencies. Thus a new term, shock metamorphism or impact metamorphism, was introduced (Pecora, 1960, p. 19). The two terms are closely related, and they both describe changes produced in minerals and rocks by transient, high-pressure shock waves. The term, impact metamorphism, however, is restricted to the observed changes produced by the hypervelocity impact of a body such as a meteorite (Chao, 1967b, p. 192). The evidence and criteria for shock metamorphism have been discussed in their geologic aspects in many reviews, e.g. those by Beals *et al.* (1956, 1963); Dence (1964); Stöffler (1966, 1971a, 1972); Chao (1967a, 1967b, 1968); French (1968); and Short (1968).

Not all geologists, however, accept meteorite impacts as a significant geologic agent. Arguments against the meteorite impact theory are mostly based (1) on the observation that some supposed meteorite impact craters are associated with tectonic or volcanic lineaments (Bucher, 1963; McCall, 1964), (2) on geochemical anomalies in melted rocks and breccias of some Canadian craters (Currie and Shafiqullah, 1967, 1968; Currie, 1971), and (3) on the assumed existence of a type of explosive volcanic mechanism — not yet discovered within any active volcano or volcanic area — called cryptovolcanism (Bucher, 1963; McCall, 1968; Currie, 1971, p. 5584). Cryptovolcanism is usually regarded to be related to gaseous or certain carbonatite explosions. Bucher (1963), Dietz (1963) and French (1968) have discussed in detail the arguments for and against the meteorite impact theory.

Stöffler (1972, 1974) has written two comprehensive papers treating the behavior of minerals under shock compression and physical properties of shocked minerals; he distinguishes four main types of residual shock effects:

1. Fracturing

2. Plastic deformations
   — planar fractures
   — planar elements (shock lamellae, planar features)
   — deformation bands
   — irregular plastic lattice deformation

3. Solid state transformations
   — high-pressure phases
   — short-range-order phases

4. Thermally induced transformations
   — decomposition
   — melting and vaporization.

Stöffler also gives detailed information about the shock effects, and about the optical and other physical properties of shocked minerals. In addition, he explains the basic properties of shock waves and describes experimental shock-wave techniques.

In rocks of shock-metamorphic origin, the shock effects have been developed and distributed very irregularly not only because of the mixing of strongly and weakly shocked or "unshocked" material during the cratering process but also because in a single rock specimen each mineral grain has its own shock history. When shock-metamorphosed rocks are classified, the strongest shock effects observed in a specimen usually determine the shock stage to which the specimen belongs. Stöffler (1965, 1966, 1971a) has elaborated six stages or zones (facies) of progressive shock metamorphism. Utilizing the Hugoniot data for quartz and feldspars he fitted the three lowest shock stages to correspond to the three regimes of the generalized Hugoniot curve of the framework silicates. Quartz and feldspars are suitable index minerals because their lattices, having low packing densities, are much more readily deformed, broken or transformed than are the lattices of, say, olivine, pyroxenes or amphiboles. Stöffler's (1917a) six shock stages are:

Shock stage 0. The low-pressure regime represents a Hugoniot of shocked material with its initial zero-pressure density. Quartz and feldspars shocked to this stage are only fractured and indicate no change of density or other physical properties.

Shock stage 1. This stage is characterized by diaplectic quartz and feldspars. i.e. quartz and feldspars with planar elements (planar features). Canadian investigators (Dence, 1968, p. 175; see also Robertson *et al.,* 1968, p. 438) have used four shock zones corresponding to Stöffler's shock stage 1. Each of these zones is characterized by the development of a defined new type of planar feature of

quartz, and thus the number and intensity of sets of planar features per grain gradually increases.

Shock stage 2. The most typical feature of this stage is the appearance of amorphous short-range-order phases of the quartz and feldspars called diaplectic glasses (von Engelhardt *et al.,* 1967, p. 93) or thetomorphic glasses (Chao, 1967a, p. 212). Another feature typical of this stage is the occurrence of high-pressure polymorphs of silica, viz. coesite and stishovite.

Stöffler (1971a) derived the following three highest shock stages from different effects of high residual temperatures.

Shock stage 3. In this stage, selective melting of feldspars takes place, and vesiculated feldspar glasses are produced. Stöffler used the melting of any feldspar of the rock to define the lower limit of this stage. The diapletic quartz glass may still contain coesite aggregates and traces of stishovite. The original texture may be only indistinctly visible.

Shock stage 4. Stage 4 is characterized by the melting of the whole rock. As a consequence of the short melting and quenching times very heterogeneous glasses are produced and are generally mixed with material of lower stages of shock metamorphism. In large craters, however, slow cooling beneath fall-back debris may allow the formation of a relatively homogeneous glass with ghost-like rock fragments.

Shock stage 5. Pressures of several megabars and temperatures exceeding 2 500°C to 3 000°C may be reached near the point of impact. This results in the total vaporization of both the meteorite projectile and the target rock. Small glassy or metallic spherules form as condensation products.

Stöffler's classification of progressive shock metamorphism is appropriate for nonporous, quartz-feldspar rocks of broad compositional variance. There is still a dearth of up-to-date knowledge of whole-rock shock data and residual temperatures, etc. This, as well as the complexity of shock metamorphism itself, prevents the drawing up of detailed classifications appropriate for all craters (see Chao, 1967a, 1968). But careful investigation of each meteorite impact crater, and research into shock metamorphism will permit further local subdivisions.

In addition to the great number of mineralogical and petrological criteria for shock or impact metamorphism there are also some structural criteria. The best known criterion of this type is the occurrence of shatter cones (Dietz, 1959, 1961, 1968). Shatter cones are conical rock fragments varying in length from less than 2 mm to some meters. The apical angles of the cones are usually close to 90°. The surface of a shatter cone is characterized by striations that radiate from the apex in horsetail fashion. The best-developed cones occur in fine-grained, homogeneous rocks. In hard, coarse-grained rocks, complete cones are rare but fracture surfaces with typical convergent striations of shatter cones are observed (E.C.T. Chao, 1971, personal communication).

Recent investigations at Sierra Madera, U.S.A., (Howard and Offield, 1968) and Gosses Bluff, Australia, (Milton *et al.,* 1972) have revealed that after the disturbed structural blocks have been set to their pre-event positions, most cones point inward and upward, a pattern favoring an extraterrestrial origin for the explosion. Robertson (1968) reported abundant shatter cones at La Malbaie structure in Canada, where the best conditions for shatter-coning existed at a distance of approximately 7 km outward from the center. This suggests that coning is a relatively low-grade shock-metamorphic feature. Shatter cones probably form during the compressive stage of the shock wave (elastic precursor). Although there is some uncertainty as to whether shatter cones are exclusively of shock-metamorphic origin, there is no doubt that they are associated with shock-metamorphosed rocks and may serve as significant megascopic indicators of an eroded meteorite impact site.

LAKE LAPPAJÄRVI AREA

Location, size and topography

The Lappajärvi crater lake is located at latitude 63°09′N and longitude 23°42′E about 100 km east of Vaasa, a town in western Finland (Fig. 1). The center and church of the commune of Lappajärvi are situated on the north-

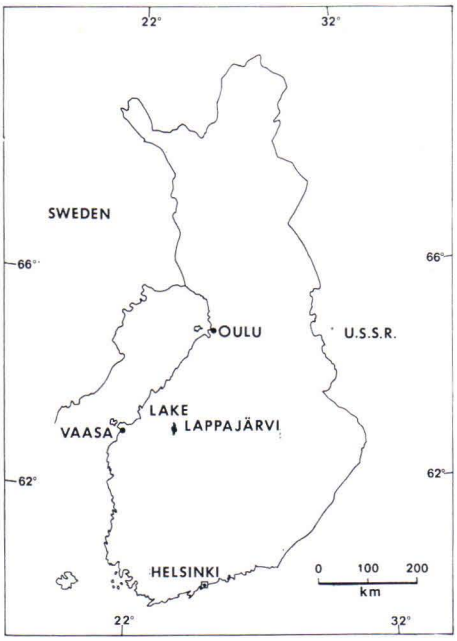

Fig. 1. Location map.

western shore of the lake. The roughly elliptical lake covers an area of about 160 km^2 (Odenwall, 1934) and is approximately 23 km long and 12 km across, being elongated in a north-south direction.

Lake Lappajärvi is of preglacial origin, but its exact age is still unknown. The breccia material of the original crater fill and the fractured and brecciated bedrock surrounding the crater had a low resistance to erosion, which accounts for the strong erosion during preglacial times and the Pleistocene glaciation. The present level of erosion of the crater is unknown. However, especially during the Pleistocene glaciation, new material deposited in the crater, principally in its northwestern corner, and altered the form of, or even "transported", the crater. According to Mölder (1948), the strongest striations in the area are N 18°—30° W. The author's measurements from a kärnäite outcrop on the eastern shoreline of the island Kärnänsaari revealed striations N 30° W, N 50°—60° W and N 76° W, of which the first is the youngest and strongest. The center of the present-day lake does not necessarily coincide with that of the original crater, and it does in fact seem to have been moved somewhat toward the southeast.

Svensson (1968, p. 438) wrote that "the size of the structure is not exactly known but must be of the order 5—6 km". More recently Svensson (1971, p. 5384) has assumed that "the impact-melt layer in the Lappajärvi structure is roughly circular and has a diameter of 5—6 km". The present author has carried out outcrop and boulder observations and calculations of the approximate volume of the impact-produced melt and breccias (see p. 29) which indicate that the diameter of the original crater was about 12 km to 14 km. According to recent gravity investigation (see p. 15), the gravity anomaly has a diameter of 17 km.

The mean depth of the lake is 7.45 m (Odenwall, 1934). The conspicuous features of the bottom topography are the two slightly inward-curved parallel bottom valleys and the shallower depression connecting the valleys and dividing the lake bottom into two shallow platforms (Fig. 2). The parallel valleys occur near the western and eastern shorelines of the lake. The maximum depth of the western bottom valley is 38 m and of the eastern one 36 m (Odenwall, 1934).

The eastern bottom valley seems to be associated with a great lineament running in a NNW—SSE direction, which comprises not only the eastern bottom valley, but also the river Ähtävänjoki (the outlet of Lappajärvi) in the north, and the lakes Kaartusenjärvi, Alajärvi and Ähtärinjärvi in the south. It has been concluded that the eastern bottom valley is of tectonic origin, but a similar origin is not so clear for the western and middle bottom valleys. According to Svensson (1971, p. 5385), the western and middle bottom valleys do not fit into the general lineament pattern of the area, and he favors a meteoritic origin for the valleys. In any case, Lake Lappajärvi is a deeply eroded crater with a layer of

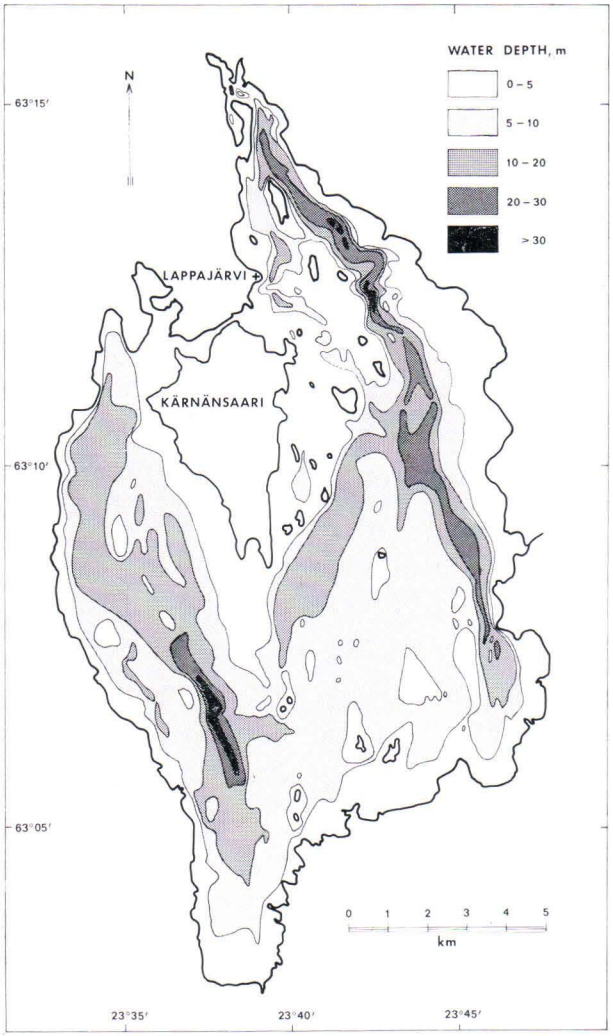

Fig. 2. Bathymetric chart of Lake Lappajärvi. Redrawn according to Odenwall (1934).

melted rock (kärnäite) in the center. This kärnäite might lie close to the bottom of the original crater. Reference is made to the Brent Crater in Canada (Dence, 1968) and to Lake Mien in Sweden (Stanfors, 1969, 1973). Kärnäite is highly resistant to mechanical disintegration, and thus the brecciated and fractured zones surrounding the kärnäite layer (or group of small lenses) have been eroded more rapidly and deeply, forming the troughs around the kärnäite. Possible shear zones and fractures connected with readily eroded rocks might have facilitated the process.

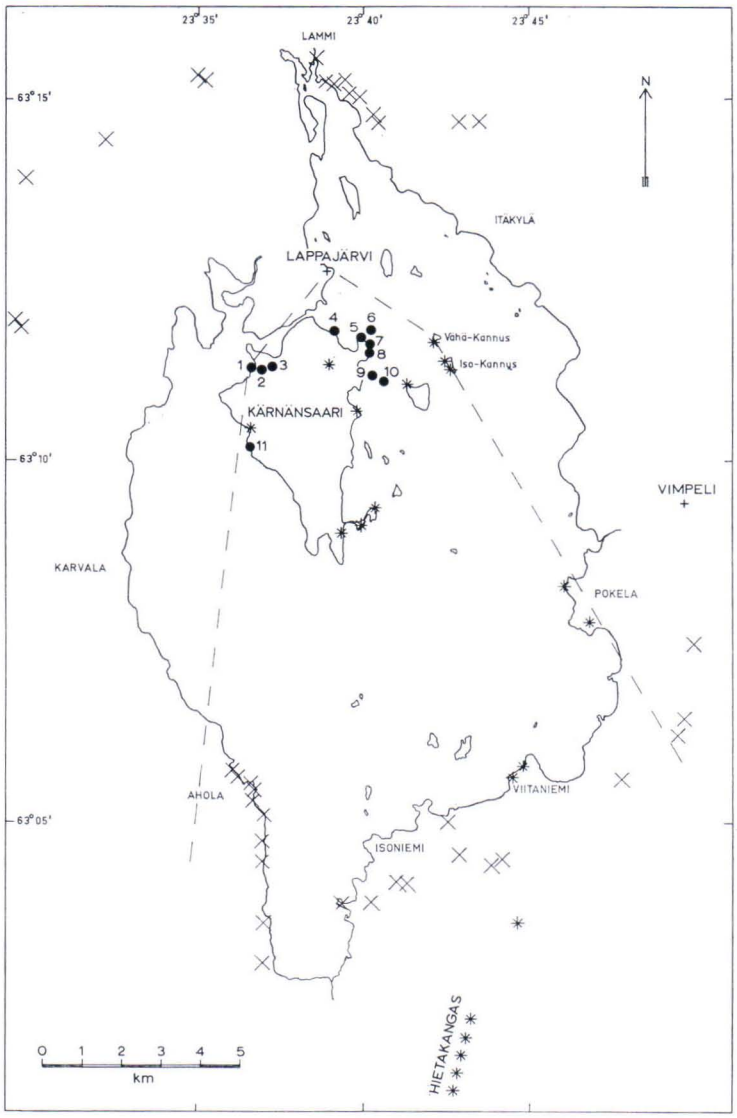

Fig. 3. Kärnäite and bedrock outcrops, breccia localities, and indicator fan of kärnäite.
Dots: Nos. 1—10, kärnäite outcrops. No. 11, a possible kärnäite outcrop. *Stars:* Most important breccia (impact breccia, suevite) localities. *Crosses:* Bedrock outcrops. *Dashed line:* Indicator fan of kärnäite.

All the largest islands in the lake, including Kärnänsaari, are situated on the northern bottom platform. The islands are very low, with the exception of the

northern part of Kärnänsaari, where some mounds of gravel and moraine lie 15 m to 20 m above the lake surface.

As a whole, the topography of the Lake Lappajärvi area is rather flat, but some hills southeast of the lake rise up some 100 m above the lake surface, which is 69.4 m above sea level. The 80 m contour forms a rough circle around the lake area, excluding the southern and northern corners of the lake. The circle goes through the Lappajärvi, Karvala, Ahola, Isoniemi, Viitaniemi, Vimpeli and Itäkylä villages (Fig. 3) and has a diameter of about 14 km. Its center lies near the southern end of Kärnänsaari, and there only kärnäite crops out within the circle.

Geology of the environment

According to the old General Geologic Map of Finland (1:400 000) Sheet B 3, Vaasa (Saksela, 1934), Lake Lappajärvi is situated in the Precambrian supracrustal rock area of southern Pohjanmaa, where the general strike of the schists is N 50°—60° E but with considerable local variations. Mica gneisses (biotite-plagioclase gneisses; Saksela, 1934, 1935; Laitakari, 1942) are the main constituents of the schists, whereas amphibolites and hornblende gneisses, quartzites, leptites and limestones occur only sporadically (Saksela, 1934, 1935, pp. 6—13; Laitakari, 1942, pp. 11—25).

The plutonic rocks of the area are made up of granite pegmatites and muscovite granites and of equigranular or porphyritic gneiss granites (Saksela, 1934). These rocks form migmatites with the schists or contain abundant schist inclusions. The chemical composition of the gneiss granites is usually that of granodiorite or quartz diorite (Saksela, 1935, p. 13).

There are few outcrops on the shores or in the vicinity of Lappajärvi. Those at the northern corner of the lake, between the villages of Lammi and Itäkylä and west of Lammi (Fig. 3), consist of muscovite granite, granite pegmatites and mica gneisses. The granite pegmatites sometimes contain elongated mica gneiss inclusions. The muscovite granite is a medium- or coarse-grained and massive rock with perthitic microcline (average triclinicity of three samples, 0.93), albite-oligoclase, quartz, muscovite and biotite as chief minerals. Apatite and zircon are common accessory minerals.

There are some outcrops of granite pegmatite and mica gneiss on the western shore of the lake, in the village of Ahola where granite pegmatites occur as large lenticular bodies in mica gneiss. Owing to the pegmatite bodies the strike of the mica gneiss varies greatly. The northernmost outcrop on the lakeside at Ahola is a migmatite consisting of mica gneiss and granodiorite; the strike is N 14° E and the dip 80° W.

The mica gneiss is a medium-grained, granoblastic rock, the chief minerals of which are quartz, biotite, oligoclase, muscovite and potassium feldspar. Apatite

is the most common accessory mineral. With a decreasing amount of biotite, the amount of potassium feldspar and quartz normally increases. The varieties rich in biotite are coarser-grained and strongly folded.

Oligoclase, which to the naked eye is dull and white or reddish, quartz with wavy extinction, muscovite and microcline perthite (triclinicity 0.91—1.0) are the chief constituents of the granite pegmatite. The amount of microcline varies. In some specimens collected from the outcrops on the western shore of the lake, oligoclase is the principal feldspar. Biotite, green apatite and graphite occur as accessory minerals. Graphite sometimes forms abundant twisted plates in biotite.

The southeastern corner of the Lake Lappajärvi area contains abundant outcrops composed of mica gneiss, granite pegmatite and granodiorite. The granodiorite is a gray, medium-grained rock with well-developed gneissoid structure. The chief minerals are oligoclase-andesine, quartz with wavy extinction, biotite, ragged green hornblende and microcline. Epidote, apatite and opaque minerals are accessory minerals.

Previous investigations (Saksela, 1934, 1935; Laitakari, 1942) and the present author's observations of the outcrops and boulders on the shore and in the vicinity of the lake indicate that the bedrock immediately adjacent to or underlying the original crater and the present lake area consists, or consisted, of granite pegmatite, mica gneiss often migmatized by granite and granite pegmatite, and some muscovite granite and granodiorite. This conclusion is consistent with the observation that all rock fragments, blocks and inclusions observed in the breccias and kärnäite consist of the forementioned rocks.

Geophysical investigations

The formation of an astrobleme by meteorite impact and the enormous explosion that ensues will produce marked changes in the physical properties of the crater fill and the adjacent bedrock (Millman *et al.,* 1960; Beals *et. al.,* 1960). Canadian scientists have carried out systematic gravity, seismic and magnetic surveys at craters of explosive origin in Canada. The gravity surveys in particular, although they do not alone yield unique solutions, are important and illustrative in that they impose certain constraints on subsurface configurations (see also Milton *et al.,* 1972; Fudali and Cassidy, 1972; Fudali, 1973; Fudali *et al.,* 1973).

A typical Bouguer anomaly map of an astrobleme or cryptoexplosion structure appears to have rather circular gravity contours (negative residual anomaly) with a minimum of several milligals at the center; for instance, the Brent (Millman *et al.,* 1960), West Hawk Lake (Halliday and Griffin, 1963), and Deep Bay craters (Innes *et al.,* 1964). The gravity pattern, however, depends on the size and age of the crater, crater fill and surrounding bedrock.

Innes (1961) used gravity data to calculate mass deficiency and hence the amount of deformed rock in the crater. His results agree well with the depth/diameter ratios of Baldwin (1963, Fig. 19) for meteorite impact craters and with the computed crater model of Rottenberg (Beals *et al.,* 1960, pp. 5—8).

The gravity data for the Lappajärvi crater are so far restricted to information on the general gravity map of Finland (Bouguer anomalies). The lake basin occupies a negative gravity anomaly of 5 mgal to 10 mgal. The preliminary results, however, of a recent gravity survey carried out in the area by the Geological Survey of Finland (S. Elo and F. Pipping, 1975, personal communication), have revealed the existence of a circular negative gravity anomaly of about 10 mgal. The diameter of this anomaly is about 17 km, and its center lies 2 km south of point No. 10 in Fig. 3.

No drill holes have been made, and consequently accurate knowledge of the crater fill is lacking. The crater is deeply eroded, and it is unlikely that there

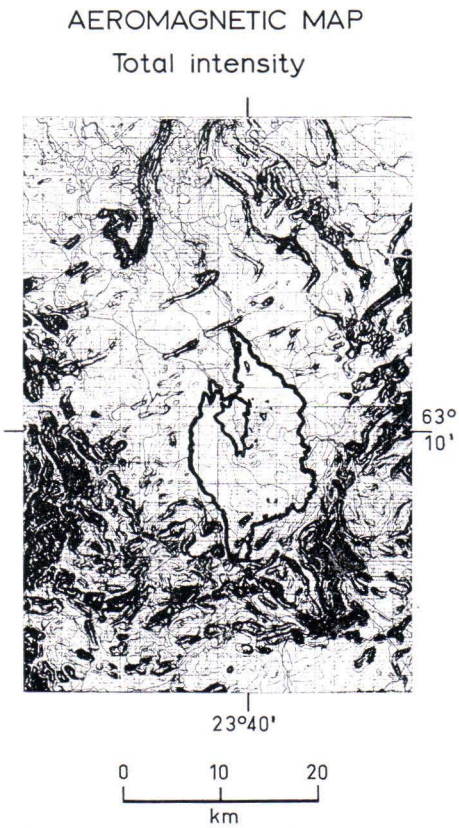

Fig. 4. Aeromagnetic total intensity map.

is an unusually large amount of low-density sedimentary material of later origin. The specific gravity of kärnäite is somewhat lower (2.52—2.58; the mean of four specimens is 2.56) than that of the adjacent bedrock. The specific gravity measured for the bedrock specimens collected varies from 2.61 (granite) to 2.82 (mica gneiss), the mean value being about 2.68 to 2.70. The negative gravity anomaly may be due in part to the difference in the specific gravity between kärnäite and the adjacent bedrock. It is, however, the porosity of the breccias, which are partly mixed with or underlying kärnäite and which have very low specific gravities from about 2.55 down to about 2.25 or 2.20, that offers the most ready explanation for the gravity anomaly. This model, proposed here for mutual positions of the shock-metamorphosed rocks in the crater, resembles that suggested for Lake Mien in Sweden, where diamond drilling has revealed that suevite-like breccia, tuff-like breccia and brecciated bedrock underlie the lava-like rock (Stanfors, 1969, 1973).

The total-intensity magnetic map of the Lake Lappajärvi area is based on an aeromagnetic survey carried out by the Geological Survey of Finland. The most outstanding feature on the map (Sheets 2313 and 2314) is the marked contrast between the intensities observed within and beyond the lake area (Fig. 4). The surrounding bedrock produces strong elongated magnetic anomalies in the direction N 50°—60° E and N 30° W. The former is parallel to the general strike of the schists of the southern part of the area, and the latter is parallel to the strike of the schists of the northern part. But within the lake area proper, the variation of magnetic intensity is rather small and uniform except at the northern and southern ends of the lake. These parts, however, to judge by observations on outcrops, for instance, lie outside the crater proper. There are some areas of low magnetic relief also in the surroundings, but they seem to associate with the known massifs of granite or granodiorite.

ROCKS OF SHOCK-METAMORPHIC ORIGIN

Three types of rocks of shock-metamorphic origin occur in the Lake Lappajärvi area: (1) impact breccia, (2) suevite, and (3) kärnäite (impact lava). The terminology of shock- and impact-metamorphosed rocks, and their classification in general, is so far tentative. Most of the terms used have been derived directly from experimental crater studies or studies of young craters and their crater fill. These terms, which are usually defined on the basis of the location of the rock unit with respect to the crater, are not very suitable for old, deeply eroded and preglacial craters like Lappajärvi. The names adopted here for the Lappajärvi rocks of shock-metamorphic origin are thus used as descriptive terms without any reference to the present or original location of the rocks.

Kärnäite is the only one of the three rocks that crops out, but all the rocks and their constituents form the main part of the glacial drift within the lake basin and south or southeast of the lake. In particular, the till near the southeastern shore (Viitaniemi) and the eskers south of the lake are rich in breccia (impact breccia and suevite). The gravel pits of Hietakangas (Fig. 3) contain abundant rounded cobbles and pebbles of breccias and kärnäite, making up about 60 % to 70 % of the esker material, breccias being more common than kärnäite. Some large kärnäite and breccia boulders have also been found on the shores of the islands. There are many large boulders of suevite on the western shore of the island Iso-Kannus but none on the eastern shore; some boulders have been found in the southern corner of the island Vähä-Kannus (Fig. 3).

The kärnäite seems to be very resistant to weathering, whereas the breccias, especially the suevite, disintegrate readily. It usually only takes a few years for even a large boulder of suevite to be transformed into a clayey material containing some rock and glass fragments, which is why almost all boulders observed on the shores consist of kärnäite. Exceptions are some suevite boulders rich in large twisted glass fragments, which occur on Iso-Kannus and Vähä-Kannus.

The occurrence of kärnäite as clasts in the glacial drift has long been investigated, starting in 1914 when Laitakari (1942, p. 40) mapped the distribution of kärnäite in the area. Since then, Mölder (1948), Saksela (1948) and Kulonpalo (1969) have augmented the information about the occurrence of kärnäite in glacial drift southeast of the lake. The present author has found more than twenty clasts of kärnäite in eskers and moraine about 100 km to 120 km southeast of Lake Lappajärvi. One of these clasts, a boulder, weighed more than 5 kg. According to Kulonpalo (1969), one kärnäite block has been discovered at a distance of 300 km southeast of the lake.

The rapid weathering and breakup of suevite clasts makes it very difficult to study the distribution of the suevite. In the lake area itself, however, breccia material prevails on the eastern side and kärnäite material on the western side.

Petrography

Impact breccia

A multiple rock type breccia, consisting of angular mineral and bedrock fragments of different sizes and showing typical features of low stage shock metamorphism, is called impact breccia (see Lehtinen, 1970). This type of breccia does not contain melted material but has a matrix of zeolites (mostly heulandite), cristobalite, and some argillaceous substance. Some impact breccia specimens, however, display a few glass fragments with fluidal texture indicating that the breccia passes gradually into the suevite. In the lake area and in the

glacial drift, cobbles and boulders of polymictic impact breccia are uncommon, whereas shock-metamorphosed cobbles of bedrock occur frequently. Fine-grained impact breccia material often found in cracks or hollows of these cobbles indicates that the cobbles have been embedded in a fine- or medium-grained breccia derived from a larger breccia unit.

The impact breccia consisting of small fragments is usually friable, and small specimens disintegrate readily when handled. The mineral fragments consist mostly of quartz with or without planar deformation structures, feldspars with different shock-metamorphic features including vesiculated feldspars, and of biotite with or without kink bands. Some biotite flakes have been partially decomposed. Graphite occurs as kinked flakes in some specimens.

Most of the rock fragments and cobbles consist of coarse- and medium-grained granite (granite pegmatite), but fragments of mica gneiss and granodiorite, and cobbles and fragments of migmatite (mica gneiss and granite) are also common. Fragments of mica gneiss in the impact breccia are usually more altered than those of granite or granodiorite. They are grayish green in color as a result of strong chloritization of the plagioclase. The biotite has been chloritized and decomposed. The differences in the mineral content, grain size, chemical composition and structure, etc. of the granite and mica gneiss fragments have produced different shock-metamorphic effects and consequently differences in secondary alteration.

Small fragments of shatter cones are found in several weakly shocked mica gneiss and granite inclusions in the impact breccia. Especially in granite the shatter cone fragments occur only as curved surfaces with convergent striations and cannot be seen unless the rock is struck with a hammer. In a single rock inclusion the shatter cones appear to have a common axial orientation but in a mica gneiss inclusion two cones intersect almost at right angles. The largest fragment of shatter cone, about 4 cm long and with a large apical angle, occurs in a mica gneiss inclusion.

Suevite

Suevite is the most common type of shock-metamorphic rock in the glacial drift in the lake basin and in the southeastern neighborhood of the lake. Almost all specimens of this rock are very friable and weather readily, which explains why it has not been previously discovered. Suevite is, however, the most important and noteworthy type of rock examined, and exhibits, among other things, the whole scale of shock-induced features, i.e. plastic deformations, and solid-state and high-temperature transformations (see Stöffler, 1972, 1974).

The term suevite — originally a local name for an unusual breccia consisting of glass bombs and crystalline rock fragments from the Ries Crater in Germany — has been used for certain strongly shocked impact rocks in general, i.e. for

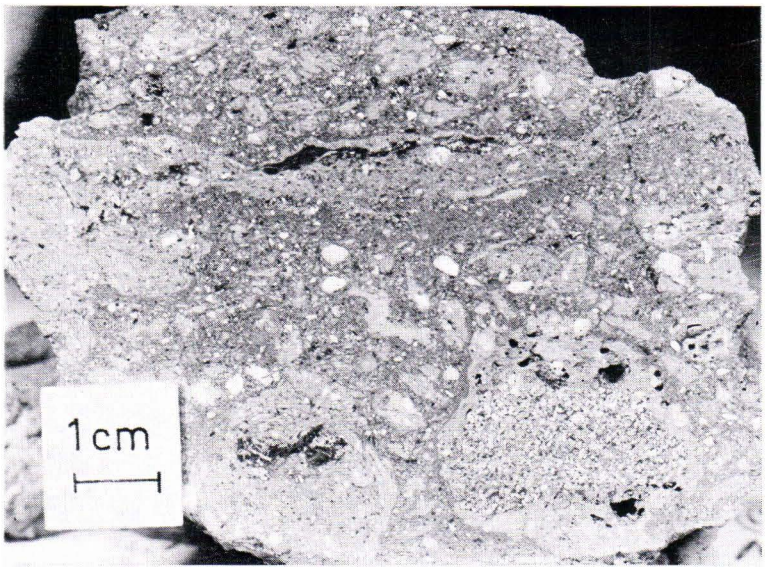

Fig. 5. Cross-section of a suevite specimen showing two mica gneiss (migmatite) fragments coated by vesicular glass with fluidal texture (lower part), and a long very heterogeneous glass body resembling a bipolar (or unipolar) fusiform bomb (upper part). Specimen No. 117.

the uppermost layer of fallback and fallout breccia (v. Engelhardt, 1971, p. 5569). The corresponding Lappajärvi rock was called fallout breccia by the author (Lehtinen, 1970). However, in the Lappajärvi region it is impossible to distinguish between actual fallback and fallout breccias because they do not crop out and occur only in the glacial drift. Thus, the name suevite has been adopted for the Lappajärvi rock, a usage consistent with the definition of suevite in the Glossary of Geology (Gary *et al.,* 1972).

The color of the Lappajärvi suevite varies from mottled brown to gray, and it is lighter in color than kärnäite. The bedrock and mineral fragments are the same as those occurring in the impact breccia and in kärnäite. Estimates made by examining the suevite boulders and cobbles in the gravel pits of the esker of Hietakangas show that half of the rock fragments having a diameter greater than 1 cm consist of mica gneiss and the rest of granite (granite pegmatite), granodiorite and migmatite (mica gneiss and granite). Mineral fragments derived from granite and granite pegmatite (quartz, feldspars, biotite) dominate among the smaller fragments, although small fragments of mica gneiss are also common. The suevite is much more heterogeneous than is the kärnäite and thus its rock and mineral fragments exhibit a wide range of stages of shock metamorphism. Most fragments belong to Shock Stage III (p. 31), but fragments belonging to Shock Stages I and II are also common. It is difficult to find large fragments

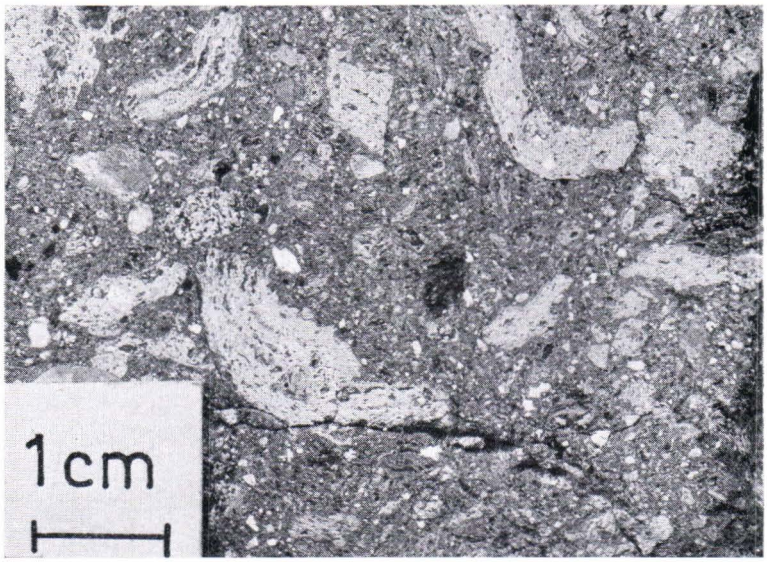

Fig. 6. Twisted and bent glass bodies with fluidal texture resembling ribbon lapilli and almond-shaped lapilli, some of which have fibrous ends. Suevite, Specimen No. 94.

without visible shock effects. Two kinds of fragments are distinguished: uncoated fragments and fragments coated by glass with fluidal texture (Fig. 5).

When struck with a hammer, some of the suevite hand specimens break into pieces revealing fragments resembling volcanic bombs and lapilli. The bombs, like their negatives in the rock, have a relatively smooth surface and are dark brown due to a coating of iron oxide. Many bombs have a form typical of fusiform or almond-shaped bombs (Macdonald, 1972, p. 125). Cross sections of the bombs, however, show that they are oblong rock fragments (usually mica gneiss of Shock Stage III) coated by vesicular glass with fluidal texture. The coat is some millimeters thick, but at the stretched ends of the bomb it may be more than 2 cm in thickness. The shape of the coat suggests that the fragments were surrounded by the melt during their passage through the air.

The most conspicuous features of the Lappajärvi suevite, however, are the glass fragments with fluidal texture, in which numerous small mineral and rock inclusions and vesicles may also be present. They are usually of lapilli size, but both their shape and size are very variable. Because of the extreme heterogenity of the breccias, it is difficult to give accurate percentages of the constituents, but the amount of glass most often varies from 20 % to 50 %. The specimens rich in glass also contain the largest glass bodies as irregular, flat and twisted bombs ("pancakes"). The largest body recognized as a single bomb has approxi-

Fig. 7. Almost unrecrystallized brown glass with schlieren, and with partly melted and angular mineral fragments. Suevite, Specimen No. 108. Plane-polarized light, 100 X.

mate dimensions of 2 cm by 10 cm by 20 cm and is strongly twisted, like the "pancakes" (Fladen) of the original suevite from the Ries Crater.

When the content of glass in the suevite decreases, the proportion of relatively regular bombs and lapilli generally increases. Bodies resembling ribbon bombs or ribbon lapilli (Macdonald, 1972, p. 125) are common among the more regularly shaped glass fragments. Twisted and bent ribbon lapilli (Fig. 6) up to 10 cm in length but only a few mm in diameter have been observed. Some of the smallest glass grains that resemble lapilli are more or less drop-like in shape. Elongated lapilli are sometimes fibrous at one or both ends.

The Lappajärvi suevite appears extremely heterogeneous under the microscope. The glassy bodies are mixtures of various molten glasses rich in schlieren and mineral fragments. Very thin but continuous dark flow lines, bent in the vicinity of mineral inclusions and vesicles, may penetrate a glass bomb several cm long. The glass is usually transparent light brown, but greenish or almost colorless varieties occur, and alternation of dark and light bands or schlieren is a typical feature. Very dark brown or almost opaque patches of glass surround or replace strongly altered, oxidized biotite flakes. Fused feldspar and silica (lechatelierite) occur as short, sinuous "worms" of clear glass. However, the glasses have recrystallized or begun to recrystallize almost everywhere, which prevents their detailed investigation and sometimes makes it difficult to determine their origin and texture.

Mineral inclusions are common in the glasses with fluidal texture. They involve "unshocked" fragments of quartz and feldspar, which may vary from clear and angular to brownish or greenish and strongly altered, diapletic quartz and feldspar, and quartz and feldspar glass. Because of the almost ubiquitous recrystallization of the glasses, it is difficult to distinguish between grains of diaplectic (or thetomorphic) and normal (fused) glass. Vesicles and flow lines are typically absent in diaplectic glass (Stöffler and Hornemann, 1972, p. 372), and it has the shape of the original mineral grain. Almost all feldspar glass inclusions in the glass bodies, however, show intensive vesiculation and flow textures indicating a fused glass origin. Devitrification has formed sheaf-like bundles of crystallites which delineate plastic distortions of the original feldspar grains. The quartz glass often occurs together with quartz inclusions displaying abundant planar elements (shock lamellae). There are (e.g. specimens Nos. 236 and 237) some unrecrystallized clear quartz glass inclusions containing coesite aggregates (see Fig. 13) very similar to some diaplectic quartz glass inclusions in the suevite from the Ries Crater (cf. Fig. 1 of Stöffler, 1971b). Along with the thetomorphs of quartz, inclusions of strongly shocked quartz occur that have a core of brownish subconchoidally broken quartz surrounded by a clear mixture of tridymite or cristobalite, or both, and, probably, quartz glass. In the vicinity of these quartz inclusions the glass with the fluidal texture is nearly colorless and may contain some serpent-shaped strings of lechatelierite. These quartz inclusions are considered to be partly melted quartz grains. Because the same small glass body also contains "unshocked" feldspar and quartz inclusions, the mixture cannot result from direct fusion of incorporated quartz grains but must have been produced by mixing of materials already wholly or partly melted by the shock waves.

Kärnäite

For kärnäite-like rocks from other meteorite impact sites (astroblemes) terms such as impact lava or impact melt, melt rock, coherent breccia and welded breccia have been used. Perhaps the most descriptive, although lengthy, name for this kind of rock would be breccia with lava-like groundmass. In this paper, however, the term kärnäite is adopted for the pertinent Lappajärvi rock because it is an old, common and well-known term in Finland.

The prevailing type of kärnäite is a peculiar, hard and megascopically lava-like rock which on the weathered surface may have the appearance of a breccia or an agglomerate. Here and there the kärnäite usually contains amygdules of calcite and chalcedony and open vesicles. In addition, vesicles containing a few rhombohedral crystals or rosettes of hematite altered to goethite are common. Some blocks or boulders with abundant open vesicles of different sizes have

been observed. Many blocks and boulders of kärnäite contain distinctly oriented vesicles, usually elongated in one direction. In Outcrop No. 10 (Fig. 3) the vesicles are elongated in a vertical direction and flattened in an east-west direction. This indicates that, probably as a consequence of slumping and deformation of the crater, the material has moved until finally consolidated.

Well-developed vertical jointing is visible in some kärnäite outcrops. The best example of this is Outcrop No. 1 where the direction of the jointing is E-W. In the eastern kärnäite outcrops, e.g. Outcrop No. 10, the direction of the jointing is about N 70° E. The jointing is best developed in outcrops where the kärnäite is relatively homogeneous, i.e. where the groundmass is dense and partly glassy and the fragments are strongly assimilated.

The rock that crops out in the northern part of the island Kärnänsaari and on the islands Vartijasaari and Lokkisaari (Fig. 3) was first mentioned in Finnish geological literature by Holmberg (1858). In his catalogue of Finnish minerals, rocks, etc., he wrote (translated from the Swedish): "On the Kärnä Island there occurs a porphyry which consists of a dark groundmass with reddish brown feldspar phenocrysts".

This peculiar rock was named kärnäite by Berghell (1921), who visited the lake area in 1913. At a meeting of the Geological Society of Helsingfors (Helsinki) in 1920, Berghell described a volcanic rock that occurs on the island Kärnänsaari in the lake Lappajärvi. The chemical composition of the rock is that of dacite — the chemical analysis was published by Eskola (1921, p. 12) and Laitakari (1942, p. 39) — but on the basis of its particular texture Berghell preferred to call it kärnäite. According to him, most of the feldspar phenocrysts of the kärnäite are monoclinic, probably mostly sanidine, but microcline and scanty plagioclase are also present. On Kärnänsaari kärnäite consists mainly of agglomerate-like tuff that is only exceptionally free from inclusions, which include granodiorite-like rocks, quartzite, mica schist and arkosic sandstone.

In an unpublished manuscript Kaikko (1921?) gave a detailed description of the kärnäite. He noted two distinct types of kärnäite: (1) common tuff-like kärnäite with a dull glassy groundmass and (2) kärnäite with a clear glassy groundmass with abundant microlites of feldspar, pyroxene and star-shaped cordierite trillings. He noted the large amount of rock and mineral fragments in the kärnäite, and the quartz fragments with subconchoidal fractures. Kaikko also described corroded skeletal feldspar fragments and a partly melted feldspar fragment with flow texture containing a quartz grain with subconchoidal fractures, and some quartz-feldspar-(biotite) rock fragments some of which were partly recrystallized or strongly altered. All the phenomena observed by Kaikko were ascribed to a volcanic origin. He concluded that kärnäite is a volcanic rock consisting of lava and tuff and has embraced a large amount of bedrock fragments.

Eskola (1921) compared the chemical analyses of rocks from four Fennoscandian crater lakes (Lakes Mien and Dellen in Sweden, Jänisjärvi in the U.S.S.R. and Lappajärvi in Finland) which have many features in common, and concluded that the rocks have somewhat uncommon chemical compositions and that their petrographic names are without significance. Eskola (1927, pp. 5—6) described the rock of Kärnänsaari as a porous dacite, sometimes resembling agglomerate, which contains rather large feldspar phenocrysts in a dense groundmass.

Laitakari (1942, pp. 38—40) described the kärnäite as a dark gray porphyritic dacite with a typical lava structure. It contains feldspar phenocrysts and oriented amygdules of calcite.

Saksela (1949, pp. 22—23) emphasized that kärnäite is not a porphyritic rock but contains a considerable number of mineral and rock fragments. The mineral fragments are of varying sizes, most of them with a diameter of a few mm, although fragments with a diameter of some cm also occur. The fragments are mostly pieces of feldspar (potassium feldspar and plagioclase) and quartz. The strongly altered bedrock fragments consist of granite pegmatite, gneissic granite and biotite-plagioclase gneiss. Moreover, Saksela stated that under the microscope the large number of mineral fragments in kärnäite and its clastic texture become visible. It is his opinion that kärnäite consists of pyroclastic material, agglomerate and tuff, and that the amount of real lava is relatively small.

Simonen (1964b, p. 115) regarded the kärnäite as a dense volcanic rock containing feldspar phenocrysts and amygdules of calcite and chalcedony. Because of the appreciable number of bedrock fragments, Simonen concluded that the rock is a volcanic breccia formed when gases, dissolved in a magma, escaped in an explosion-like eruption.

The present author has studied several hundred kärnäite specimens in the field and in the laboratory as well as more than 40 thin sections made from the kärnäite specimens collected. Numerous X-ray diffractograms and powder patterns of the constituents of kärnäite have been recorded and measured. The author is convinced of the following:

1) Kärnäite is not a porphyritic rock but contains numerous small and large mineral and bedrock fragments.

2) The mineral fragments consist of quartz and feldspar.

3) The bedrock fragments are the same as those in the impact breccia, i.e. medium- to coarse-grained granite, and mica gneiss, granodiorite and migmatite. Most of the fragments are strongly altered because of recrystallization and partial melting or assimilation (ghost-like remnants), which sometimes makes it difficult to determine the original rock type.

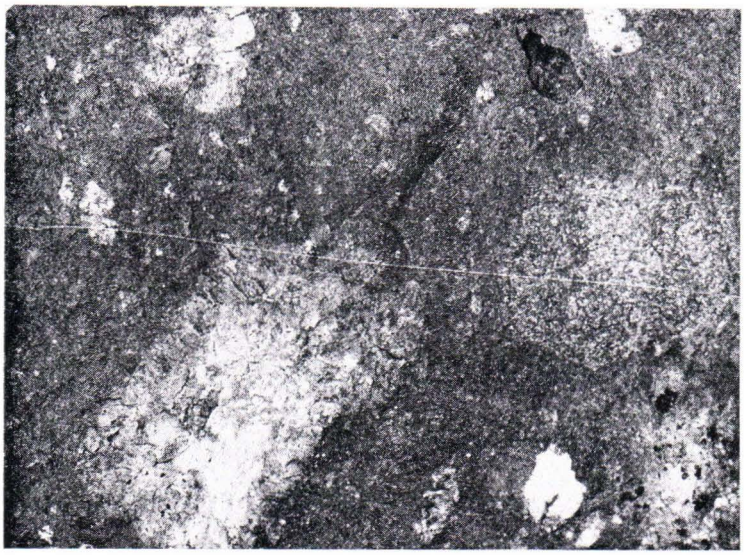

Fig. 8. Kärnäite, broken surface. Ghost-like remnants of granite (some of them vesiculated) and mica gneiss (right side). Outcrop No. 1 in Fig. 3. Half of natural size.

4) Only one fragment of sandstone has been discovered, and it is likely that the subconchoidally broken pieces of quartz are those formerly recognized as quartzite or sandstone inclusions.
5) The structural state of all feldspar fragments X-rayed is of the high sanidine — high albite series (Wright and Stewart, 1968, p. 65), sanidine being much more common than anorthoclase and high albite.
6) As Kaikko (1921?) observed, there are at least two different types of kärnäite.
7) No contact is visible between the kärnäite and bedrock or between the kärnäite and breccias in the outcrops investigated.

Kärnäite may be defined as a breccia with lava-like groundmass. This definition is valid for both types of kärnäite, but especially for the Type I kärnäite (Lehtinen, 1970), which is characterized by the following features:

1) Clear or dull glassy groundmass with microlites of sanidine, anorthoclase, pyroxene and cordierite.
2) Quartz fragments always rimmed by pyroxene microlites.
3) Feldspar fragments that have been corroded and transformed into, or recrystallized as, sanidine or anorthoclase, or both.
4) Bedrock fragments (granite, granite pegmatite, granodiorite, mica gneiss) that sometimes indicate eutectic melting at quartz-feldspar grain boundaries. The melt-

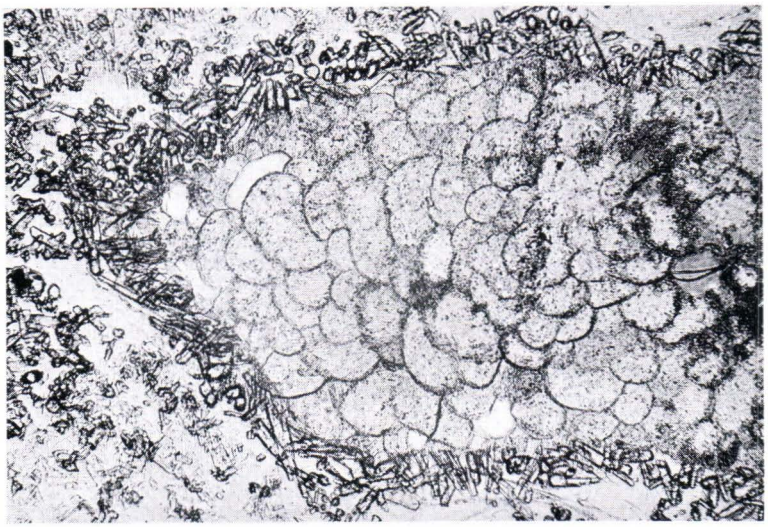

Fig. 9. Subconchoidally fractured quartz fragment rimmed by orthopyroxene microlites. Type I kärnäite, Specimen No. 216, plane-polarized light, 100 X.

ing has produced a clear glass which contains only a few feldspar and pyroxene microlites.

The quartz fragments have usually been fractured subconchoidally (Fig. 9), obviously because of thermal stress, but fragments with well-developed planar features are also present. Some small quartz fragments have been wholly altered to cristobalite and cracks in large quartz fragments filled with cristobalite. Some cristobalite grains probably represent the recrystallized thetomorphic or melted quartz fragments that French *et al.* (1970, p. 4401) have observed in the "melt rock" of Tenoumer Crater, Mauritania. X-ray diffractograms of groundmass drilled from kärnäite specimens show that cristobalite is an important constituent of several kärnäite specimens, especially of those showing secondary alteration. The alteration also appears as chloritization of corroded feldspar fragments and transformation of colorless orthopyroxene microlites into greenish clinopyroxene (or amphibole) or brownish biotite, although most of the specimens of Type I kärnäite are fresh and unaltered.

Cordierite microlites exhibit a very complicated twinning which, because of their small size, is difficult to solve, but most of them are twelve-pointed stars or twelvelings (cf. "cross-sexuplet" of Strunz *et al.*, Fig. 4h, 1971). This kind of complex twinning is generally regarded as typical of cordierite of high-temperature origin.

In a hand specimen, Type I kärnäite is bluish or grayish black, and the rock splits subconchoidally into harp-edged fragments with fairly even surfaces. The

specimens that have a groundmass of clear glass show a silky luster on their broken surfaces. They are, however, rare because the groundmass has commonly recrystallized as laths or spherules of feldspar and quartz which include microlites of pyroxene, feldspar and cordierite.

The other type of kärnäite, Type II (Lehtinen, 1970), is distinctly heterogeneous even megascopically and often contains numerous small cavities of very irregular shape. When struck with a hammer, the rock is not as hard as Type I, but it crumbles or breaks along rugged surfaces. The rock is bluish or grayish black, but very light-colored, gray or greenish gray varieties occur. The light color is often due to secondary alteration (leaching). Some rare, highly vesicular varieties of kärnäite have also been included in Type II.

The heterogeneity of Type II kärnäite is much more pronounced in thin sections studied under the microscope. The rock consists of fragments or "bombs" of varying size, rich in glass but relatively poor in mineral and rock inclusions, and of breccia material occurring among the glassy fragments. The glass of the glassy fragments has been devitrified, and it now contains laths and spherules of feldspar (cf. the groundmass of Type I) or it has been altered into chlorite, clay minerals and hematite. Nevertheless, the original textures have survived and are still recognizable (cf. Dence, 1968, p. 173). Mineral inclusions and vesicles within the glass fragments are usually more or less clearly oriented, and the glass itself has a distinct fluidal texture. In adjacent glass fragments, however, the flow lines have different orientations, as in suevite. Most of the glass fragments are elongated, sometimes bent and of irregular shape. Their length varies from about 1 mm up to several cm. Pyroxene and feldspar microlites occur only sporadically in the glass fragments, but numerous small crystallites may make the groundmass (glass) very dull. A very thin rim of small slender pyroxene microlites surrounds some quartz inclusions. Most the pyroxene microlites have been altered to green mica.

Cristobalite is a common constituent of Type II kärnäite. In places, subconchoidally fractured quartz has been partly or wholly transformed into cristobalite (cf. the quartz fragments of Type I kärnäite). Some twinned crystals of tridymite have been identified in the vesicles in the rock.

The fragmental breccia that occurs between the glassy fragments or "bombs" contains the same minerals as the impact breccia, except that unaltered flakes of biotite are extremely rare in the kärnäite. Nevertheless, some flakes have been observed and display strong kinking, somewhat lowered birefringence and are faded brown in color. Almost all biotite flakes have been decomposed into a brown or green substance with numerous small crystallites, probably magnetite or spinel. Partly melted feldspar fragments with fluidal vesicular texture are common. All the fragments X-rayed were composed of sanidine or anorthoclase, or both.

The Type II kärnäite contains more abundant mineral and rock fragments than the Type I kärnäite. The content of unaltered fragments and of fragments displaying weak shock effects is also high. Thus, for instance, "unshocked" fresh feldspar fragments absent in Type I kärnäite are common in Type II. The mineral and rock inclusions found in the glassy "bombs" are, however, very similar to those described in Type I kärnäite.

A noteworthy feature of Type II kärnäite is that the matrix of the fragmental breccia occurring between the glassy "bombs" in some specimens consists of clear glass without a fluidal texture. Consequently, there are two kinds of glass in these specimens: an "older" dull glass where the fluidal texture is still recognizable, and a "younger" one forming the clear matrix of the rock. This means that during or after the deposition of the glassy bodies and breccia material the temperature was high enough to produce some new melt or, more probably, that some melt and hot gases squeezed into the loose fallback immediately after its deposition. During a hypervelocity meteorite impact, the melts and gases formed are partially in a superheated state (Dence, 1971, p. 5556) and are very effective as melting and assimilating agents. This process is greatly facilitated because, as a result of the action of shock waves, unfused but shocked rocks and their constituents are broken into pieces, and the mineral lattices are deformed or disordered. This kind of rock is very readily affected by melting and assimilation.

However, features indicating rapid cooling and assimilation on a small scale only are widespread in the kärnäite, particularly in Type II. Shards of clear quartz or quartz with planar features, angular fragments of feldspar and kinked biotite, absence of phenocrysts, survival of vesiculated feldspar inclusions and glassy "bombs" with fluidal texture show that assimilation was of little significance during the formation of kärnäite. Instead of being assimilated or altered by the melt, most of the material melted totally or partially, or was already transformed into a glass by the shock waves. The melt did not supply additional heat but only promoted homogenization in the more massive parts of the melt layer (Type I kärnäite).

The author regards Type II kärnäite as a firmly welded fallback breccia, i.e. as a fragmental rock consisting of originally fused material and bedrock fragments that fell back into the crater at high temperatures and were transformed by sintering and crystallization into a hard rock which in a hand specimen may resemble an agglomerate. The sintering, crystallization and recrystallization homogenized the rock and obscured its texture.

Type I kärnäite passes gradually into Type II, and it is sometimes difficult to distinguish between Type II and suevite. The groundmass of Type I kärnäite with microlites of pyroxene, feldspar and cordierite represents the impact melt proper which has incorporated brecciated but unmelted or partially melted (selec-

tively melted) material although it could not assimilate all of it. The author believes that Type I kärnäite consists of material that was not actually ejected from the crater. If so, the difference between the types was originally due to differences in their depth of formation and burial in the crater.

Amount of kärnäite

On the basis of the data available it is impossible to decide whether the kärnäite outcrops represent separate small units or are peaks of a large uniform lens in the crater. The occurrence of kärnäite blocks and boulders in the vicinity of the outcrops and between them suggests, however, that there are, not one large unit, but at least two separate units, viz. a western and an eastern one. The western unit consists of outcrops Nos. 1—3 and the eastern unit Nos. 4—10 (see Fig. 3). In addition, there are areas where the number of kärnäite blocks abruptly increases possibly indicating an outcrop nearby. For instance, in Figure 3, point No. 11 represents a small area where almost 100 % of the blocks and boulders consist of kärnäite, but no outcrop has been discovered.

The occurrence of kärnäite in the crater is presumably somewhat similar to the occurrence of welded breccia at Nicholson Lake in Canada (Dence *et al.*, 1968, pp. 349—358). This lake seems to occupy a deeply eroded crater where a discontinuous and irregular cover of tough and compact welded breccia resembling kärnäite overlies strongly fractured gneisses that show distinct evidence of weak shock metamorphism. Dence *et al.* (1968, p. 353) observed, further, that "the breccias were laid down on an irregular surface with a relief more than 160 feet (50 m)". They could not, however, find eroded breccias thicker than 5 m. They also noted that the contact was generally subhorizontal and sharp but irregular in detail.

The energy required to produce a crater with a known diameter may be calculated using Baldwin's (1963, p. 162) empirical formulas 8-1A and 8-3A for the relation of the crater diameter to the total energy of the impacting meteorite. The first formula is for explosions of surface-burst type and the second is valid when the scaled depth of the burst is 0.10; for craters of "medium" size one must interpolate between the two formulas (Baldwin, 1963, p. 186). If it is assumed that 5 % of the total energy (cf. Beals, 1965, p. 909) is used to melt the granitic target rock with specific heat 0.25 cal/g, that the heat of fusion is 70 cal/g and that the melting point is 1500°C (Birch *et al.*, 1942), the data in Table 1 are obtained. In the calculation of the volume of impact melt a value of 2.65 g/cm^3 was used for the density of the melt, and the effect of vaporization was partly omitted by assuming that the vaporized material condensed and returned its heat of vaporization to the melting process.

Table 1

Production of impact melt

Crater diameter (km)	Total energy (cal)	Volume of impact melt (km^3)
5	1.6×10^{18}	0.067
6	3.2×10^{18}	0.14
8	9.8×10^{18}	0.42
10	2.4×10^{19}	1.0
12	4.9×10^{19}	2.1
15	1.2×10^{20}	5.1

It is impossible to state precisely the amount of impact-produced melt in the crater of Lappajärvi, but on the basis of Mölder's (1948) data the amount of kärnäite in the glacial drift may be roughly estimated. Most of it seems to occur within an indicator fan having an apical angle of about 35°, a radius of 60 km, and an average kärnäite content of 5 % to 10 % (Mölder's map No. 2). If the thickness of the kärnäite-bearing glacial drift is supposed to be 3 m to 4 m (see Okko, 1964, p. 243) the indicator fan would contain 0.15 km^3 to 0.45 km^3 kärnäite. The glacial drift also contains suevite, which greatly increases the amount of melted material, but the content of melt is usually 30 % to 50 % in kärnäite and still lower in suevite. This very rough estimate suggests that, taking the amount of melted material still left in the crater into consideration, the diameter of the original Lappajärvi crater could have been of the order of 10 km to 12 km (cf. Table 1).

Stages of shock metamorphism

On the basis of detailed investigations, described in the previous and following chapters, of the shock-metamorphosed rocks and their constituents, the following tabulation showing the stages of shock metamorphism for the Lappajärvi rocks has been established.

Shock Stage I:

— kink bands in micas
— planar deformation structures in quartz and feldspars

Shock Stage II:

— diaplectic (thetomorphic) glasses of quartz, plagioclase (maskelynite) and microcline. The impact breccia consists mainly of unshocked material (no shock effects visible) and of material of Stages I and II

Shock Stage III:

— recrystallization of feldspars to form vesicular grains or spherules of sanidine and anorthoclase
— recrystallization of quartz (formation of quartz mosaic and chalcedony)
— partial decomposition of biotite
— partial destruction of original textures of the rock

Shock Stage IV:

— individual melting of minerals according to their melting points (quartz often remains unmelted)
— eutectic melting at quartz-feldspar grain boundaries. The partly melted material is mixed with other mineral and rock fragments of lower shock stages to form "impact lava" or firmly welded breccia. The rock produced is locally called kärnäite
— total destruction of original textures of the rock

Shock Stage V:

— individual melting of minerals to form heterogeneous glasses of various colors and refractive indices and with fluidal texture. This material is ejected from the crater, mixed in flight with various mineral and rock fragments of lower shock stages, and deposited in and around the crater. Many of the glass fragments, especially those with a rock-fragment core, have aerodynamically shaped forms. The brittle depositional breccia produced is called suevite.

In a hand specimen it is often difficult to distinguish among certain shock stages. Thus, for descriptive purposes and field work, the rocks have been classified as follows: (1) from weakly to moderately shocked (Shock Stages I and II), (2) strongly shocked (Shock Stage III), (3) intensely shocked (Shock Stages IV and V).

Mineralogy

The most important criteria for shock metamorphism are not petrological but rather mineralogical, although recent investigations seem to support the idea that the occurrence of rocks like suevite or kärnäite is in itself a valid criterion proving that the crater (scar) from which the rocks derive originated by meteorite impact.

Because the target rocks of the Lappajärvi Crater are mostly granitic, the most important minerals of the shock-metamorphosed rocks are quartz, alkali feldspars, plagioclase and biotite. Detailed descriptions of both naturally and synthetically shocked quartz and plagioclase have been published. The present author has paid particular attention to the investigation of minerals with shock-

induced properties not known in detail, viz., alkali feldspars, some accessory minerals and coesite, the critical mineral.

Quartz

The behavior of quartz under very high pressures and strain rates has been investigated during the last ten or twenty years. Information has been obtained from a variety of sources, e.g. (1) Experiments under controlled laboratory conditions to synthesize high-pressure polymorphs of silica or to produce plastic flow and mechanical deformation in quartz; (2) Shock wave experiments (laboratory-scale and underground explosions) to determine the Hugoniot equation for quartz or to investigate shock-induced deformations and transformations of quartz under well-known or moderately well-known pressure and temperature conditions; (3) Investigation of petrographic shock properties, deformation and transformation of quartz and quartz-bearing rocks from terrestrial meteorite impact craters.

The voluminous literature, including many unpublished reports, dealing with these investigations has been reviewed by Stöffler (1972, 1974). In the present paper, only quartz with planar deformation structures will be discussed briefly. Solid-state and thermally induced transformations of quartz will be examined later in connection with coesite (p. 38).

The Hugoniot elastic limit (HEL) for quartz is about 100 kb to 120 kb (Wackerle, 1962, p. 933; Ahrens and Rosenberg, 1968, p. 71). When shocked above the HEL, the compressibility of quartz increases anomalously, and the mineral yields plastically, producing certain crystallographically controlled, thin planar deformation structures (Hörz, 1968; Müller and Déforneaux, 1968). At lower pressures, open irregular or planar fractures, mainly parallel to $\{10\bar{1}1\}$, $\{10\bar{1}0\}$ and $\{0001\}$, are formed. The experimental results also show that sets of closely spaced planar structures parallel to $\{10\bar{1}3\}$ are produced by shock pressures just exceeding the HEL of quartz but that to produce $\{10\bar{1}2\}$ orientations, pressures above 160 kb to 200 kb are needed. In addition, planar structures parallel to eight less prominent orientations have been observed in experimentally shocked quartz (Hörz, 1968; Müller and Défourneaux, 1968). The experimental results are in very good agreement with data on naturally shocked quartz from several meteorite craters (e.g. Bunch, 1968, pp. 415—423; v. Engelhardt *et al.,* 1968; Robertson *et al.,* 1968, pp. 435—444; Short, 1970b, pp. 627—629) and with data on quartz from underground nuclear explosion craters (Short, 1966, Fig. 8; 1968, pp. 207—208; 1970a, pp. 711—719).

The thin planar deformation structures of quartz (and of some other silicates) have been variously described as planar features (e.g. Carter, 1965, pp. 798—801; Robertson *et al.,* 1968, p. 445), planar elements (v. Engelhardt,

1967a, pp. 179—180), and shock lamellae (Chao, 1967b, p. 194). The usage of these and related terms is still confused, especially because until about 1966 the structures were usually described as planar fractures or cleavages. They are not real fractures but consist of glassy material or of material with lower refractive indices and birefringence (Chao, 1967b, pp. 193—195; v. Engelhardt *et al.*, 1968, p, 480). In addition, Carter (1965, 1968a,b) distinguished between planar features and deformation lamellae of quartz. Because it is difficult to distinguish between the two planar deformation structures (phase-contrast illumination is needed), they are often combined as "planar features" (see Hörz, 1968, p. 247; cf. also Robertson *et al.*, 1968, pp. 444—445).

v. Engelhardt and Bertsch (1969) described not only planar fractures but also decorated and non-decorated planar elements, filled lamellae and homogeneous lamellae of quartz from the Ries breccias, Germany. The decorated planar elements seem to be typical of quartz of old meteorite craters (Robertson *et al.*, 1968, p. 444; Short, 1970a, p. 711; see also v. Engelhardt *et al.*, 1968, p. 480).

The number of multiple sets of planar elements and their intensity increases with rising peak pressure. The increase is accompanied by gradual, although not linear, changes in the physical properties of quartz (see Chao, 1967b, pp. 193—196; 1968, pp. 223—232; Hörz, 1968; v. Engelhardt and Bertsch, 1969; Stöffler and Hornemann, 1972; see also Stöffler, 1974). The mixed phase regime of quartz (region of anomalous high compression) extends at least to about 330 kb (Wackerle, 1962, p. 935; Ahrens and Rosenberg, 1968, Figs. 6—7; Hörz, 1968, p. 252), and higher pressures cause total transformation to a short-range-order phase. W. v. Engelhardt (French, 1966, p. 906) proposed the term diaplectic, and Chao (1967a, p. 212; 1967b, p. 195) used the term thetomorphic to describe such solid-state glasses. To describe minerals with shock-induced deformation structures the term diaplectic minerals (e.g. diaplectic quartz) has been suggested (v. Engelhardt and Stöffler, 1968, p. 163).

Hugoniot and release adiabat data (Wackerle, 1962, p. 936; see also Chao, 1968, pp. 216—219; Stöffler, 1974, pp. 261—262) indicate that when peak pressures exceed about 500 kb, post-shock temperatures exceed the melting point of quartz, and "normal" liquid-state glass (lechatelierite) is obtained. Porous rocks like sandstone have a different thermal shock history, and in them weaker shock waves can produce fused silica glass (De Carli and Milton, 1965, Table 1; Kieffer, 1971, pp. 5463—5471).

The whole range of shock-induced deformations and transformations, excluding stishovite, observed in naturally or experimentally shocked quartz also occurs in the Lappajärvi quartz. Diaplectic or thetomorphic glass of quartz is, however, uncommon, but the presence of the high-pressure polymorph of silica, coesite, has been verified in the Lappajärvi rocks. Because shocked quartz is

the most common mineral at Lappajärvi, its associations and some properties are discussed in connection with the descriptions of kärnäite, suevite, and coesite and microcline (alkali feldspars).

Quartz with planar deformation structures occurs in all the Lappajärvi rocks of shock-metamorphic origin, but these structures are generally best preserved in the impact breccia. For the Lappajärvi quartz the method of v. Engelhardt and Bertsch (1969, pp. 211—213) was applied to study the orientation of the deformation structures. Although it is thus possible to identify the pole of each set of planar structures, a distinction cannot always be made between positive and negative forms. Another method of studying the orientation is to measure the angle between the c-axis and the pole normal for each set of planar structures and to plot it in a histogram to express the frequency distribution of the total number of sets (e.g. Carter, 1968a, pp. 454—456; Short, 1970b, pp. 627—629).

The present author has not systematically studied the orientations of planar deformation structures in the Lappajärvi quartz, but has measured 20 to 30 quartz grains from suevite and impact breccia. Svensson (1968) measured the orientation of planar elements of 16 quartz grains from kärnäite. He found elements parallel to $\{10\bar{1}3\}$, $\{10\bar{1}2\}$, $\{21\bar{3}1\}$ and $\{10\bar{1}1\}$, the most common orien-

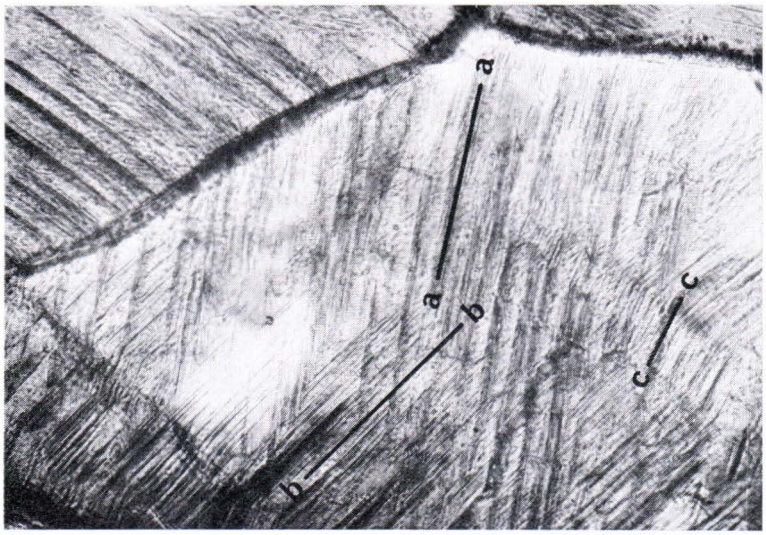

Fig. 10. Two quartz grains with multiple sets of closely spaced planar elements. a—a = $(\bar{1}103)$, b—b = $(\bar{1}013)$, c—c = (0001). In addition, the lower grain has planar elements parallel to $(3\bar{1}\bar{2}1)$ and $(6\bar{5}\bar{1}1)$. In the upper grain, planar elements are parallel to $\{10\bar{1}3\}$ (strong elements) and to (0001). Granite fragment of impact breccia, Specimen No. 105. Crossed polarizers, 310 X.

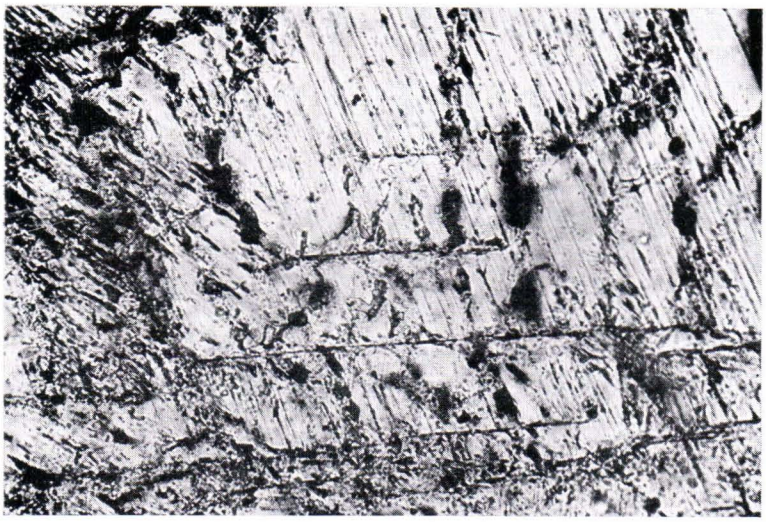

Fig. 11. Small, oriented patches of diaplectic glass (left side), planar fractures (nearly horizontal) and partly decorated planar elements. Quartz from Specimen No. 105. Crossed polarizers, 280 X.

tation being $\{10\bar{1}3\}$. In addition to the orientations observed by Svensson, the present author found planar elements parallel or closely parallel to $\{10\bar{1}0\}$, $\{11\bar{2}0\}$, $\{51\bar{6}1\}$ and $\{0001\}$ in the quartz of the breccias. As in the quartz of kärnäite, most of the sets of planar elements are in the $\{10\bar{1}3\}$ orientation; these sets also prevail because their elements are relatively long, well-developed, multiple and very closely shaped.

In single grains of quartz from kärnäite and the breccias, the number of sets of planar elements per grain varies greatly, but within a single rock fragment the quartz is more evenly shocked. Nevertheless, it is very typical of coarse-grained quartz of granite or granite pegmatite fragments that the planar elements are irregularly distributed and tend to concentrate near grain boundaries. The marginal areas that are full of closely spaced sets of planar elements have a reduced birefringence (from gray to almost black) and lowered refractive indices. The central parts of large quartz grains may be devoid of planar elements. Another feature typical of the large grains is that in different parts of a single grain the planar elements may have different orientations, although the number of sets may be the same (cf. Robertson *et al.,* 1968, pp. 441—442).

Both decorated and non-decorated planar elements occur in the Lappajärvi quartz, but homogeneous lamellae (v. Engelhardt and Bertsch, 1969, pp. 207—209) have not been observed. Bent elements and filled lamellae are not uncommon, but instead of stishovite or coesite (v. Engelhardt and Bertsch, 1969,

Fig. 5; Stöffler, 1971b, Figs. 2, 3 and 6) the filling appears to consist of material with very low refractive indices and birefringence, conceivably glassy quartz or a zeolite mineral of secondary origin.

Large grains of quartz or their interiors usually have a very irregular optical extinction pattern accompanied by planar and irregular fractures. Within a single grain, the extinction may be irregular or patchy and occur within a range of 10° to 20°. This kind of quartz is very often brownish owing to irregularly distributed pigment. The plastic deformation, mosaicism, appears to have been produced by relatively weak shock waves, although quartz with mosaicism often has planar elements, too. If the quartz has partly recrystallized the mutually disordered domains are still recognizable because recrystallization begins from fractures and areas of strong deformation. In quartz with mosaicism, deformation bands (Bunch, 1968, pp. 419—422) have been observed but they are not common at Lappajärvi.

It is not always possible to distinguish between planar fractures and planar elements (see Stöffler, 1972, p. 81). They often occur together, and a deformation structure regarded as a planar fracture near the grain boundary may, in the interior of the grain, continue as a very thin non-decorated planar element. Two types of planar fractures exist in the Lappajärvi quartz: (1) fractures at which planar elements terminate (sometimes in a fish-bone fashion), and (2) younger fractures cutting sets of planar elements. The planar fractures were evidently formed upon both shock compression and pressure release. Planar fractures following the $\{10\bar{1}1\}$, $\{10\bar{1}0\}$, $\{10\bar{1}3\}$ orientations and the basal pinacoid have been found, but many irregular fractures also occur.

A peculiar type of microstructure (ovoids) is fairly common in the Lappajärvi quartz. Similar microstructures have been found in quartz from the Deep Bay Crater, Canada (Short and Bunch, 1968, Fig. 13; Short, 1970a, Plate 6D). The formation of such ovoid microstructures is apparently not related directly to the shock itself but must depend on the post-shock environment. At Lappajärvi, quartz with ovoid microstructures is associated with kinked biotite, diaplectic feldspars, or diaplectic glasses of feldspars, or both, in moderately shocked specimens of granite. Quartz grains with and without ovoids may occur in the same specimen. There are quartz grains consisting of subrounded or ovoid bodies of "normal" quartz with some weak planar elements. The ovoids ($\varnothing = 0.02$ mm) are surrounded by almost isotropic rims of silica. This kind of microstructure passes gradually into another type in which only a few ovoids of intact quartz occur in a single grain. The quartz between the ovoids displays abundant very closely spaced sets of planar elements and has a low birefringence and low refractive indices. In addition, this "matrix" quartz is often colored by a brownish pigment, or has been altered to zeolites, or begun to recrystallize.

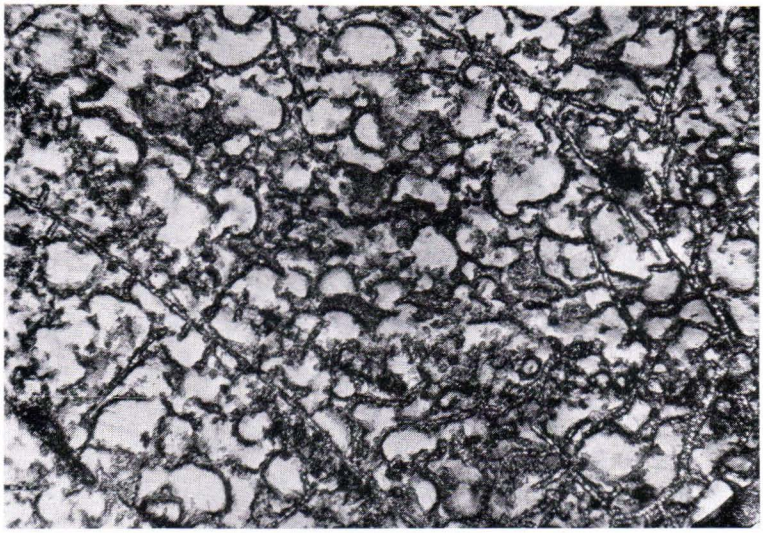

Fig. 12. Subrounded bodies of "normal" quartz surrounded by almost isotropic rims of silica. Granite pegmatite fragment from impact breccia, Specimen No. 134. Crossed polarizers, 100 X.

X-ray diffractograms were recorded on a Philips Norelco diffractometer with Cu/Ni radiation for the quartz specimens listed in Table 2. The peak intensities on the diffractograms do not differ much from those of standard quartz. The only exception is specimen No. 257, whose diffractogram exhibits nothing but traces of the strongest peaks. The peak positions were corrected by means of an internal silicon standard. The corrected 2 Θ-values show a small but systematic transition toward lower values, indicating enlarged unit-cells. The calculated unit-cell parameters are 0.002 Å to 0.005 Å greater than those of standard quartz, although in most cases the parameter differences are included in the estimated limits of error. Chao (1968, p. 227) assumed that so enlarged a unit cell is thermally induced, and thus the expanded cell is an indicator of the residual temperature from which the quartz was quenched. However, by use of the calculated unit-cell parameters of the Lappajärvi quartz and the thermal expansion data for quartz (Skinner, 1966, Table 6—7), very low ($<100°$ C) residual temperatures are obtained.

The refractive indices were determined for eight single grains of diaplectic quartz (Table 2). To the naked eye the grains were white or yellowish, only very weakly transparent and with a pearly luster on their broken surfaces. The quartz from specimen No. 257, however, was colorless and almost transparent. These findings are consistent with those of Chao (1968, p. 226) on the Ries quartz.

Table 2

Refractive indices (n_o, minimum value, n_e, maximum value), n_e—n_o (not actual birefringence), mean refractive index [$n_m = (2n_o + n_e)/3$], and content of amorphous silica (x_n, volume %) in eight quartz grains with planar deformation structures. Immersion method, sodium light, accuracy ± 0.001

No.	n_o	n_e	n_e—n_o	n_m	x_n
207.....	1.541	1.551	0.010	1.544	3.0
265.....	1.539	1.548	0.009	1.542	5.7
93.....	1.537	1.547	0.010	1.540	7.6
262.....	1.536	1.548	0.012	1.540	8.0
215.....	1.536	1.547	0.011	1.540	8.3
208.....	1.535	1.547	0.012	1.539	9.1
218.....	1.533	1.545	0.012	1.537	11.4
257.....	1.492	1.496	0.004	1.493	61.0

If, in agreement with v. Engelhardt and Bertsch (1969, p. 222) it is assumed that diaplectic quartz is a mixture of thin lamellae of amorphous silica embedded in a matrix of crystalline quartz, it is possible, on the basis of the measured refractive indices, or density, or both, to calculate the content of disordered phase present. For each grain of quartz (Table 2) the minimum and maximum values for the refractive indices were determined. In a certain part of a diaplectic quartz grain, however, both birefringence and refractive indices were lowered. Thus the range in refractive indices is a measure of heterogeneity within a single grain. A quartz grain from specimen No. 257, contained also a very small amount of quartz that had refractive indices above 1.53.

Material from a crushed quartz grain (No. 208) was heated at 500° C ± 20° C for five days. No change in refractive indices was observed.

Coesite

The presence of one or both of the two high-pressure polymorphs of silica, viz. coesite and stishovite, is generally taken as important evidence of shock metamorphism induced by a hypervelocity meteorite impact.

Pressures in excess of 20 kb are required to synthesize coesite under static conditions (see Coes, 1953; Boyd and England, 1960; Bell and Boyd, 1968). Still higher pressures, exceeding 75 kb (Stishov and Popova, 1961; Sclar *et al.*, 1962; Stishov, 1963; Gigl and Dachille, 1968), are needed to produce stishovite. The p-T conditions required for the quartz-coesite transition are reached at a depth of the order of 100 km (MacDonald, 1965, Fig. 2), and it seems impossible that even coesite could form in normal regional metamorphism.

In shock metamorphism, the formation and preservation of coesite and stishovite is controlled, not only by the p-T conditions within the shock wave itself and the kind of material shocked, but also by the sluggishness of the quartz-coesite transition (Boyd and England, 1960, p. 750; De Carli and Milton, 1965) and by the instability of stishovite at high temperatures (Dachille *et al.*, 1963; Skinner and Fahey, 1963).

The Hugoniot curve for material, say, quartz, undergoing a shock-induced phase change consists of three regimes: the low-pressure, the mixed-phase and the high-pressure regimes (Ahrens *et al.*, 1969a, Fig. 1). Quartz, shocked to the mixed-phase regime (about 100—380 kb), gradually transforms to a high-pressure phase, i.e. to stishovite or a dense glass with Si in sixfold coordination, or both. The formation of a mixture like quartz and coesite, or stishovite and coesite, is also possible. According to this model, at least some stishovite or coesite is then retained upon the pressure release (Wackerle, 1962, p. 937; McQueen *et. al.*, 1963; p. 2322; De Carli and Milton, 1965; Ahrens and Rosenberg, 1968, pp. 71—75).

Stöffler (1971b) studied a number of coesite- and stishovite-bearing crystalline rocks from the Ries Crater in Germany. Coesite usually occurs as very fine-grained aggregates embedded in diaplectic quartz glass and as aggregates arranged along planes in diaplectic quartz. Coesite-bearing diaplectic quartz glass always has a very low birefringence and mean refractive indices below 1.480. Stishovite, however, is generally enclosed within filled lamellae of diaplectic quartz, but it is possible that traces of stishovite also occur in diaplectic quartz glass. Stöffler (1971b, pp. 5481—5482) concluded that in nonporous rocks stishovite forms during the shock compression, whereas coesite crystallizes during the pressure release and that even so a metastable growth of coesite may continue after the pressure release. He suggested that peak pressures ranging from about 300 kb to about 550 kb are required for the formation of coesite and that pressures ranging from about 120 kb to about 450 kb are required for the formation and preservation of stishovite.

Kieffer (1971), who studied a suite of shocked sandstone specimens from Meteor Crater, Arizona, noted that the behavior of porous material under shock conditions differed markedly from that of nonporous or single-crystal material. There is a strong reverberation of shock and rarefaction waves in porous rocks, which results in a highly nonuniform pressure and temperature distribution. Locally, both high shear pressures and peak pressures occur at grain boundaries and along planes within crystals, speeding up the sluggish quartz-coesite transition. Kieffer (1971, p. 5469) observed a little coesite in regions believed to have been subjected to high shearing stress. Weaker shock waves can thus produce coesite and stishovite in porous rocks. Another difference is due to varying temperature distribution in shocked porous rocks in which stishovite may be

inverted, not to diaplectic glass as suggested by v. Engelhardt and Bertsch (1969, p. 231) for nonporous material, but to coesite and fused silica glass (lechatelierite) (Kieffer, 1971, pp. 5470—5471).

Coesite and stishovite are not only important indicatory minerals of shock metamorphism but also very sensitive indicators of post-shock thermal conditions. Stöffler (1971b, pp. 5484—5486) pointed out that in rocks in which shocked rock fragments are incorporated in rock melt (cf. kärnäite) all the stishovite and coesite decompose during cooling. Suevite-like breccias (fallout and fallback breccias) contain rapidly quenched material, but the bulk temperature of the rock after its deposition may still be several hundred degrees C. This causes annealing during which the stishovite is wholly destroyed and the amount of coesite decreases (see also Dachille *et al.,* 1963; Skinner and Fahey, 1963; Gigl and Dachille 1968).

With the exception of kimberlites (Harris, 1968, p. 405), tektites and fulgurites (Reid and Cohen, 1962; Walter, 1965; Stöffler and Arndt, 1969), coesite has been found in nature solely in meteorite craters and in cryptoexplosion craters. The following coesite localities are recorded in the literature: Meteor Crater in Arizona, U.S.A. (Chao *et al.,* 1960), Ries Crater in Germany (Pecora, 1960, p. 19; Shoemaker and Chao, 1961), Wabar Crater in Arabia (Chao *et al.,* 1961), Lake Bosumtwi Crater in Ghana (Littler *et al.,* 1961), Serpent Mound in Ohio and Kentland Structure in Indiana, U.S.A. (Cohen *et al.,* 1961), Holleford Crater in Canada (Bunch and Cohen, 1963), Lake Mien in Sweden (Svensson and Wickman, 1965) and Lake Wanapitei in Canada (Dence *et al.,* 1974).

The identification of coesite is, however, not entirely conclusive in instances where it is based merely on an X-ray powder pattern with relatively weak line intensities. The powder pattern of coesite displays only two strong lines and, therefore, in the opinion of the present author, the identification must be verified by means of the infrared absorption spectrum. An X-ray powder pattern with weak lines may lead to misidentification as actually happened in the case of the Richat Dome, Mauritania (Cailleux *et al.,* 1964; Fudali, 1970).

In the beginning of this study, coesite was enriched for X-ray identification in the usual way (Chao *et al.,* 1961; Bunch and Cohen, 1963), viz. one gram of the rock was powered and dissolved in a water solution of 5 % hydrofluoric acid and 5 % nitric acid. The treatment was continued until the undissolved residue was less than 5 % of the original material. An X-ray diffractogram of the residue was recorded on a Philips Norelco diffractometer. Several samples of the Lappajärvi rocks were treated by this method. Very weak peaks of coesite were observed in diffractograms of only two samples, and it was concluded that the method was not suitable for the Lappajärvi rocks. Therefore the pre-enrichment procedure was modified as follows. About 50 g of very finely powdered sample

was eluted for one hour in a 600 ml beaker. The suspension, which usually contained 1 g to 3 g of the most fine-grained material, was transferred into a 600 ml plastic beaker. Hydrofluoric acid and nitric acid were added to the solution until the concentration of both was about 5 %. The suspension was allowed to stand for one day at room temperature. The residue was then filtered off and washed, and an X-ray diffractogram was recorded. When the undissolved residue was very small (it was embedded in the filter paper), the filter paper with the residue was slowly heated in a platinum crucible to burn off the paper. This procedure is possible because dry coesite is stable at atmospheric pressure and high temperature (Dachille *et al.,* 1963). If necessary, the acid treatment was continued until no quartz lines remained on the X-ray diffractogram.

The elution of the material before acid treatment has the following advantages:

1) The grain size of the material is uniform and thus facilitates a more selective solution.

2) Because the grain size of coesite is exceedingly small (see Fahey, 1964, p. 1646; Kieffer, 1971, p. 5459; Stöffler, 1971b, p. 5475), coesite may be considerably enriched during elution.

3) Most Lappajärvi rocks contain small amounts of graphite. This mineral appears to float in water and can be removed before acid treatment.

4) Slightly soluble heavy minerals such as zircon are largely enriched in the coarse-grained and heavy fraction that settles on the bottom and do not enter the acid treatment.

Almost 70 specimens, or one fourth of the rock specimens collected, were treated by the method given. It was found that the kärnäite contains no coesite, which is consistent with Stöffler's (1971b) findings, and indicates that the kärnäite was formed in an environment where the post-shock temperature was high or the temperature release was slow enough to allow the possible high-pressure polymorphs to transform into low-pressure polymorphs or silica glass. Coesite was detected in seven samples, six of which are suevite samples and one of which is a granite fragment embedded in a coarse impact breccia.

Six of the seven coesite-bearing Lappajärvi rock specimens contain only minute amounts of coesite, and the X-ray diffractograms obtained from the enriched residues display only the two strongest peaks of coesite. In the seventh specimen (suevite, No. 236), however, the coesite content was found to be rather high (about 0.25 %). Because the coesite-bearing concentrate obtained was contaminated by spinel, it was difficult to accurately determine the coesite content. This specimen consists of brown or greenish brown glass with fluidal texture containing relict quartz grains, and of fine-grained bedded breccia material. Graphite is relatively abundant (about 1 %). Coesite occurs most probably in

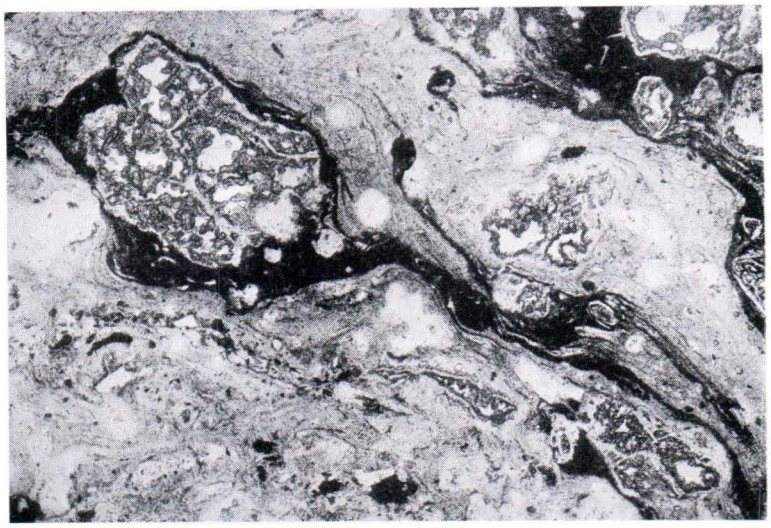

Fig. 13. A quartz grain with aggregates (gray) of diaplectic quartz glass and coesite (cf., Fig. 12 of Kieffer, 1971, and Fig. 1 of Stöffler, 1971b). The grain is embedded in light brown glass with schlieren, vesicles and mineral fragments. Suevite, Specimen No. 236. Plane-polarized light, 100 X.

connection with the relict quartz grains. An X-ray powder pattern of a piece of a quartz grain showed that the relict quartz consists mostly of quartz (chalcedony), but the two strongest peaks of coesite are also visible. This specimen resembles specimens Nos. 601 and 620 from the Ries Crater made available for reference by the courtesy of Professor Th. G. Sahama, University of Helsinki. In the Ries specimens fluidal textures are visible in the brownish glass displaying embedded pieces of diaplectic quartz glass and diaplectic quartz grains, both of which contain coesite (v. Engelhardt, 1967a, pp. 181—184, Figs. 14—15). At Lappajärvi, the silica inclusions have partly recrystallized to very finely crystalline quartz and chalcedony. The grain size of coesite in the Lappajärvi rocks is too fine to allow its microscopic detection in thin section.

The X-ray diffraction powder pattern of coesite from specimen No. 236 is reproduced in Table 3. For the stronger lines the accuracy in the measurement of the 2Θ-angles is $\pm$ 0.01° and for the weak lines (relative intensity less than 6) $\pm$ 0.04°. The lines caused by contaminating graphite (d = 3.360 Å) and zircon (d = 3.302 Å), and the seven lines produced by spinel have been omitted. The spinel lines are listed in Table 8.

The powder pattern of coesite from specimen No. 236 agrees well with that of the synthetic coesite given by Boyd and England (1960, Table 1) and by Keat (ASTM Card 14-654). In addition, the X-ray powder pattern recorded from pure coesite from Meteor Crater, Arizona, which was available for

Table 3

X-ray powder pattern of Lappajärvi coesite

I	*hkl*	2Θ	dmeas.	dcalc.
2	020	14.32	6.186	6.184
1	021	20.23	4.389	4.377
35	130 111	26.00	3.426	3.429 3.426
100	$\bar{2}02$ 002 $\bar{2}21$ 040	28.83	3.096	3.105 3.097 3.095 3.092
9	220	32.40	2.763	2.762
11	131	33.19	2.699	2.697
4	$\bar{2}41$	38.51	2.338	2.338
6	$\bar{3}11$ 150 112	39.22§	2.297	2.296 2.296 2.296
7	$\bar{2}42$ 240	41.28	2.187	2.191 2.185
4	151	44.60	2.032	2.032
4	330	49.52	1.8405	1.8412
7	$\bar{2}04$	50.90	1.7938	1.7931
3	$\bar{2}61$	51.14b	1.7860	1.7857
10	062 260	53.46	1.7138	1.7161 1.7143
7	113 170	53.98	1.6989	1.6990 1.6987
6	$\bar{4}23$	55.46	1.6568	1.6572
4	350	58.24	1.5841	1.5820
4	$\bar{4}04$	59.57	1.5518	1.5524
1	$\bar{1}73$ 153 172	a66.28b	1.4095	1.4114 1.4096 1.4093
9	$\bar{4}62$ 262 $\bar{1}15$ $\bar{5}32$ 190	a69.98b	1.3432	1.3503 1.3484 1.3454 1.3419 1.3414
1	$\bar{4}25$	a71.15	1.3240	1.3243
4	$\bar{5}34$ $\bar{1}35$	a73.67	1.2848	1.2862 1.2859

Cu/Ni radiation, internal silicon standard

a = measured as α_1-peak
b = broad peak
§ = strong interference from adjacent peak of spinel
Calculated unit-cell parameters:

$$\begin{cases} a_0 = 7.148 \pm 0.005 \text{ Å} \\ b_0 = 12.369 \pm 0.008 \\ c_0 = 7.173 \pm 0.005 \\ \beta = 120°16 \pm 05' \end{cases}$$

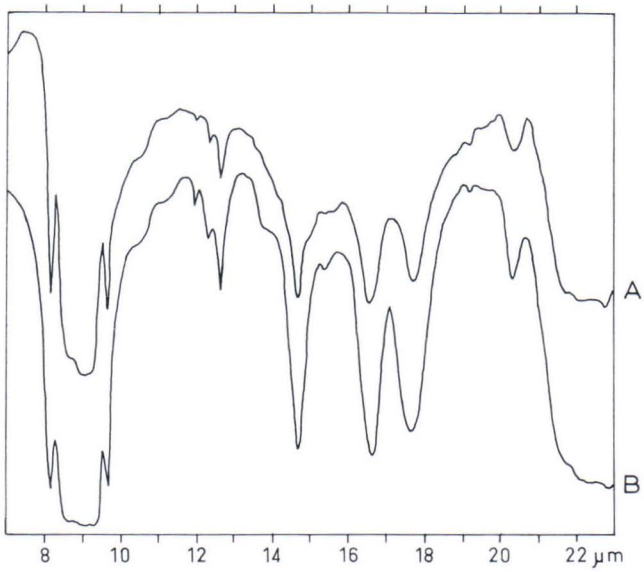

Fig. 14. Infrared absorption spectrograms of coesite from Lake Lappajärvi (Specimen No. 236, tracing A) and from Meteor Crater in Arizona (tracing B).

reference by the courtesy of Dr. E. C. T. Chao, U.S. Geological Survey, Washington, D.C., is virtually identical with that of coesite No. 236 from Lappajärvi.

The occurrence of coesite in specimen No. 236 from Lappajärvi was verified by recording the infrared absorption spectrum on a Leitz Douple Beam Infrared Spectrophotometer using NaCl and KBr prisms; wavelength range 1 μm to 24 μm. Details of the method have been given previously by Sahama and Lehtinen (1968, pp. 32—33). The spectrum of the Meteor Crater coesite was also recorded for reference. The photometer tracings of both coesite spectra are reproduced in Fig. 14. Because of the scarcity of the Meteor Crater material available, the potassium bromide pellet prepared from this sample contained only 0.265 weight % coesite instead of the 0.30 weight % used for the Lappajärvi coesite. This slight difference in content between the potassium bromide pellets causes some difference in the intensities of the bands. For the purpose of identification, however, the difference is of no importance.

The infrared absorption spectra of the Meteor Crater and Lappajärvi coesite samples are virtually identical. In the spectrum of the Lappajärvi coesite, the background between the bands shows a lower transparency than in the Meteor Crater spectrum, a difference apparently due to spinel contamination in the Lappajärvi sample. In the range from 9 μm to 24 μm the infrared absorption spectrum of spinel displays a low transparency without any sharp bands. Conse-

quently, the spinel contamination does not interfere with the coesite bands but merely raises the background.

Feldspars

The unusual properties of some meteoritic and terrestrial glasses of plagioclase composition have long been known, but the work of De Carli and his co-workers (De Carli and Jamieson, 1959; Milton and De Carli, 1963) demonstrated that the glasses can form by shock-induced solid-state transformations. Since their experiments on minerals and rock specimens, new recovery shock experiments, including both those on a laboratory-scale and underground nuclear and chemical explosions, have been carried out. More information on the behavior of feldspars under shock conditions has been obtained by means of Hugoniot and release adiabat experiments. The experimental data have been applied to naturally shocked feldspars of terrestrial impact craters and meteorites, and recently to those of returned lunar specimens.

The Hugoniot elastic limit (or HEL) of the feldspars appears to lie within the limits of 40 kb to 90 kb (Ahrens *et al.,* 1969b, pp. 2732—2736; Ahrens and Liu, 1973, p. 1275). Above the HEL, the mixed-phase regime of microcline and othoclase extends from 115 kb—120 kb to about 300 kb. In oligoclase, this region of anomalously high compression ranges from about 115 kb to about 400 kb. The very high compression is caused by the partial transformation into a high-pressure phase with a zero-pressure density of about 3.8 g/cm^3. This phase is most probably of hollandite structure (Ahrens and Liu, 1973, p. 1277; Ringwood *et al.,* 1967). The possibility of the formation of high-pressure phases in the structures of jadeite and stishovite also exists, and James (1969) observed a mixture of jadeite and dense plagioclase glass, maskelynite, in several strongly shocked specimens from the Ries Crater, Germany.

Shock pressures exceeding the upper limit of the mixed-phase regime are needed for a complete transformation into a high-pressure phase. Upon pressure release, however, the high-pressure phase, or phases, will usually revert to a low-pressure amorphous phase. Depending on the post-shock temperature, the amorphous phase is formed either in a solid-state transformation resulting in a dense glass (SRO phase, diaplectic glass or thetomorphic glass) or, if the post-shock temperature exceeds the transition temperature of the glass, the amorphous phase consists of normal liquid-state glass (fused, vesicular mineral glass). Stöffler and Hornemann (1972) studied and summarized the physical properties (density, refractive index, infrared absorption) of both of these feldspar glasses. They confirmed that diaplectic feldspar glasses represent a series of intermediate structural states between shocked but still birefringent feldspars and fused glasses of the same composition. Annealing experiments on terrestrial, meteoritic and

lunar maskelynite (Bunch *et al.,* 1967, 1968; Duke, 1968; Bell and Chao, 1970; Chao *et al.,* 1970, p. 301; v. Engelhardt *et al.,* 1970, p. 372) indicate that the physical properties of the dense glasses reflect their pressure-temperature histories.

Disordered but still birefringent feldspars are produced if the shock pressures are within the mixed-phase regime. Such shocked feldspars are characterized by closely spaced planar deformation structures, and, as in quartz, the planar elements or shock lamellae consist of highly disordered or amorphous material. Consequently, with an increasing number of planar elements, the feldspars gradually approach the diaplectic or thetomorphic feldspar glasses. Stöffler (1967) and Dworak (1969) have given detailed descriptions of shocked plagioclase specimens from the Ries Crater and from the Manicouagan Crater, Canada. In addition to maskelynite, they found diaplectic plagioclase specimens with several deformation structures (planar fractures, planar elements, deformation bands). They also observed, as did Chao (1967b, pp. 196—197), that the planar elements in plagioclase prefer to develop parallel to crystal planes with low crystallographic indices. Papers on shocked potassium feldspar are few (see Aitken and Gold, 1968; Kleeman, 1971; Stöffler and Hornemann, 1972; Ahrens and Liu, 1973), and the author has not found any detailed investigation of naturally shocked microcline.

Alkali feldspars

Microscopic investigations

1) In unshocked Lappajärvi rocks potassium feldspar is bluish white or reddish. In a hand specimen, the distinction is not always readily made between potassium feldspar and plagioclase; however, the potassium feldspar is fresh and the plagioclase is usually dull owing to secondary alteration (sericitization). In thin sections made of granite and granite pegmatite specimens, potassium feldspar always has the cross-hatched pattern typical of microcline, but in some granodiorite specimens collected from the southeastern corner of the area untwinned microcline has been found. Stringlets, strings and rods of microperthite are common in granite and granite pegmatite.

2) In weakly and moderately shocked rocks (Shock Stages I and II) potassium feldspar is often dull bluish, and its color grows darker with increasing shock intensity. Under the microscope, a number of shock-induced deformations and changes (planar deformation structures, bending, complex mosaic texture) are observed.

Planar fractures are usually thin and develop preferably parallel to certain simple crystal planes, such as (110), ($1\bar{1}0$), (001) and ($1\bar{2}0$). A zig-zag or

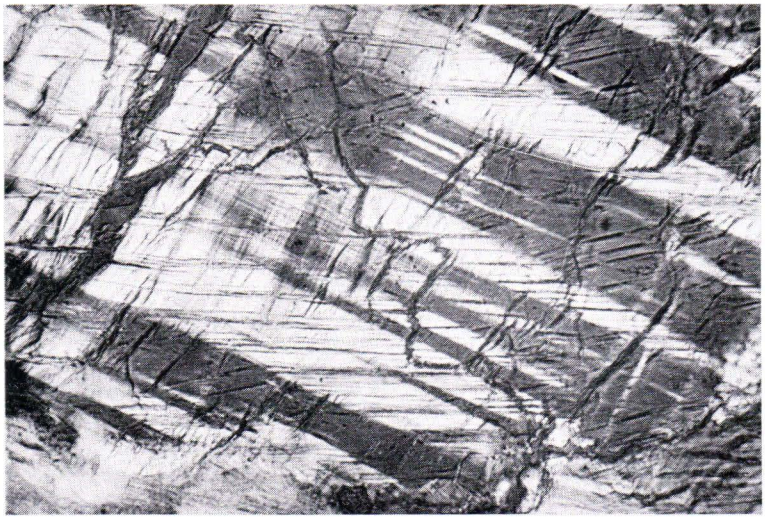

Fig. 15. Fractured cross-hatched microcline with non-decorated planar elements parallel to (120) (nearly horizontal set). Granite fragment of impact breccia, Specimen No. 215. Almost crossed polarizers, 230 X.

chevron pattern of fractures is often seen in cross-hatched twinnings. In addition to planar fractures, closely spaced and very short non-decorated planar elements (shock lamellae) are common in the Lappajärvi microcline. They are best developed in grains or areas with reduced birefringence. Owing to the intimate twinning of microcline it was possible in only a few cases to determine the crystallographic orientation of the planar elements. The elements are crystallographically controlled because they have identical orientation in those individuals of a complex twin crystal that have the same optical orientation. The number of distinct sets of planar elements per grain is small (1—3). In some grains with a relatively coarse cross-hatched pattern planar elements parallel to (110), $(1\bar{1}0)$, (120), $(1\bar{2}0)$ and $(1\bar{4}0)$ were observed. Elements parallel to some (hkl) pinacoids may also occur. In regions where albite twinning dominates, one set of twin lamellae has planar elements parallel to (120), and the other set parallel to $(1\bar{4}0)$, and thus a stair-like pattern of planar elements is formed. In more strongly shocked microcline, plastic deformation (incipient flow?) and deformation bands with diffuse boundaries are visible, especially in large grains. In such microcline an almost chaotic texture has been produced and prevents a detailed optical investigation.

In coarse- or medium-grained granite fragments of the impact breccia or suevite, microcline with planar deformation structures is associated with oligoclase that has been at least partly transformed into maskelynite. The adjacent

quartz grains always have numerous planar elements, but quartz grains that have been shocked to make diaplectic glass do not occur. The biotite is fresh but full of kink bands. Diaplectic microcline glass is rare in the Lappajärvi rocks, but it should be noted that if diaplectic glasses of microcline and plagioclase occur together it may be impossible to distinguish between them (Stähle, 1973, p. 385).

3) In strongly (Shock Stage III) and intensely shocked (Shock Stages IV and V) Lappajärvi rocks alkali feldspars have recrystallized as vesicular grains or spherules of sanidine and anorthoclase. There are, however, a few strongly shocked specimens in which only incipient flow or vesiculation of potassium feldspar is observed, e.g. specimen No. 132, which is a piece of coarse-grained granite characterized by yellowish potassium feldspar full of small open fractures, and by clear or milky translucent quartz. The plagioclase exhibits secondary alteration. In thin section, no vesiculation of the potassium feldspar is visible but the feldspar bears witness to small-scale flow and contains long deformation bands. This suggests that the post-shock temperature barely exceeded the softening point of the corresponding feldspar glass (Stöffler and Hornemann, 1972, pp. 388—389). Most grains have a strong zonal extinction of about 20° to 30° and an anomalously blue interference color. Their *2V* is small and negative. Remnants of deformed cross-hatched pattern are observed, as are also thin, bent shock lamellae. The potassium feldspar grains are permeated with sets of cleavages (planar fractures), but irregular, even radial fractures are also present. The adjacent quartz is largely recrystallized, but traces of penetrative planar fractures and patches with planar elements are still visible. The biotite is faded brown and intensely kinked and has a little reduced birefringence and pleochroism. In this specimen, the plagioclase has been zeolitized into heulandite, but some unaltered maskelynite is still manifest. The original texture of the rock is unchanged.

In strongly and intensely shocked Lappajärvi rocks the post-shock temperature has usually reached the melting point of the feldspar. The material has been vesiculated and homogenized, and the feldspar now present is high sanidine or anorthoclase. The principal cause of the vesiculation must be the outgassing of adsorbed water, and silicate vaporization played only a minor role.

Specimens Nos. 55 and 403 were originally graphic granite. The feldspar now consists of spherulitic sanidine (Specimen No. 55, Or 78 weight %) or anorthoclase (Specimen No. 403, Or 6 weight %). The largest spherules and cavities occur near the quartz-feldspar contact. The quartz has wholly recrystallized into a mosaic of tiny crystals, partly chalcedony, and in Specimen No. 403 the quartz is brownish. The gross texture is largely unchanged in these specimens, but in some others the original texture has been destroyed. Specimen No. 55 is from suevite and Specimen No. 403 is from Type I kärnäite.

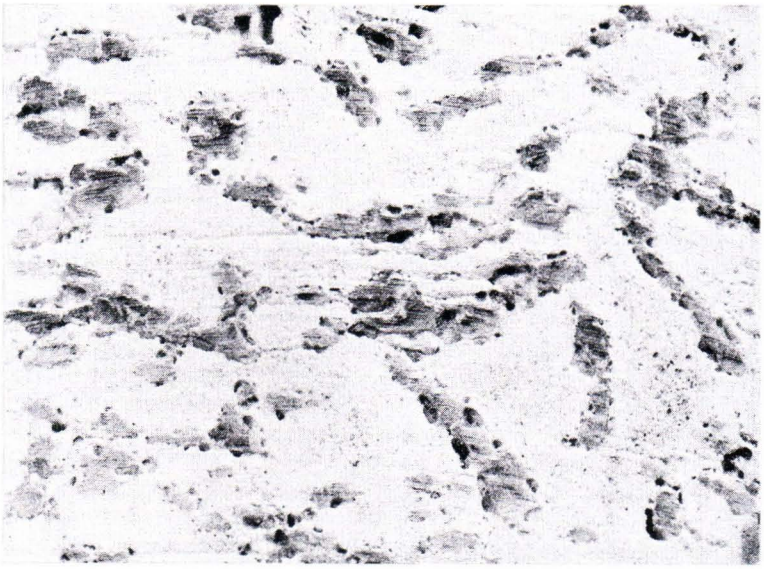

Fig. 16. Vesiculated and recrystallized graphic granite. Specimen No. 55, 2.5 X.

X-ray diffractograms

X-ray diffractograms were recorded for a large number of feldspar specimens from different stages of shock metamorphism on a wide-angle Philips diffractometer by use of Ni filtered Cu radiation and an internal silicon or quartz standard. The usual range of measurement was from $2\Theta = 60°$ to $2\Theta = 15°$. The diffractograms were measured and corrected, and reflections of unambiguous indexing were employed to calculate provisional unit-cell parameters. The remaining reflections were either indexed or their indexing was checked by using the provisional parameters, and the refined unit-cell parameters were calculated by the least-squares method, all 2Θ values being given unit weight. All the calculations were performed on a Hewlett-Packard calculator, Model 9100B, equipped with the extended memory 9101A.

The existence of three different groups of alkali feldspars in the Lappajärvi rocks was verified by X-ray diffractograms (see Table 4), as follows.

Group I: unshocked off-structure specimens. They are characterized by sharp peaks on the diffractograms and by highly ordered structural state.

Group II: weakly and moderately shocked specimens. They yield diffractograms with broad peaks and reduced peak intensities. The specimens have highly variable structural states.

Table 4

c^*/b^* ratios, Δbc values and Or contents of the alkali feldspar specimens investigated

a No.	b	c^*/b^*	Δbc	form	Or wt %
Unshocked (off-structure) specimens (Group I):					
16	1	1.996	0.984	tr	99
128b	2	1.996	0.944	tr	99
151	3	1.997	0.921	tr	98
249	4	1.996	0.945	tr	97
264	5	1.997	0.935	tr	99
402a(1)	6	1.995	0.996	tr	96
402a(2)	7	1.996	0.992	tr	94
Weakly and moderately shocked specimens (Group II):					
54 (1)	8	2.003	0.82	tr	98
54 (2)	9	2.013	0.57	m	94
93 (1)	10	1.997	1.0	tr	95
93 (2)	11	1.999	0.87	tr	99
93 (3)	12	1.998	0.92	tr	90
105	13	2.006	0.73	m	91
131	14	2.010	0.66	m	87
134 (1)	15	1.998	0.99	tr	99
134 (2)	16	2.009	0.61	m	91
215	17	1.999	0.94	tr	98
250 (1)	18	2.003	0.78	m	91
250 (2)	19	2.004	0.73	m	92
253	20	2.000	0.84	tr	96
409	21	2.005	0.94	tr	98
410	22	2.008	0.59	m	80
453	23	2.001	0.84	tr	90
Strongly and intensely shocked specimens (Group III):					
55	24	2.019	0.52	m	78
60	25	2.020	0.48	m	72
65	26	2.022	0.46	m	67
85	—	2.020	0.50	m	71
91	27	2.0195	0.53	m	69
104	28	2.018	0.47	m	84
110	29	2.021	0.50	m	68
111	—	2.023	0.48	m	79
125	—	2.021	0.51	tr	14
132	30	2.021	0.47	m	85
206	31	2.019	0.53	m	73
244	32	2.0205	0.49	m	72
246	33	2.021	0.48	m	78
247	34	2.021	0.49	m	82
252	35	2.021	0.52	m	82
400	36	2.017	0.55	m	73
403	—	2.019	0.54	tr	6

a: original specimen number (numbers enclosed in parentheses refer to separate single grains of the same specimen)
b: number used for the specimen in Figs. 17—19
tr = triclinic, m = monoclinic form

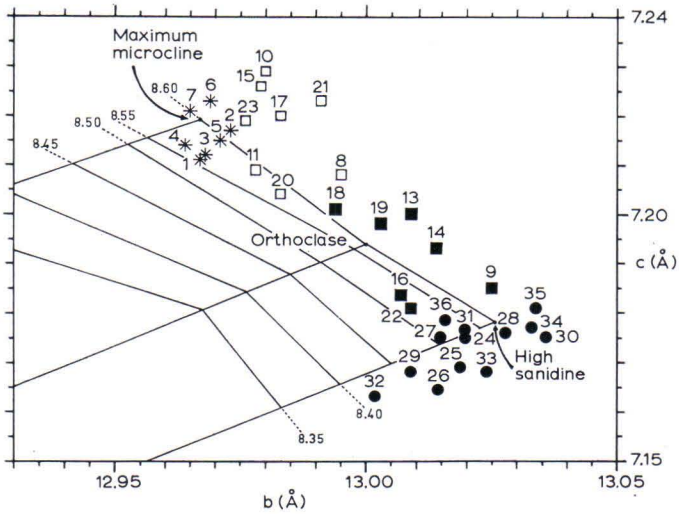

Fig. 17. *b—c* plot of potassium feldspars. Contours according to Wright and Stewart (1968, Fig. 2b). Specimen numbers, see Table 4.

Stars: Unshocked. *Open squares:* Weakly and moderately shocked, dominating form triclinic. *Solid squares:* Weakly and moderately shocked, dominating form monoclinic. *Solid circles:* Strongly and intensely shocked.

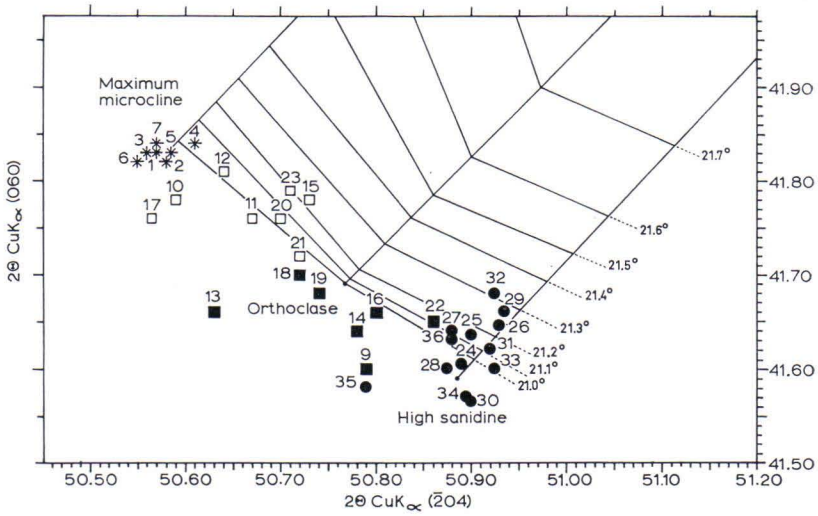

Fig. 18. Observed 2Θ values of 060 plotted against observed 2Θ values of 204 for potassium feldspars. Contours according to Wright (1968, Fig. 3). For symbols and numbers, see Fig. 17 and Table 4.

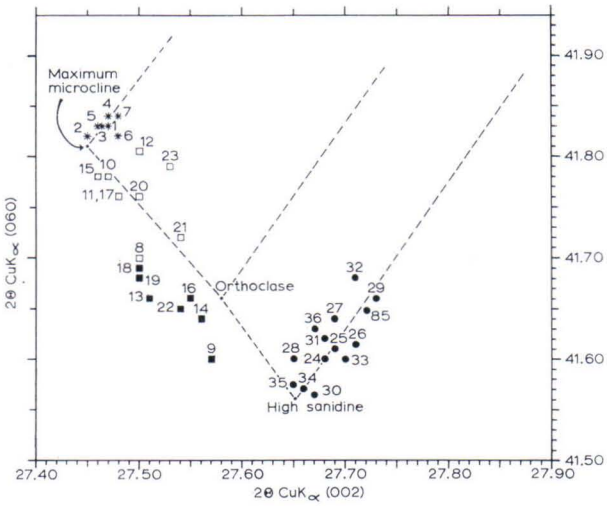

Fig. 19. Observed 2Θ values of 060 plotted against observed 2Θ values of 002 for potassium feldspars. Contours and dots for maximum microcline (Or 98), orthoclase (Or 90) and high sanidine (Or 91) have been drawn according to the calculated powder patterns of Borg and Smith (1969). For symbols and specimen numbers, see Fig. 17 and Table 4.

Group III: strongly and intensely shocked specimens. They are characterized by the uniform structural state of a completely random Al/Si distribution. Most specimens yield diffractograms with very sharp peaks. There is a gradual change especially between Group II and Group III, and it appears to be reasonable to use $c^*/b^* = 2.016$ (Table 4) as the value dividing the two groups (see Jones, 1966). The following is a more detailed description of each group.

Group I consists of single grains of microcline from seven unshocked granite and granite pegmatite specimens collected from outcrops around Lake Lappajärvi. The Or content of these microcline-perthite specimens, obtained from the calculated unit-cell volumes (Orville, 1967, Fig. 5; Wright and Stewart, 1968, Fig. 1a), averages 97 weight %. All the specimens X-rayed display the two or three strongest peaks of albite. In the *b-c* plot (Fig. 17), all the points representing these microcline specimens are located near the corner of maximum microcline, which is indicative of a highly ordered structural state (between Spencer B and maximum microcline in Fig. 2b of Wright and Stewart, 1968). Their Δbc values, which were obtained in the manner suggested by Stewart and Ribbe (1969, p. 446), are close to 1.0. The high degree of Al/Si order is also visible in the plot of 2Θ (060) versus 2Θ ($\bar{2}04$) (Fig. 18) and in the plot of 2Θ (060) versus 2Θ(002) (Fig. 19).

Table 5

Triclinicity of potassium feldspar

a	No. b	1	2
	Unshocked (off-structure) specimens:		
16	1	0.89	0.91
128b	2	0.94	0.94
151	3	0.94	0.96
249	4	0.96	0.98
264	5	0.94	0.95
402a(1)	6	0.975	0.99
402a(2)	7	0.925	1.01
	Weakly shocked specimens:		
93 (1)	10	0.81	0.83
93 (2)	11	0.70 d	0.60
134 (1)	15	0.79 d	0.80
215	17	0.84	0.86
253	20	0.825d	0.84
409	21	0.36 d	0.67
453	23	0.60 d	0.82

a: original specimen number (numbers enclosed in parentheses refer to separate single grains of the same specimen).

b: number used for the specimen in Figs. 17—19.

1: triclinicity calculated from equation $\Delta = 12.5\,[d(131)—d(1\bar{3}1)]$ (Goldsmith and Laves, 1954, p. 5).

2: triclinicity obtained plotting α against γ (Orville, 1967, Fig. 13).

d: diffuse separation of the 131 and $1\bar{3}1$ peaks.

More information about the structural state was obtained by calculating the triclinicity as proposed by Goldsmith and Laves (1954, p. 5) and by plotting α against γ as outline by Orville (1967, Fig. 13). A good correlation exists between the values obtained by the two methods, except in the case of specimen No. 402a(2) for which the latter method gives a value clearly too high (Table 5).

The X-ray diffractograms indicate that the pre-shock structural state of the potassium feldspar (microcline-perthite) in the Lappajärvi target rocks has been ordered and between that of Spencer B and maximum microcline.

Group II comprises weakly or moderately shocked (Shock Stages I or II) microcline specimens. Their X-ray diffractograms are characterized by broad peaks and reduced peak intensities, which makes it difficult to obtain a sufficient number of unambiguously indexed reflections for the calculation of unit-cell parameters. Therefore, and because of the broadening of the peaks, the parameters calculated are not very reliable. The broadening of the peaks is the result of the destruction of long-range order in the crystal lattice (development of diaplectic or thetomorphic glass) and of the development of a monoclinic form of potassium feldspar. Thus, pairs of peaks like 130—$\bar{1}30$ and 131—$1\bar{3}1$ are

not distinctly separated, or the corresponding peak of orthoclase is visible between the two peaks. There are also single grains that yield diffractograms without any peaks, or with only the two or three strongest peaks visible. It is impossible to obtain unit-cell parameters for such specimens, and even the method of Wright (1968), in which the 2Θ values for the 060 and $\bar{2}04$ reflections are plotted against each other (Fig. 18), is not applicable.

The best way to classify weakly and moderately shocked microcline specimens (diaplectic microcline) appears to be to calculate the c^*/b^* ratio for each specimen by using the 060 and 002 reflections (Table 4; Fig. 19). The method yields reasonable results even when only the 002, 060 and, as is usual, the $\bar{2}01$ peaks are measurable.

The X-ray diffractograms prove the occurrence of significant shock-induced Al/Si disordering in the microcline samples studied. It is manifest from Table 4 that the c^*/b^* ratio of the weakly and moderately shocked microcline specimens ranges from 1.997 to 2.013, the mean and standard deviation being 2.003 $\pm$ 0.005. For the unshocked microcline specimens the ratio is 1.996 $\pm$ 0.001, and for the strongly and intensely shocked specimens it is 2.020 $\pm$ 0.001. This demonstrates the structural uniformity of Groups I and III.

A conspicuous difference between Groups I and II is apparent when the calculated values of triclinicity are compared with one another (Table 5). Weakly shocked microcline samples have a reduced triclinicity and not all their peaks are distinctly separated. For specimens slightly more strongly shocked, the calculation of triclinicity is impossible or unreliable because they are almost amorphous to X-rays or the crystalline material is mostly monoclinic.

The estimated Or contents of the weakly and moderately shocked microcline specimens (Table 4) were obtained by using the $\bar{2}01$ reflection and thus they are not very reliable (Wright, 1968, p. 97). In addition, some of these specimens have an anomalously large unit cell, even 2 Å^3 to 3 Å^3 larger than the unit cell of the microcline or anorthoclase end member. In all such instances, the dominating crystalline phase is monoclinic (orthoclase), and the lowered Or content indicates incipient homogenization of the microcline-perthite.

The heterogeneous nature of the weakly and moderately shocked microcline specimens (Group II) is likewise seen in Figs. 17—19. Such specimens form an intermediate group between the unshocked (Group I) and strongly and intensely shocked (Group III) Lappajärvi feldspars. In particular, specimens in which the dominating crystalline phase is monoclinic (solid squares in Figs. 17—19) lie out of the contours for potassium feldspar of intermediate structural state. The behavior of these specimens is a result of their enlarged unit-cell size. The cell is evidently thermally expanded and thus is an indicator of the post-shock temperature from which it was quenched. However, the observed enlargements (about 3 Å^3 to 5 Å^3) would at atmospheric pressure be caused by

quenching from only relatively low temperatures (200°C to 450°C; Skinner, 1966, Table 6-1).

The strongly and intensely shocked feldspar samples (Group III) were obtained from kärnäite and suevite. This group is structurally uniform (Table 4; Figs. 17—19), and the structural state corresponds to that of the high sanidine — high albite series (Wright and Stewart, 1968). The Or contents of these specimens were calculated from their unit-cell volumes and, in general, prove that the entire feldspar material is homogeneous. However, the two or three strongest peaks of high albite or anorthoclase are visible in diffractograms of several sanidine specimens. Specimens in which the feldspar has recrystallized as anorthoclase are not common. Two of them (Nos. 125 and 403) are included in Table 4.

Microscopic determinations indicate that Group III may consist of two subgroups because it contains a few specimens in which only incipient flow or vesiculation is visible. Some of the potassium feldspar grains from specimens Nos. 111 and 132 actually yielded X-ray diffractograms with somewhat broader peaks and reduced peak intensities, but the existence of the two subgroups could not be verified by means of X-rays because the calculated c^*/b^* ratios correspond to those of high sanidine samples (Table 4).

Infrared absorption

Infrared absorption spectra were recorded for a number of alkali feldspar specimens on a Leitz Double Beam Spectrophotometer. The experimental procedure has been described earlier by the present author (Lehtinen, 1974, p. 225). The Lappajärvi specimens used for this study ranged from unshocked (nearly maximum microcline) to intensely shocked (high sanidine). Their structural states were estimated by means of the c^*/b^* ratio as described in the previous chapter (see Table 4).

Both the unshocked (Group I) and the strongly and intensely shocked (Group III) microcline specimens yielded infrared absorption spectra that are typical of maximum microcline and high sanidine, respectively. The weakly and moderately shocked (Group II) Lappajärvi microcline yielded infrared spectra with weak band intensities. This is particularly obvious within the range of 12.5 μm to 20 μm (800 cm^{-1} to 500 cm^{-1}) where the variation of transparency is only 20 % to 30 %. Stöffler and Hornemann (1972, Fig. 9) observed similar features in the infrared spectra of diaplectic alkali feldspar glasses of naturally and experimentally shocked samples. The corresponding variation in the spectra of high sanidine, orthoclase or microcline is 50 % to 70 %. According to Iiishi *et al.* (1971, Table 3), the absorption bands within the forementioned

Table 6

Or content (weight %), positions of the characteristic infrared absorption bands (A and B), $\Delta\mu$ values (wave lengths in μm), and c^*/b^* ratios of unshocked and shocked alkali feldspar specimens

No.	Or	A	B	B—A = $\Delta\mu$	c^*/b^*
Unshocked specimens (Group I):					
128b	99	15.43	18.60	3.17	1.996
264	99	15.44	18.59	3.15	1.997
Weakly and moderately shocked specimens (Group II):					
93 (3)	90	15.48	18.51	3.03	1.998
97	99	15.50	18.54	3.04	1.999
105	91	15.61	18.34	2.73	2.006
131 (2)	99	15.67	18.26	2.59	2.012
250 (2)	92	15.68	18.34	2.66	2.004
Strongly and intensely shocked specimens (Group III):					
55	78	15.72	18.24	2.52	2.019
111	79	15.77	18.18	2.41	2.023
125	14	15.76	18.33	2.57	2.021
246	78	15.685	18.25	2.565	2.021
403	6	15.75	18.38	2.63	2.019

range result principally from Si-Si and Si-Al(Si) stretching, and from O-Si-O and O-Si(Al)-O bending vibrations. The two strong bands or band groups at about 9 μm and 10 μm, which are due to Si-O and Si(Al)-O stretching vibrations, are only somewhat rounded.

The positions of the characteristic absorption bands of Lappajärvi alkali feldspars and the c^*/b^* ratios are listed in Table 6. The common (unshocked) potassium feldspars exhibit a good linear correlation between the c^*/b^* ratio and the positions of the characteristic absorption bands (Lehtinen, 1974, p. 229). This holds true also for the unshocked (Group I) and strongly and intensely shocked (Group III) Lappajärvi specimens (Fig. 20). The weakly and moderately shocked specimens (Group II), however, do not behave like the corresponding unshocked potassium feldspar with an intermediate degree of Al/Si order, probably because the infrared absorption spectrum is affected by both the crystalline and X-ray amorphous material whereas the c^*/b^* ratio is obtained only for the crystalline phase. The positions of the characteristic absorption bands of the moderately shocked samples [single grains, Nos. 105, 131(2) and 250(2)] lie close to those of high sanidine, although the crystalline phases have intermediate c^*/b^* ratios. Thus, the degree of Al/Si order is lower in the X-ray amorphous material (diaplectic glass) than in the crystalline material.

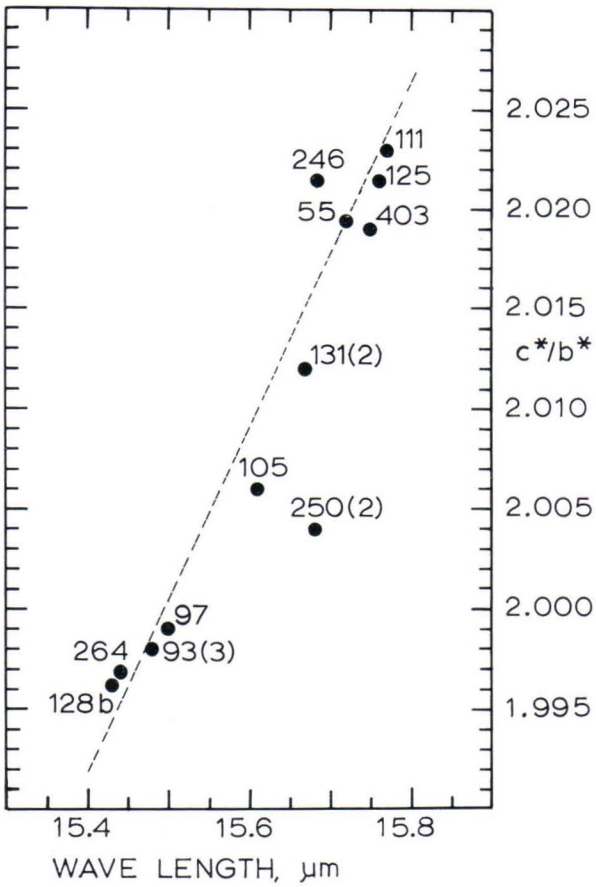

Fig. 20. c^*/b^* plotted against the wave length of one of the characteristic absorption bands (the A-band of Laves and Hafner, 1962, p. 61) of 12 alkali feldspar specimens. The dashed line is the best fit from Lehtinen (1974, Fig. 2).

Discussion

The optical, X-ray and infrared studies of weakly and moderately shocked Lappajärvi microcline (Group II) establish the formation of certain deformation structures and anomalous optical properties accompanied by Al/Si disordering and the formation of diaplectic potassium feldspar glass. These features are indicative of peak pressures of about 115 kb to 300 kb (mixed phase region, Sclar and Usselman, 1970; Kleeman, 1971, p. 5501; Ahrens and Liu, 1973, Fig. 1). Strongly and intensely shocked microcline (Group III) exhibits small-

scale flow at least, and its structural state corresponds to that of the high sanidine — high albite series. The adjacent minerals show signs of high temperature, and the quartz has recrystallized. In most cases, the feldspar material (microcline + albite-oligoclase) has been melted, homogenized and recrystallized.

The nature of shock-induced Al-Si disordering in the weakly and moderately shocked microcline (Group II) is somewhat puzzling. It cannot be of thermal origin alone, as in the Group III specimens, because the adjacent minerals show no signs of having been affected by high temperature. Furthermore, rapid heating at 1050°C does not change the angles of maximum microcline (Grundy and Brown, 1967, Table 2). When shocked to the mixed phase region, microcline partially transforms into a high-pressure phase, possibly with hollandite structure (Ahrens and Liu, 1973, p. 1277), and upon pressure release this phase reverts to diaplectic microcline glass. The Group II specimens consist of diaplectic glass and crystalline material in which the degree of Al/Si order is highly variable. In some specimens, the crystalline material is a mixture of triclinic and monoclinic form of potassium feldspar. Thus, it is possible that the high-pressure phase can revert to a mixture of diaplectic glass and potassium feldspar with a lower, variable degree of Al/Si order, as a consequence of highly variable post-shock thermal distribution (temperature gradient) in the rock (see Kieffer, 1971, p. 5470).

In many Lappajärvi specimens of medium-grained granite the diaplectic plagioclase glass (maskelynite) is common, but the corresponding microcline glass is rare and occurs only as small patches. This suggests that the transformation within the mixed-phase region is more sluggish in microcline than in plagioclase (oligoclase), thereby indicating that a given shock pressure results in partial isotropization and Al/Si disordering of microcline whereas the adjacent plagioclase is almost totally transformed to maskelynite.

Plagioclase

In unshocked Lappajärvi rocks, plagioclase ranges from albite to oligoclase. In granite and granite pegmatite specimens, it ranges from An_6 to An_{14}, and in granodiorite, mica gneiss and quartz diorite specimens, the observed range is An_{18} to An_{28} (weight %). The mean refractive index and the An content (weight %) of some unshocked (off-structure) plagioclase specimens are included in Table 7.

Maskelynite may be distinguished from crystalline plagioclase even in a hand specimen: maskelynite breaks along curved surfaces and is clear and colorless, although grains with birefringent patches may have a bluish or yellowish tint. Consequently, when one set of lamellae of a multiple-twinned plagioclase grain

Table 7

Refractive indices (n) of maskelynite specimens and mean refractive indices [$n_m = (n_x + n_y + n_z)/3$] of unshocked plagioclase specimens. Immersion method, sodium light, accuracy $\pm$ 0.001

Maskelynite		Unshocked plagioclase		
No.	n	No.	n_m	An weight %
251	1.522	16	1.535	8
212	1.522	128b	1 539	14
93	1.523	264	1.540	14
214	1.523	151	1.543	22
250	1.525	150	1.544	23
		128a	1.545	25
		152	1.547	27

has been transformed into maskelynite, the maskelynite lamellae are clear and colorless whereas birefringent lamellae are dull and white.

In thin sections and under the binocular microscope perlitic cracks are visible in the Lappajärvi maskelynite. A clear single pearl of maskelynite was 1 mm in diameter, and its refractive index was 1.523 $\pm$ 0.002 (specimen No. 93). An X-ray powder pattern was taken from the pearl (Cu/Ni radiation, camera diameter 57.3 mm), but even after an exposure time of 14.5 hours, there were still no lines visible. The pearl was heated at (1100 $\pm$ 20°)C for two hours,

Fig. 21. Maskelynite with irregular birefringent patches. Sillimanite needles (left) are full of planar deformation structures. Granite fragment from impact breccia, Specimen No. 212. Crossed polarizers, 130 X.

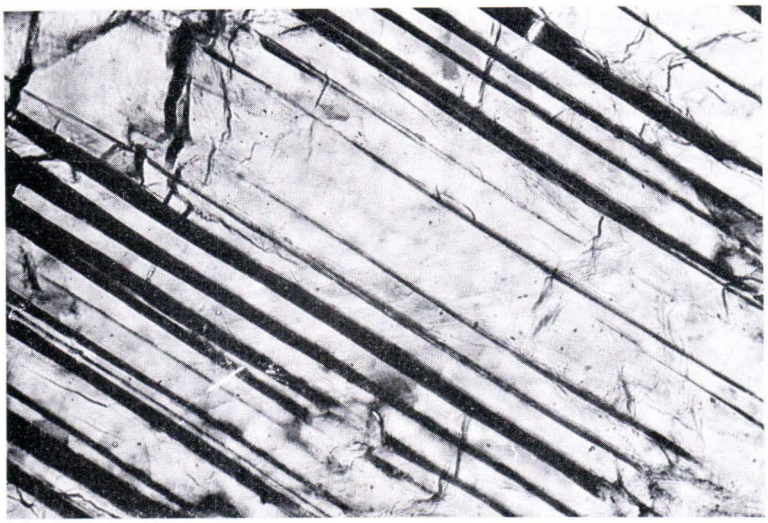

Fig. 22. Asymmetric isotropization (transformation into maskelynite) of oligoclase which is multiply twinned according to the albite law. The birefringent twin lamellae contain thin planar elements. Granodiorite fragment from impact breccia, Specimen No. 98. Crossed polarizers, 130 X.

and a new pattern was recorded (exposure time 15 hours). This heat treatment transformed the pearl into a white and dull polycrystalline mosaic of oligoclase yielding a good powder pattern with sharp peaks.

Although maskelynite is a common constituent of shocked granite, granite pegmatite and granodiorite fragments of the impact breccia and suevite, it is very difficult to find plagioclase grains that have been transformed totally into maskelynite. The grains usually contain small, irregular anisotropic patches which often pass gradually into isotropic plagioclase (Fig. 21). Asymmetric isotropization of plagioclase is visible in many plagioclase grains multiple twinned according to the albite law (Fig. 22). This demonstrates how in a polycrystalline rock, reverberations and the interaction of both shock and rarefaction waves may produce a different shock history for each mineral grain. It also shows that the shock effect may be vectorial, because the spatial orientations of the lattices of the two sets of twin lamellae are different and only one set has been transformed into a glassy state.

Thin, short and non-decorated planar elements are very common in the Lappajärvi plagioclase. They occur as closely spaced parallel sets, usually 1 to 3 sets in each grain or twin lamella. Bent elements and deformation bands are more common in plagioclase than in quartz. Planar elements are abundant in grains that have isotropic patches or are strongly deformed. Thus, it is difficult to obtain reliable crystallographic orientations of the elements, and their

orientation in the Lappajärvi plagioclase was not studied systematically. The reader is referred to the detailed measurements on the Ries (Stöffler, 1967) and Manicouagan (Dworak, 1969) plagioclase, which suggest that the lamellae tend to lie along simple crystallographic directions of the plagioclase lattice.

The refractive index was measured for five grains (pearls) of Lappajärvi maskelynite, and the results are presented in Table 7. Although the chemical composition of the grains is not occurately known, there is no doubt they are oligoclase, and in that case, the results agree well with those of Stöffler and Hornemann (1972, Fig. 5), that is, the refractive indices are intermediate between normal glasses and the mean refractive indices of normal oligoclase of the same chemical composition.

On the basis of X-ray diffractograms and the unit-cell parameters measured an attempt was made to group or classify the shocked Lappajärvi plagioclase (cf. the three groups of alkali feldspars, Table 4). But the classification failed for the following reasons:

1) Moderate and strong shock effects in Lappajärvi plagioclase are almost always accompanied by incipient secondary zeolitization (formation of erionite or heulandite, or both). Some important plagioclase peaks, such as 131, displays strong interference from adjacent or overlapping zeolite peaks.

2) Microcline and oligoclase appear to behave differently under similar shock conditions (see also p. 58). Weakly and moderately shocked microcline samples (Group II) are mixtures of glassy material and potassium feldspar with lower, variable Al/Si order, and it is difficult to find homogeneous diaplectic glass of microcline. The adjacent oligoclase is, on the contrary, usually X-ray amorphous or has very diffuse peaks, and it appears that shocked plagioclase tends to have either a rather intact or an X-ray amorphous crystal lattice.

3) To group plagioclase the chemical composition of each grain should be known because the X-ray properties of plagioclase depend strongly on its chemical composition.

A number of X-ray diffractograms of unshocked and shocked specimens of Lappajärvi oligoclase were recorded and measured. It was found that, for example, the angle between the 131 and $1\bar{3}1$ peaks is $1.42 \pm 0.02°$ for unshocked grains of oligoclase (specimens Nos. 128b, 151 and 264), whereas the angle for three shocked plagioclase grains from specimens Nos. 213 and 251 is $1.56 \pm 0.01°$ (2Θ CuKα). According to Bambauer *et al.* (1967, Fig. 5), the greater angle is indicative of a lower degree of Al/Si order in the shocked oligoclase. This finding is consistent with Dworak's (1969, pp. 335—336) measurements on the Manicouagan labradorite. But it must be kept in mind that the result may be misleading because of the unknown composition of the shocked oligoclase and owing to broadening of the peaks.

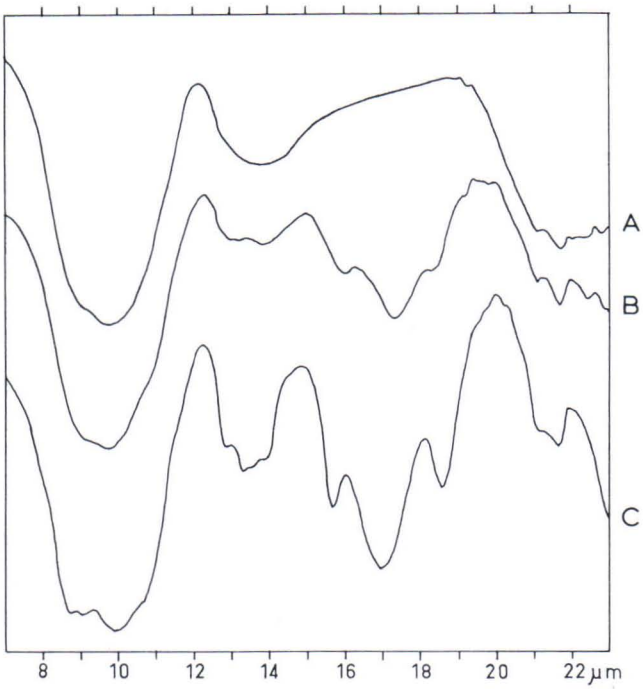

Fig. 23. Infrared absorption spectrum of A) maskelynite (specimen No. 93), B) heated maskelynite (at 1100°C for two hours, specimen No. 93), C) oligoclase An_{21}.

Infrared absorption spectra from 1 μm to 24 μm were recorded for five unheated and heated samples of Lappajärvi maskelynite. Two of the spectra are reproduced in part in Fig. 23. For reference, infrared spectra of an oligoclase specimen, An_{21}, (Kracek and Neuvonen, 1952, oligoclase 103086) and of the Manicouagan maskelynite (Bunch *et al.,* 1967; Dworak, 1969) were also recorded; the reference material was made available by the courtesy of Professor Th. G. Sahama, University of Helsinki. Details of the experimental procedure were given in an earlier publication by the present author (Lehtinen, 1974, p. 225).

Four groups of absorption bands are manifest in the region from 7 μm to 24 μm (about 1 400 cm^{-1} to 420 cm^{-1}) on the infrared spectrum of unshocked oligoclase (Fig. 23, Spectrum C; see also Thompson and Wadsworth, 1957, Fig. 1). The groups lie at 8.5 μm to 11 μm, 12 μm to 15 μm, 15 μm to 19 μm and 20.5 μm to 22 μm, respectively. The spectra of the Lappajärvi and Manicouagan maskelynite are rather similar to each other. Their most characteristic feature is the disappearance of the third band group at 15 μm to 19 μm (Fig. 23, Spectrum A), which is to be expected in shocked lattice because,

according to Iiishi *et al.* (1971, Table 4), the said group consists of absorption bands resulting from the relatively complicated O-Si(Al)-O and O-Si-O bending vibrations. The fourth band group contains absorption bands caused by Na-O/Ca-O stretching and by Si-O-Si deformation vibrations (Iiishi *et al., ibid.);* it shows no significant loss of details. The first and second band group are caused by Si-O, Si(Al)-O and Si-Si, Si-Al(Si) stretching vibrations, respectively, and in the maskelynite spectrum both groups have lost details. The second group is moreover much weaker in the maskelynite spectrum than in the oligoclase spectrum. As in the spectra on Manicouagan maskelynite and diaplectic labradorite (Dworak, 1969, Fig. 15), the presence of the third band group was observed in spectra of plagioclase grains that were only partially transformed into maskelynite.

Five maskelynite specimens were heated for two hours at temperatures ranging from 930° C to 1 100° C, and their infrared absorption spectra and X-ray powder patterns were then recorded. The powder patterns showed that the maskelynite grains were converted to crystalline oligoclase. Their devitrification is also visible in their infrared absorption spectra (Fig. 23, Spectrum B), but especially the third band group is weaker than that of normal oligoclase. The infrared absorption spectra and the tentative annealing experiments on the Lappajärvi maskelynite indicate that its physical properties correspond closely to those of maskelynite from the Clearwater West and Manicouagan craters in Canada (Bunch *et al.,* 1967, 1968; Dworak, 1969).

Biotite and muscovite

Kink bands in biotite (micas) are almost omnipresent in rocks associated with known or suspected astroblemes, e.g. the Ries Crater (Stöffler, 1966, Fig. 3), the American and Canadian craters (Bunch, 1968, pp. 429—431; Dence *et al.,* 1968, pp. 345, 348, 355; Short and Bunch, 1968, Table 1), and with nuclear explosion craters (Cummings, 1965, 1968; Short, 1966, p. 1204, Fig. 4 c, 1968, pp. 202—205).

Kinking is a combination of translation gliding and external rotation around an axis lying in the glide plain perpendicular to the glide direction (Turner *et. al.,* 1954, p. 896; Hörz and Ahrens, 1969, p. 1213). Kink bands occur in biotite as sharply defined, usually lenticular features. A large number of kink bands are often observed within a single grain.

Cummings (1965, 1968) studied kinked biotite in granodiorite from the vicinity of the 5-kiloton Hardhat Event, Nevada Test Site. He found three distinct preferred orientations of kink bands with respect to the shock-wave propagation direction. More than half of the kink bands were oriented with the long axis of kink band perpendicular to the impact direction. The other two preferred orientations were 70° ± and 50° ±.

Investigating the Hardhat granodiorite samples, Short (1968, p. 202) found that at pressures between 30 kb and 60 kb only a few mica grains were kinked. In a sample shocked at 120 kb, nearly all biotite grains contained kink bands. Shock-attenuation calculations by Cummings (1968, Fig. 6) and Hörtz (1969, p. 373) suggest that the minimum shock pressure for the kinking of biotite is about 10 kb. Hörz and Ahrens (1969) performed a series of shock experiments on single crystal biotite and demonstrated that kinking occurs at pressures as low as 9 kb. They also found that virtually no kink bands form in specimens in which the shock wave propagated perpendicular to the {001} cleavage. This is in agreement with the observations of Cummings (1968, p. 214), who stated that kink bands only form in biotite grains whose *c*-axis is steeply inclined (>45°) to the shock wave propagation.

The kink bands in biotite, however, are not exclusively caused by shock. The occurrence of kinked biotite in tectonites has long been known (Turner, 1964, p. 347 and references included). Kink bands have also been produced artifically in biotite from a granite by static deformation at 500° C and 5 kb confining pressure (Griggs *et al.,* 1960, p. 65). The kink bands tended to form preferentially in grains oriented with the *c*-axis steeply inclined to the compression axis.

Cummings (1968, pp. 214—215) and Hörz and Ahrens (1969, pp. 1224—1226) reported anomalous interference figures in shocked biotite specimens. Hörz and Ahrens (1969) also noted a marked decrease of the $2V$ of biotite with increasing shock pressure, and asterated reflections in Laue photographs They interpreted this as a result of the irregular stacking order of mica sheets which were partly tilted and probably rotated under shock conditions.

Hörz (1970, p. 975) demonstrated that there is no clear crystallographic control of the kink angles or of the angle of external rotation in the kink bands formed. In addition, his data show that shock-induced kink bands are more asymmetric than their statically generated counterparts.

In the Lappajärvi area, kinked biotite is a common constituent of fragmental breccias, especially of their biotite-plagioclase gneiss, granite and granodiorite fragments. Single crystals of biotite with or without kink bands occur in the fine-grained impact breccia and rarely also in suevite but never in kärnäite.

Many factors seem to affect the development of kink bands. They prefer to form in flakes oriented with their cleavage plane parallel to the direction of shock-wave propagation (Cummings, 1968, p. 214; Hörz and Ahrens, 1969, p. 1216) but, because of reverberations and the complexity of the shock wave, orientation does not play an important role in large meteorite craters (see Cummings, 1966, p. 952). The grain size, texture and mineral composition of rocks must also have a marked influence on the formation of kink bands. For example, specimen No. 213 is a migmatite composed of medium- or fine-grained biotite-

plagioclase gneiss and of coarse-grained granite. Both components contain diaplectic quartz and microcline, and the plagioclase has partially transformed into maskelynite (Shock Stage II). In the gneiss component, which contains about 40 volume % biotite, only few biotite flakes are notably kinked, whereas in the granite component almost all flakes are kinked, and many of them intensely.

The decomposition and oxidation of biotite and the vesiculation of feldspars are closely related in the Lappajärvi rocks. When some feldspar grains in a rock show incipient vesiculation or fluidal texture ("softening"), biotite displays reduced birefringence and pleochroism and "dirty" colors. With increasing vesiculation of feldspars the decompositon of biotite into a brownish or greenish substance with numerous black or brown crystallites (magnetite and spinel) also increases rapidly. In places, the kink bands or their remnants remain visible even until the final stage of decomposition. Totally decomposed biotite flakes either have lost their original shapes or their grain boundaries are indistinct. At this stage of shock metamorphism (Stage III), quartz is generally subconchoidally fractured and is brownish in thin section. In still more strongly shocked rocks (Stage IV), in which the feldspars have melted and recrystallized as spherules of sanidine or anorthoclase, or both, decomposed biotite has flowed in the feldspar melt and is now present as irregular dark streaks within the feldspar. However, there are mica-bearing rock fragments in the kärnäite, and rarely in the suevite, that contain completely decomposed biotite occurring together with diaplectic quartz and feldspar. This indicates that (1) the cooling rate was more rapid in suevite than in kärnäite and (2) that the residual temperature was generally higher in kärnäite. According to Chao (1968, Fig. 1) and Stöffler (1971a, Table 2), the decomposition of biotite begins when the peak pressure exceeds about 400 kb to 450 kb raising the residual temperature to about 650° C—900° C.

Well-developed and symmetrical kink bands due to tectonic deformation have been observed in a strongly folded biotite-rich gneiss near the southern end of Lake Lappajärvi, and rare kinked muscovite flakes are present in the muscovite granite in the northern part of the area. It is thus possible that some of the kink bands manifest in the shocked rocks are of pre-shock origin. Such kink bands are, however, rare, and in strongly shocked rocks in which the biotite is intensely kinked they are unimportant.

In the Lappajärvi rocks, the plane of kinking (kink band boundary) is steeply inclined to the cleavage plane of mica. But because the kink bands in shocked mica are more or less lenticular and asymmetrical and are obscured by secondary alteration, the measurement of accurate kink angles is difficult. About 100 orientations of kink bands were, however, measured on biotite flakes whose cleavage planes were nearly perpendicular to the plane of the thin section.

Most of the kink bands were oriented with the long axis of the kink band lens inclined 60° to 85° to the {001} cleavage plane; the angles measured ranged from 47° to 89°.

The kinked biotite grains are usually characterized by only one set of almost parallel kink bands. In strongly shocked specimens, the biotite is more intensely kinked, and two or three different systems of kink-band orientations are present. They are often visible, even with the naked eye, on the cleavage plane of biotite. In thin sections, parallel sets of very thin planar elements occur in some biotite flakes with horizontal or nearly horizontal cleavage planes. The planar elements are parallel to the $(11l)$ and $(1\bar{1}l)$ planes in biotite, most probably to (111) and $(1\bar{1}1)$.

In many mica gneiss specimens of Shock Stage III rather thin (width <7 μm) planar deformation structures are seen in the biotite when it is viewed parallel to {001}. They usually occur as two parallel, closely spaced sets inclined symmetrically 55° to 75° to the cleavage to the cleavage plane (a herringbone pattern). The deformation structures have been explained as planar fractures formed parallel to the (111), $(1\bar{1}1)$, $(11\bar{1})$ or $(1\bar{1}\bar{1})$ plane, or planes, of biotite (see Stöffler, 1972, Fig. 23 b). But it is also possible that some of them are thin, parallel kink bands. It is known that as the peak pressure increases the width of the kink bands in biotite decreases (Stöffler, 1972, p. 85).

Muscovite is as an accessory mineral in some shocked granite specimens (Shock Stages I and II) which have evidently been embedded in the impact breccia. In addition, a few single flakes of muscovite occur in the medium-grained impact breccia. The behavior of muscovite under shock conditions seems to be very similar to that of biotite, although decomposed muscovite has not been detected in the Lappajärvi rocks.

Spinel

Spinel was detected in the X-ray diffractograms of the residues of some rock specimens chemically treated for the coesite test. Two powder patterns and the unit-cell parameters calculated are listed in Table 8. For comparison, a part of the powder pattern of synthetic $MgAl_2O_4$ (Swanson and Fuyat, 1953, p. 37; ASTM Card 5-0672) is included in the table. Unit-cell parameters were also calculated for two additional Lappajärvi spinel specimens, the results are $a_0 = (8.062 \pm 0.006)$ Å (No. 38) and $a_0 = (8.130 \pm 0.004)$ Å (No. 241).

Spinel is insoluble in the acids used (see p. 40) and thus even minute amounts of it may be detected. Specimen No. 236, the piece of suevite described on p. 41, and the biotite-plagioclase gneiss specimens Nos. 241 and 243 have all undergone strong or intense shock metamorphism (Shock Stages III—IV). The feldspars are almost completely fused, the quartz with planar elements is

Table 8

X-ray powder patterns of spinel specimens, Cu/Ni radiation, internal Si standard

	ASTM 5-0672 (synthetic)		No. 243, Lappajärvi		No. 236, Lappajärvi	
hkl	d(Å)	I	d(Å)	I	d(Å)	I
111	4.67	4	4.67	4	4.65	3
220	2.858	40	2.859	51	2.849	32
311	2.436	100	2.438	100	2.432	100
222	2.333	3	2.334	4	—	—
400	2.021	58	2.021	38	2.015	40
422	1.649	10	1.649	17	1.645	12
511	1.555	45	1.556	45	1.550	47
440	1.429	58	1.429	56	1.424	59
531	1.366	3	—	—	—	—
620	1.278	2	—	—	—	—
533	1.232	9	1.231	5	1.229	8
	$a_0 = 8.0800$ Å		$a_0 = 8.080$ Å ± 0.005 Å		$a_0 = 8.058$ Å ± 0.005 Å	

subconchoidally fractured, and the biotite has been entirely decomposed and recrystallized into a microcrystalline or cryptocrystalline mass. Specimen No. 38 represents Type I kärnäite in which, at 400 X magnification, spinel was detected in thin section as perfect octahedrons (diameter <0.01 mm) that vary from almost colorless to violet or reddish brown. The spinel crystals or crystal groups are embedded in recrystallized feldspar (sanidine spherules) on the margin of a granite fragment. No spinel crystals were detected in the inner parts of the fragment, but biotite is decomposed into almost opaque substance.

In specimens Nos. 38, 236 and 243, spinel has a remarkably small unit-cell parameter. For example, according to Strunz (1970, pp. 176—177), a_0 is 8.102 Å for pure $MgAl_2O_4$ and 8.135 Å for pure $FeAl_2O_4$ (hercynite). Gahnite ($ZnAl_2O_4$) has the smallest unit-cell parameter (a_0 = 8.078 Å) of the spinels. For chemical reasons, the Lappajärvi spinel cannot be gahnite. In all other natural spinel minerals a_0 is greater than in hercynite.

Binns *et al.* (1969) found ringwoodite, $(Mg,Fe)_2SiO_4$ spinel, in the Tenham meteorite, an olivine-hypersthene chondrite that shows effects of strong shock metamorphism. At high pressures the olivine structure is transformed into the spinel structure (ringwoodite) and the pyroxenes break down to ringwoodite plus stishovite (Ringwood and Major, 1966). The unit-cell parameter, a_0, of the Mg-ringwoodite is (8.07 ± 0.02)Å (Ringwood, 1958, pp. 20, 28; Akimoto and Ida, 1966) and of Fe-ringwoodite, 8.235 Å (Ringwood, 1958, p. 24).

It is not possible to decide on the basis of the data available whether or not the spinel in the Lappajärvi rocks may be interpreted as a high-pressure shock-metamorphic decomposition product of biotite. The extremely small grain size

of the spinel prevents a close optical study, and the spinel-bearing residues contain also other minerals and are not suitable for chemical analysis. It seems reasonable, however, to assume that the spinel is a product of shock metamorphism, and that the small unit-cell parameter is due to high temperature and rapid cooling.

There are no descriptions to date of a spinel similar to that in the Lappajärvi rocks. Chao (1967a, p. 230), however, found a fine-grained spinel in the schist fragments in the suevite of the Amerdigen quarry in the Ries Crater, and he has proposed that it formed after the passage of the shock wave, i.e. during cooling. More recently Stähle (1975) has described the formation of small spinel crystals (hercynite or pleonaste) as a high-temperature breakdown product of almandine from the Ries Crater.

Sillimanite

Naturally shocked sillimanite has so far been reported only from the Ries Crater, Germany (Stöffler, 1970), where it occurs as a main constituent of some garnet-sillimanite schist fragments in suevite. The physical properties of shocked sillimanite differ considerably from those of unshocked sillimanite, and it has been partially transformed into mullite plus cristobalite. The most striking feature, however, visible under the microscope, is the formation of sets of planar deformation structures (diaplectic sillimanite). Stöffler (1970, p. 115) estimated that the sillimanite was shocked to a peak pressure exceeding 300 kb to 400 kb.

In the Lappajärvi rocks, sillimanite crystals are found in two specimens only. Specimen No. 212 is medium-grained shocked granite in which the plagioclase has been almost totally transformed into clear maskelynite with perlitic cracks and the quartz has many sets of planar elements (shock lamellae); the quartz also contains some worm-like stringlets of diaplectic glass. Microcline is strongly deformed and has lowered birefringence. Biotite is fresh but kinked. Sillimanite is present as long, subhedral crystals embedded in plagioclase (maskelynite) as it is also in Specimen No. 136, migmatite composed of mica gneiss and granite. Specimen No. 136 also contains an appreciable number of garnet and apatite crystals with abundant planar fractures. Feldspars have been partially vesiculated and have a fluidal texture. Quartz is recrystallized almost totally as a polycrystalline mosaic, and, especially in the gneiss, it is very "dirty", i.e. with a large amount of brown pigment derived from the adjacent decomposed biotite. Furthermore this migmatite was affected by secondary alteration resulting in the sericitization of feldspar and sillimanite. The solid-state transformation products (diaplectic glass of quartz and feldspar) in specimen No. 212

indicate that it was shocked at a peak pressure of about 350 kb. Specimen No. 136, showing incipient vesiculation of feldspar, was evidently subjected to a somewhat higher post-shock temperature.

Planar fractures and planar elements (shock lamellae) are omnipresent in the sillimanite in specimens Nos. 212 and 136, although it is often difficult to distinguish between planar fractures and planar elements. In long sillimanite needles three or four sets of multiple fractures seems to be the most common combination of planar structures. Two of the sets lie parallel to (010) and (001), and the other two sets are inclined 30°—40° to the *c*-axis. Fifteen sillimanite crystals were studied by use of the universal stage. When all the measured orientations of the planar deformation structures were plotted within a common (*hkl*)-quadrant of the stereographic projection (see Fig. 2 of Stöffler, 1970) eight types (or groups) of planar structures in sillimanite were distinguished: (1) {010}; (2) {001}; (3) {*hk*0}; (4) {021} or {031}, or both; (5) {201} or {301}, or both; (6) {313} or {101}, or both; (7) {133} or {122}, or both, and {011}; (8) {112} or {113}, or both. The common {010} cleavage of sillimanite suggests that the {010} type may be of pre-shock origin. Planar fractures of type {*hk*0} are not necessarily parallel to any common prism face (see also Stöffler, 1970, Fig. 2). The orientations of the planar structures of the Lappajärvi sillimanite closely resemble those of the Ries sillimanite, although the Lappajärvi mineral has some additional orientations, viz. types (4) and (5). [The ρ-angles for (131), (311), (133) and (313) of sillimanite are 67.4°, 67.7°, 38.6° and 39.1°, respectively. Stöffler's (1970) Fig. 2 shows that the orientations {311} and {131} reported by him, must actually be {313} and {133}.]

The content of sillimanite is too low to permit either infrared absorption or X-ray studies. Its optical properties are, unlike those of the Ries sillimanite, almost unchanged by shock metamorphism. For example, the direct measurement of $2V$ was possible in seven crystals (specimen No. 212) and the result, $2Vz = 29°—31°$, does not differ from $2V$ of unshocked sillimanite. Evidently the Lappajärvi sillimanite has not been so strongly shocked as the Ries sillimanite, although at both localities the shock effects in the adjacent feldspar and quartz indicate approximately the same stage of shock metamorphism.

Apatite

A few apatite crystals with planar deformation structures have been observed in specimens Nos. 84 and 136, consisting of medium-grained mica gneiss migmatized by medium- or coarse-grained granite. Both specimens are pieces of suevite boulders, and specimen No. 136 is partly covered by glass with fluidal texture. The apatite crystals are relatively large, from 0.4 mm to 1.4 mm in

diameter, subhedral or anhedral, and embedded in feldspar that shows marks of incipient flow and vesiculation. The small apatite crystals, usually 0.1 mm to 0.2 mm in diameter and which are very common in many specimens, do not display clear shock-induced deformation structures. The number of different sets of planar deformation structures varies commonly from four to six within a sizable single grain. Irregular fractures may also occur. The planar fractures are long and penetrate a crystal whereas the planar elements are short and very thin and form multiple, closely spaced sets. Both decorated (with dark, very small inclusions) and undecorated planar elements occur but the undecorated ones are more common. Some strongly fractured grains have begun to recrystallize, and form a mosaic of tiny crystals.

Six apatite crystals, about 0.7 mm to 1.2 mm in diameter, were studied on the universal stage. Most of the orientations measured for the planar deformation structures coincide with those of rational crystallographic planes. Planar fractures are often parallel to (0001). In addition fractures parallel to $(10\bar{1}0)$ and $(21\bar{3}0)$ or the symmetrically equivalent faces were found. The planar elements (shock lamellae) lie parallel, or closely parallel, to the faces of forms $\{10\bar{1}1\}$, $\{10\bar{1}2\}$, $\{11\bar{2}1\}$, $\{11\bar{2}2\}$ and $\{31\bar{4}2\}$. The forms are listed in the order of decreasing frequency, but only 25 sets were measured. The result is, nevertheless, in close agreement with the data for shocked apatite reported by Stöffler (1972, Table 4). Shock effects in minerals associated with the apatite at Lappajärvi indicate peak pressures exceeding 300 kb and high post-shock temperatures.

Garnet

Garnet, a rare accessory mineral in the Lappajärvi rocks, was observed in three shocked specimens. Two of them (Nos. 84 and 136) are migmatite fragments from the suevite with shock effects typical of Shock Stages II and III. Specimen No. 136 was described on p. 68. The garnet grains in the migmatite specimens are anhedral or subhedral, with a diameter of less than 4 mm. All the garnet grains studied contain numerous, partially curved cracks and thin planar deformation structures which are evidently crystallographically controlled. In places, the planar structures are abundant enough to give the mineral an almost opaque look. Nevertheless, there are patches in the garnet that contain no cracks or lamellae.

The third garnet-bearing specimen (No. 255) is strongly shocked medium-grained aplitic granite. Small kinked flakes of muscovite are rather common, and the garnet is violet brown, euhedral and up to 6 mm in diameter. It yields an X-ray diffractogram with sharp peaks ($a_0 = 11.538 \pm 0.005$ A), although the associated quartz and feldspar have been largely transformed into diaplectic

glass. In the outer parts of the granite specimen, the feldspar has been sanidinized, and the diaplectic quartz glass shows incipient recrystallization (Shock Stage III).

The coexistence of fractured or diaplectic but crystalline garnet and diaplectic (thetomorphic) quartz and feldspar glass affords, as do the diaplectic sillimanite crystals embedded in maskelynite, conclusive proof of great differences in resistance to shock deformation in minerals with different structures and packing densities.

Zircon, sulfide minerals and graphite

Zircon is a common accessory mineral in the Lappajärvi rocks, but its crystals are usually too small for a detailed optical study. Some zircon crystals are large enough (0.2 mm to 0.4 mm in diameter; Specimens Nos. 79 and 134) to allow reliable universal-stage measurements. The crystals have different sets of multiple planar structures, i.e. planar fractures and very thin lamellae similar to the shock lamellae (planar elements) of quartz. The number of sets of the planar structures observed varies from four to seven in a single grain. They are predominantly oriented parallel to or close to {100}, {110}, {210}, {102}, {201}, {001}, {112} and possibly also {212}. The uneven spatial distribution of the planar structures within a single crystal is typical of the planar structures. Strong planar fractures penetrate the crystal, but a certain set of thin lamellae may occur only in a given part or a corner of the crystal. The zircon with planar structures are found in shocked granite specimens and are embedded in diaplectic quartz, plagioclase and microcline. The two rock specimens also contain some maskelynite indicating peak pressures exceeding 350 kb (Shock Stage II).

The strongest peak of zircon (d = 3.302 Å) was observed in some X-ray diffractograms of the acid-treatment residues of suevite (see p. 42). El Goresy (1965, p. 3454) found that the decomposition of zircon into baddeleyite plus silica is a common feature of the zircon crystals in the suevite of the Ries Crater. The decomposition of zircon into baddeleyite plus silica is indisputable proof of extremely high temperatures (exceeding about 1775°C, El Goresy, 1965, p. 3453). It might, consequently, be possible to detect baddeleyite also in the Lappajärvi suevite, but not without a detailed study of a number of polished sections.

Minor accessory minerals include sulfides (pyrite, pyrrhotite, chalcopyrite), magnetite and graphite. The pyrite occurs as veins filling fractures in feldspar and quartz grains embedded in kärnäite and suevite. The groundmass of some kärnäite specimens also contains numerous very small grains of pyrite. In kärnäite and suevite, the pyrite is often accompanied by pyrrhotite and sometimes

by chalcopyrite, but seldom by magnetite. The pyrrhotite appears to surround sizable pyrite grains in kärnäite. In a small vesicle in Type I kärnäite a hexagonal single crystal of pyrrhotite was detected, which is bronze-colored, tabular on the basal pinacoid and has a rounded, pyramidal habit. Precession photographs (Mo/Zr radiation) revealed a 2A 4C superstructure (identical with that observed by Fleet and Macrae, 1968, Fig. 4).

Graphite is a common accessory mineral in shocked migmatite in which it often occurs in the contact against mica gneiss and granite. In polished sections, the graphite flakes exhibit well-developed kind bands with very sharp band boundaries. As in biotite, the kink-band boundaries tend to be steeply inclined to the basal plane. The presence of a hexagonal form of carbon, chaoite, which occurs together with strongly shocked graphite in the Ries Crater (El Goresy and Donnay, 1968, 1969), was also studied in polished sections but with negative results.

An error in the preliminary report on Lake Lappajärvi (Lehtinen, 1970, p. 92) is corrected here. The crystalline phase which occurs in the residual material after treatment with hydrofluoric acid and nitric acid (this paper, p. 41) was reported erroneously to be very fine-grained pentlandite with a remarkably small unit cell. It is, however, ralstonite formed during the acid treatment, and it is the presence of this ralstonite that explains the small unit cell observed.

Secondary minerals

Almost all Lappajärvi rocks of shock-metamorphic origin are more or less porous or amygdaloidal. Zeolites are the most common secondary minerals in the porous suevite, whereas calcite is rather common in the kärnäite.

Calcite usually occurs as white spherulite-like aggregates. The *d* spacing of the 112 reflection (Goldsmith *et al.*, 1955, Fig. 1) indicates that the samples studied are almost pure $CaCO_3$ with the exception of a yellowish sample that contains about 5 mol % $MgCO_3$.

Heulandite is the most common zeolite in the specimens investigated. It generally occurs as small, clear and well-developed orthorhombic-looking crystals in rock cavities, but there are some specimens (mica gneiss fragments in the suevite) in which heulandite is an alteration product of plagioclase. Heulandite was discovered in X-ray diffractograms together with chabazite or erionite, or both. A few single crystals of heulandite were identified by means of precession photographs (Mo/Zr radiation). The systematic extinctions agree with the space group C2/m. Although heulandite was not distinguished from clinoptilolite in the X-ray study, unit-cell parameters were calculated for two heulandite specimens, viz. Nos. 316 and 327 (Cu/Ni radiation, internal quartz standard).

According to Boles (Figs. 5 and 6b, 1972), the findings ($a_0c_0\sin\beta$ = 118.1 Å^2 and 117.9 Å^2, b_0 = 17.93 Å and 17.91$_5$ Å, respectively) indicate a large number of aluminum ions and a low number of magnesium ions per unit cell. This suggests that the two specimens are heulandite (Mg-poor).

Stilbite occurs as radiate crystal groups in cavities in the suevite. It is white or colorless. Chabazite is often present on stilbite spherulites as colorless rhombohedrons (penetration twins) a few millimeters in diameter. Chabazite rhombohedrons occur, for example, in specimen No. 314, and the X-ray powder pattern recorded is very similar to that given on the ASTM Card 15-618.

Mordenite is cottony or hairy and white or reddish. It is fairly common within vesicular feldspar (sanidine) fragments in suevite. Erionite, which was identified in four breccia specimens, occurs as almost colorless fiber aggregates in small cavities and fissures. Erionite and offretite are closely related, but they can be distinguished by some relatively strong lines in their X-ray powder patterns (Sheppard and Gude, 1969, pp. 881—883; Kerr *et al.*, 1970, Fig. 1 and Table 2). Reflections 201 and 211 are medium strong in the diffractograms of Lappajärvi erionite but 101 is weak. In addition, *c*-axis rotation photographs were taken to confirm the length of the *c*-axis, found to be about 15.0 Å to 15.1 Å. An X-ray diffractogram (Cu/Ni radiation, internal quartz standard) yielded unit-cell parameters for erionite (Specimen No. 228; impact breccia) as follows: a_0 = 13.261 ± 0.005 Å, c_0 = 15.068 ± 0.008 Å. The values are consistent with the parameters reported for erionite by Sheppard and Gude (Table 3, 1969).

Chemistry

Rocks

A single analysis of a Lappajärvi rock was published by Eskola (Analysis III, 1921). The collections of the Department of Geology and Mineralogy of the University of Helsinki contain several kärnäite specimens collected by Eskola, and after studying these, the present author came to the conclusion that the specimen selected for chemical analysis by Eskola was most probably Type I kärnäite.

The new chemical analyses of the Lappajärvi rocks were made on specimens representing different types of shocked rocks. The analyses are given in Tables 9a and 9b.

Specimen No. 38 is a homogeneous and compact Type I kärnäite with only a few small (diameter < 2 mm) quartz and feldspar fragments. Microscopically, specimen No. 14 is very similar to specimen No. 38, but megascopically it is

Table 9

Chemical composition of Lappajärvi rocks

a. Weight %

	14 [1]	38 [1]	69 [1]	117 [1]	»Dacite» [2]
SiO_2	62.20	65.86	67.88	62.42	67.20
TiO_2	0.65	0.64	0.48	0.75	0.63
Al_2O_3 ...	13.63	15.46	14.81	14.39	14.65
Fe_2O_3 ...	3.89	0.97	3.11	4.80	—
FeO	3.33	4.41	1.65	0.26	3.24
MnO	0.06	0.05	0.02	0.05	—
MgO	2.74	2.24	1.27	1.42	1.03
CaO	2.34	2.72	2.20	2.28	2.78
Na_2O ...	2.22	2.45	2.80	0.27	3.28
K_2O	2.51	3.37	3.05	2.58	4.28
H_2O+ ..	3.10	0.71	0.69	5.37	1.10
H_2O- ..	2.11	0.70	2.22	5.22	
CO_2	—	—	—	—	0.10
Total	99.71	100.07	100.26	100.19	99.51 [3]

[1]) Analyst, Tapio Koljonen
[2]) Analyst, Eero Mäkinen (Eskola, 1921, p. 12)
[3]) Includes 1.09 % Fe_6S_7

b. Weight percentages recalculated to 100

	14	38	69	117	»Dacite» [1]
SiO_2	65.82	66.75	69.73	69.67	68.29
TiO_2	0.69	0.65	0.49	0.84	0.64
Al_2O_3 ...	14.42	15.67	15.21	16.06	14.89
Fe_2O_3 ...	4.12	0.98	3.19	5.36	—
FeO	3.52	4.47	1.69	0.29	3.29
MnO	0.06	0.05	0.02	0.06	—
MgO	2.90	2.27	1.30	1.58	1.05
CaO	2.48	2.76	2.26	2.54	2.82
Na_2O ...	2.35	2.48	2.88	0.30	3.33
K_2O	2.66	3.42	3.13	2.88	4.35
P_2O_5	0.98	0.50	0.08	0.42	0.13
CO_2	—	—	—	—	0.10

[1]) Contains 1.11 Fe_6S_7

exceedingly rich in cavities. Specimen No. 69 is a well-preserved Type II kärnäite. Specimen No. 117 is a typical suevite with numerous aerodynamically shaped, grayish glassy "bombs" of fluidal texture embedded in a brownish fine-grained breccia matrix. A thin section of the rock, however, reveals vesicles filled with zeolites and products of secondary alteration (clayey substance, probably montmorillonite).

Depending on the amount of zeolites and secondary alteration, the water content varies greatly in the specimens analyzed; hence all the analyses were

recalculated free from water (Table 9b) for their mutual comparison. Owing to the limited number of analyses, a statistical approach to the problem as to whether or not the rocks form by melting and partial melting, brecciating and mixing of a given target rock is not possible. In any case, the chemical composition of the rocks is rather similar, with the exception of specimen No. 117 which has a low Na_2O content but a high Fe_2O_3 content, evidently caused by zeolitization and secondary alteration. By mixing granite plus mica gneiss (or biotite-plagioclase gneiss) with granodiorite, which are the target rocks of the Lappajärvi Crater, a rock is obtained with a chemical composition similar to that of kärnäite. Thus, there is no reason to invoke an eruption of lava to explain the chemical composition of kärnäite or of the other Lappajärvi rocks.

The same conclusion holds good, for instance, for some other possible astroblemes of Fennoscandia, such as Lake Mien and Dellen in Sweden and Lake Jänisjärvi in the U.S.S.R. Jänisjärvi is an illuminating example because the basement in that area "consists of sedimentogeneous crystalline schists whose average composition must be almost exactly the same as that of the lava-rock" (Eskola, 1921, p. 9). Eskola, who regarded this crater lake as a volcanic neck, explained the peculiar chemical composition including the very high content of Al_2O_3 [18.11 weight %—19.08 weight % Al_2O_3 (Eskola, 1921, p. 6)] as the result of assimilation on a large scale. If, however, the rock is considered an impact lava, no difficulties arise in explaining its chemical composition.

The lava-like rocks associated with shock-metamorphic rocks generally show a markedly similar chemical composition to the adjacent country rocks. Dence (1971) and Currie (1971), who studied about fifteen craters, observed, however, that some lava-like rocks were depleted in silicon and sodium and enriched in potassium, magnesium and heavy metals when compared with the target rocks. Selective melting (see v. Engelhardt, 1976b, p. 1686) and differences between the estimated average chemical composition of the target rock and the chemical composition of the rock actually melted usually suffice to explain the difference observed (Dence, 1971, p. 5560). This does not mean that all the lava-like rocks of the supposed meteorite craters are actually impact melts but that the close agreement in chemical composition is worth remembering when the origin of the craters is discussed.

As to kärnäite, it contains numerous small quartz fragments, often subconchoidally fractured and surrounded by an apparent chill zone of pyroxene microlites, whereas the feldspar fragments are strongly corroded and usually without a chill zone. The quartz fragments, constituting 10 % to 25 % (volume %) of the rock, are remains of unfused mineral components, such as some feldspar, but the feldspar fragments have, unlike quartz, reacted with the melt during cooling. A similar phenomenon, which seems to be typical of shock-induced incomplete melting, has been observed in impact-produced melts, such as the

Table 10

Normative mineral composition (Rittmann norm)

	14	38	69	»Dacite»
Quartz	27.8	25.1	29.9	24.2
Sanidine	28.2 Or_{59}	31.4 Or_{60}	29.8 Or_{56}	41.4 Or_{60}
Plagioclase	16.4 An_{29}	22.4 An_{39}	26.0 An_{33}	27.3 An_{36}
Orthopyroxene	7.7	6.3	3.5	4.4
Cordierite	15.8	12.1	9.4	—
Apatite	1.9	1.0	0.1	0.3
Ilmenite	0.8	0.7	0.5	0.7
Magnetite	1.3	1.0	0.8	0.5
Calcite	—	—	—	0.2
Pyrrhotite	—	—	—	1.1

glass bombs from the Ries Crater (v. Engelhardt, 1967b) and the lava-like rock of the Tenoumer Crater (French *et al.,* 1970). In addition, this phenomenon is common in the lava-like rocks of Lake Jänisjärvi (specimen Nos. 753, 4462, 7831, 7832, 7836—7838, 7841, 7842, 7844 and 8733) and Lake Mien (specimen Nos. A 4870 and A 4871), which the present author has studied for reference. The reference material is from the collections of the Department of Geology and Mineralogy of the University of Helsinki.

The normative mineral compositions of the unaltered Lappajärvi rocks were calculated on the basis of chemical analyses using the scheme of calculation suggested by Rittmann (1973). The results are presented in Table 10. In the quartz — alkali feldspar — plagioclase triangle, the points representing the unaltered rocks lie near each other and thus, according to the classification of Streckeisen (1967), specimen No. 14 is to be named a rhyolite and Nos. 38, 69 and the specimen labeled "dacite" by Eskola (1921) rhyodacite. This means that they all lie close to the center of the granite field of plutonic rocks. Because chemical equilibrium was not reached in the kärnäite, orthopyroxene was not transformed into clinopyroxene. Thus, the calculated norm (the Rittmann norm) agrees well with the modal compositions observed in thin sections and in X-ray powder patterns. The only exception seems to be that the Rittmann norm of cordierite is too high for most specimens and that of orthopyroxene too low. This discrepancy is corrected by changing some cordierite back into orthopyroxene, a procedure which would also increase the amount of normative quartz.

Cobalt and nickel in the rocks

The average Co/Ni ratio in meteoritic material is 0.06 and, especially in the iron meteorites, 0.07 (Mason, 1962, Table 18 and p. 131). In crustal rocks such as shales and basalts the ratio is 0.21 and 0.32, respectively, and in granite

Table 11

Co and Ni contents. Analyst, Tapio Koljonen

No.	Rock type	Co (ppm)	Ni (ppm)
Off-structure specimens:			
15	Mica gneiss	5	6
16	Muscovite granite	12	6
150	Mica gneiss	≤ 1	30
Shocked specimens:			
14	Kärnäite, Type I	4	40
17	Suevite	3	40
38	Kärnäite, Type I	5	350
67	Suevite	≤ 1	1 400
69	Kärnäite, Type II	≤ 1	200
72	»	≤ 1	70
74	Suevite	5	1 400
94	»	3	200
108	»	3	100
117	»	7	10
236	»	5	280

as high as 2 (Krauskopf, 1967, Appendix III). In addition, the content of cobalt and nickel is very low in granitic rocks, only a few ppm at the most. Thus, if, for example, an iron meteorite containing about 8 weight % nickel strikes a granitic target rock and produces an explosion crater, it should be possible to detect a change in the Co/Ni ratio or in the content of nickel, or both, in the crater rocks produced by the higher stages of shock metamorphism.

Cobalt and nickel were determined in seven suevite specimens, four kärnäite specimens and three unshocked specimens collected from the nearby bedrock outcrops. The results are listed in Table 11. The cobalt and nickel contents in the unshocked specimens are similar to the contents in granites and mica gneiss. Lonka (1967, Table 5) has determined from mica schist at Seinäjoki, southwest of Lappajärvi, the following contents: average cobalt content (12 ± 3) ppm, nickel content (39 ± 15) ppm. In mica schists and phyllites occurring in the same Bothnian schist belt but northwest of the lake corresponding contents are (9 ± 3) ppm Co, (44 ± 22) ppm Ni.

Relatively high nickel contents and even iron-nickel spherules occur in impactites at some meteorite craters, e.g. Henbury and Wabar (El Goresy *et al.*, 1968, pp. 532—536) and in lunar soil. An iron meteorite certainly contaminates chemically the strongly shocked crater rocks as is observed by the increased nickel content. Currie (1971) summarized the chemical composition of rocks from Canadian astroblemes and gave trace element data for some crater rocks. A definite or strong increase in nickel and chromium was observed in the shock-metamorphosed rocks at the Brent, East Clearwater Lake and Carswell Lake

craters. The cobalt content of these rocks remained almost unchanged, which is consistent with the supposed contamination by meteoritic material. Stähle (1970) reported that nickel, unlike cobalt, is very irregularly distributed in the glass bombs in the suevite of the Ries Crater. In addition, a small but real increase in nickel seems to be associated with the making of the glass bombs. Stanfors (1973, p. 56) reported an average nickel content of 15 ppm for the basement of Lake Mien and of 40 ppm for the "rhyolite" specimens from the crater. The increase in average cobalt content from 20 ppm Co in the basement to 70 ppm Co in the "rhyolite" is, however, considerable and thus the change in the Co/Ni ratio is not consistent with the effect of meteoritic contamination.

Table 11 shows that the cobalt content of the Lappajärvi rocks does not change markedly. Consequently, there is no connection between the cobalt content and the effects of shock metamorphism. There is, however, a large nickel content in the suevite and kärnäite specimens, but it is very irregularly scattered over a large range of about 1400 ppm. The mean of eleven determinations is 370 ppm Ni, which is about 10 times as high as the suggested average nickel content in the surrounding bedrock (target rock). To produce an explosion crater having a diameter of 8 km would require an iron meteorite with a velocity of, say, 40 km/sec and a diameter of, say, 230 m (see p. 29). The meteorite would vaporize and its material would be partly mixed with the fallout and fallback breccias, and with the impact melt. If all the nickel in the meteorite were to be mixed with the most strongly shocked target rocks at Lappajärvi comprising, say, 5 % of the crater volume, the average nickel content of these rocks would rise from about 40 ppm to about 500 ppm. At the same time the average cobalt content should have risen to about 40 ppm. Why the cobalt content of the Lappajärvi rocks seems to have remained unchanged is, however, still obscure.

RADIOMETRIC DATING

Lake Lappajärvi lies within the Precambrian Bothnian schist belt that is part of the Svecokarelian (Svecofennian) orogenic system. Svecokarelian intrusive rocks yield ages varying from 1750 Ma to 1900 Ma. For instance, for zircon from a granodiorite (gneissose granite) at Vimpeli near Lappajärvi, a discordant uranium-lead age of about 1900 Ma has been obtained (Kouvo, 1964, p. 15). Potassium-argon and rubidium-strontium ages for micas from the Svecokarelian schists are somewhat lower, about 1750 Ma to 1800 Ma (Kouvo, 1964, Table 1; Kouvo and Tilton, 1966, Tables 2 and 3), and these ages are believed to indicate the age of the last stage of metamorphic recrystallization (Simonen, 1964a, p. 92, 1964b, p. 121).

Professor Paul D. Fullagar, Department of Geology, University of North Carolina, in the U.S.A. studied a number of kärnäite and other shock-metamorphosed and unshocked Lappajärvi rocks for their whole-rock Rb/Sr ratios. Unfortunately, all the samples studied turned out to have a low and essentially similar Rb/Sr ratio (Paul D. Fullagar, 1972, written communication). This prevents the radiometric determination of their age by use of an isochron plot of $^{87}Rb/^{86}Sr$ and $^{87}Sr/^{86}Sr$ ratios. A type II kärnäite (No. 6506) was, however, analyzed and yielded a $(^{87}Sr/^{86}Sr)N$ ratio of 0.7373 and a $^{87}Rb/^{86}Sr$ ratio of 1.12. Although interpretation based on a single determination is doubtful, the result suggests, assuming a reasonable or a somewhat high initial $^{87}Sr/^{68}Sr$ ratio of 0.7090 and using the decay constant $\lambda^{87}Rb = 1.39 \times 10^{-11}y^{-1}$ (see Fullagar *et al.,* 1971, p. 437), (1) that this kärnäite may be a 1800 Ma-old igneous melt, or (2) that it may have been formed by the melting of a closed system consisting of 1800 Ma-old basement rock (Paul D. Fullagar, 1972, written communication). To get a lower age a still higher initial $^{87}Sr/^{86}Sr$ ratio must be assumed, and if an unlikely initial $^{87}Sr/^{86}Sr$ ratio of 0.7137 is assumed, an age of 1500 Ma is obtained.

The first explanation, viz. that kärnäite is an igneous melt 1800 Ma old, is ruled out by bedrock fragments of exactly this age in the kärnäite. It may be assumed that the isotopic constitution of rubidium and strontium results from intense assimilation (cf. Eskola 1921, p. 10), meaning that a young igneous melt has assimilated old basement rocks with a high $^{87}Sr/^{86}Sr$ ratio. Some recent rubidium-strontium and potassium-argon datings of meteorite impact structures (astroblemes), however, have revealed evidence strongly favoring the second explanation.

The rubidium-strontium and potassium-argon ages are known for four astroblemes: the Tenoumer Crater in Mauritania (French *et al.,* 1970, pp. 4402—4404), the Rochechouart structure in France (Kraut and French, 1971, p. 5409), the Charlevoix structure in Canada (Rondot, 1971, pp. 5419—5421) and the Sudbury structure in Canada (Fullagar *et al.,* 1971); the latter, however, which seems to have a complicated history (a later thermal event), is not treated here. In the three other structures mentioned, rubidium-strontium dating of the fused rock (impact melt) has yielded ages consistent with both rubidium-strontium and potassium-argon ages of the surrounding bedrock. Potassium-argon dating of the fused rock has yielded younger ages, considered to correspond to the ages of the three structure-forming events. For instance, "melt rock" samples from the Tenoumer Crater yielded rubidium-strontium ages closely agreeing with the rubidium-strontium and potassium-argon ages of the basement rock, which has an age of about 2000 Ma, whereas the potassium-argon age of the "melt rock" is only about 2.5 Ma (French *et al.,* 1970, pp. 4402—4404).

This means that nonvolatile elements like rubidium and strontium are retained in the fused rock during shock metamorphism, whereas argon has been almost totally degassed. Potassium-argon dating of rocks from the Brent Crater in Canada also revealed that shock-metamorphosed rocks containing shocked mineral fragments may yield low ages (Hartung *et al.,* 1971, p. 5444). Hartung and his co-workers stated that this is probably because the strong lattice distortion of shocked mineral grains reduces their ability to retain argon.

The advantage of rubidium-strontium dating is that it yields data indicating whether a lava-like rock is derived from the mantle (low $^{87}Sr/^{86}Sr$ ratio) or whether it results by the fusion of the crust (target rocks). Likewise, the observed discrepancy between the rubidium-strontium and potassium-argon ages of lava-like, shock-metamorphic rocks gives evidence of their meteorite impact origin.

To establish the age of the crater-forming event in the Lake Lappajärvi structure a number of potassium-argon ages, especially from Type I kärnäite or wholly recrystallized granite pegmatites with sanidine or anorthoclase, are required. Although the rubidium-strontium age of Type II kärnäite is not very reliable it supports the interpretation that the structure is a result of a meteorite impact.

SUMMARY AND CONCLUSIONS

Lappajärvi is a deeply eroded crater lake located within the Precambrian schist belt of southern Pohjanmaa. Near the lake the bedrock outcrops consist of coarse-grained granite and granite pegmatite, mica gneiss often migmatized by granite and granite pegmatite, medium-grained muscovite granite and granodiorite. No signs of shock metamorphism are visible in the specimens collected from the outcrops. Inside the crater proper only shock-produced melt rock with numerous bedrock and mineral fragments (kärnäite) crops out, but the glacial drift within the indicator fan consists chiefly of material at all stages of shock metamorphism.

The lake bottom is characterized by two inward-curved but partly parallel deep valleys running from NNW to SSE and by a shallower valley which divides the lake bottom into two parts. The eastern valley joins with a tectonic zone which also comprises the outlet of Lappajärvi. The other valleys do not fit well into the lineament pattern of the area but are explained as products of the preferential erosion of brecciated target rocks which, near the margins of the crater, have not been covered by the resistant layer of kärnäite.

An almost circular negative gravity anomaly of 10 mgal and 17 km in diameter is associated with the lake. The center of this anomaly coincides closely with the center of the crater. The aeromagnetic total-intensity map shows a

marked contrast between the intensities observed within and beyond the lake area, and the lake occupies an area of low magnetic relief. The negative gravity anomaly and the low magnetic relief support the interpretation of the crater as having been produced by meteorite impact. The crater is filled by low-density breccias (sp. gr. approx. 2.55—2.25) and kärnäite (sp. gr. 2.56). The specific gravity of the bedrock ranges from 2.61 for granite to 2.82 for mica gneiss.

Small fragments of imperfect shatter cones have been found in several weakly shocked mica gneiss and granite inclusions within the impact breccia. They are regarded as typical low-stage shock-metamorphic features.

A complete series of rocks of shock-metamorphic origin exists at Lappajärvi. Boulders and blocks of brecciated or weakly shocked bedrock are common. Smaller, brecciated or weakly shocked rock and mineral fragments are cemented by fine rock flour and zeolites to form a polymictic breccia, called impact breccia. This rock does not usually contain melted material.

High post-shock temperatures were associated with stronger shock waves, but, because of lack of time for equilibrium reactions and for the formation of eutectic melts and because of rapid particle motion in the strongly shocked material, each mineral grain tended to melt according to its own melting point. The resulting melt was rich in quartz particles and, when mixed with the numerous weakly and moderately shocked and partly melted rock and mineral fragments, a dark breccia with a lava-like groundmass (kärnäite) was produced.

The material of suevite was also affected by similar processes but was ejected from the crater, whereby a mixture was produced consisting of fragmental breccia and aerodynamically sculptured glassy bodies resembling volcanic bombs and lapilli. Small rock and mineral fragments were also captured by or wrapped in the melt during their passage through the air.

Some textures, such as the survival of fluidal and vesicular textures in mineral and rock glasses, and the presence of sharp-edged quartz particles with planar elements, indicate very rapid cooling of both rock types. The presence of many ghost-like remnants and the absence of phenocrysts in kärnäite also suggest that the shock waves caused the total and partial melting of the target rocks as well as the strong mixing of the material. The shock waves may also have produced some "superheated" melt which caused assimilation, but because of the rapid rates of mixing and cooling this was an unimportant process.

Two types of kärnäite are distinguished. Type I (the impact lava proper) has a colorless glassy groundmass with microlites of pyroxene, sanidine, anorthoclase and cordierite. Type II, the more common type, is very heterogeneous. It is interpreted as a firmly welded mixture of shock-produced melt, glassy bodies and shocked mineral and rock particles. It is a transitional rock between Type I kärnäite and suevite. As the amount of glass fragments of fluidal texture decreases the suevite passes gradually into the impact breccia.

Petrographic information favoring the impact origin of Lake Lappajärvi is strongly supported by mineralogical evidence that includes all the main types of residual shock effects: fracturing, plastic deformation, solid-state transformation and thermally induced transformation.

Planar fractures and planar elements (shock lamellae) occur in quartz, feldspars and biotite, and in accessory minerals (apatite, sillimanite, zircon and garnet). The planar elements are formed at peak pressures exceeding the Hugoniot elastic limit of the mineral, usually within the 100 kb to 350 kb range. Orientations of multiple, closely spaced sets of planar elements, both decorated and non-decorated, were measured on the universal stage and are presented for quartz, microcline, apatite, sillimanite and zircon. In these diaplectic minerals, the gradual decrease of birefringence and refractive indices is associated with the increasing number and intensity of the planar elements. The impact breccia, kärnäite and suevite contain minerals with planar elements, but the elements are best preserved in the impact breccia.

The solid-state transformations that require pressures in the 300 kb to 500 kb range include the formation of diaplectic (thetomorphic) glasses of quartz and feldspars. Diaplectic plagioclase glass (maskelynite) is very common in granite fragments within the impact breccia. Maskelynite is usually accompanied by diaplectic quartz and microline. Diaplectic glasses of quartz and microline are rare. Thus, in coarse- or medium-grained granitic rocks, quartz and microline are more resistant to shock compression than is plagioclase (oligoclase).

Combined X-ray diffraction and infrared absorption were applied to study the degree of Al/Si order in feldspars. Weakly or moderately shocked alkali feldspars, with planar elements and patches of diaplectic glass, have an intermediate, strongly varying degree of Al/Si order. Solid-state transformation also markedly affects those infrared absorption bands that result from O-Si(Al)-O and O-Si-O bending vibrations of the feldspars lattice. Strongly or intensely shocked alkali feldspars, recrystallized from thermally induced liquid-state glasses, have a completely random Al/Si distribution.

The most convincing example of solid-state transformations in the Lappajärvi rocks is the presence of coesite. It was separated by means of hydrofluoric and nitric acid, and its presence was verified by X-ray diffraction and infrared absorption. Coesite has so far been identified from only ten impact craters, and its presence in strongly shocked rocks is regarded as reliable evidence of hypervelocity meteorite impact.

The glass fragments of fluidal texture in suevite are products of very rapid high-temperature melting. The glass contains tiny mineral fragments at all stages of shock metamorphism, even serpent-shaped colorless strings of lechatelierite. The liquid-state glasses have in places recrystallized, and consequently their detailed investigation is difficult, even though their gross textures are recognizable.

In several kärnäite and suevite specimens thermally decomposed biotite has been transformed into tiny crystals of spinel with very small unit-cell parameters.

Calcite is the most common secondary mineral in kärnäite vesicles. Zeolites are almost omnipresent in cavities and fissures in suevite and impact breccia. Heulandite, stilbite, chabazite, mordenite and erionite have been identified in the breccias by X-ray diffraction.

Three new chemical analyses are presented of the shock-metamorphosed Lappajärvi rocks. The analyses and the normative mineral compositions (Rittmann norms) show that there is no reason to invoke lavas or other volcanic agencies to explain the chemical composition of the Lappajärvi rocks.

Nickel is very irregularly distributed in suevite and kärnäite. The average nickel content, unlike that of cobalt, has clearly increased, being about ten times as high as in the target rocks. The strong relative increase of the nickel content in the shock-metamorphosed rocks suggets contamination by a nickel-rich iron meteorite.

Whole-rock Rb/Sr ratios were determined for a number of Lappajärvi rocks. Low and essentially the same Rb/Sr ratio prevents dating by use at an isochron plot. The isotope ratios, however, suggest that kärnäite could have been formed by the melting of 1 800 Ma-old target rocks in a closed system.

Until recently, Finnish geologists have believed that Lappajärvi is of volcanic origin. Svensson (1968) published evidence in favor of meteorite impact origin, but based his conclusions mainly on the existence of planar elements (shock lamellae) in quartz grains in kärnäite. McCall (1968) questioned this evidence and suggested an endogenic origin involving highly explosive carbonatite or alkalic volcanism.

A complete series of shock-metamorphosed rocks exhibiting all the stages of shock metamorphism and containing a great number of indisputable mineralogical and petrological criteria, together with chemical and geophysical data and analogies with several well-established meteorite craters are regarded as convincing evidence that Lake Lappajärvi is a meteorite impact site (astrobleme). The structure and the shock-metamorphosed rocks were formed by an enormous explosion and strong shock waves. The only natural mechanism known to produce such an explosion is a hypervelocity meteorite impact.

No evidence in favor of McCall's proposal has been found at Lappajärvi and the present author would like to concur with Barringer (1905, p. 873) who wrote of Meteor Crater (Barringer Crater) in Arizona: "Briefly, it would seem to me to be impossible that any geologist carefully examining the region could reach the conclusion that this is a volcanic crater, or in any way produced by volcanic agencies".

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