INTEGRATED GEOPHYSICAL STUDY OF
THE KEURUSSELKÄ IMPACT STRUCTURE, FINLAND

Selen Raiskila

HELSINKI 2013
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Cover picture:
Lake Keurusselkä

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Selen Raiskila

ACADEMIC DISSERTATION IN GEOPHYSICS

To be presented, with the permission of the Faculty of Science of the University of Helsinki, for public examination in the Auditorium E204 of Physicum, Gustaf Hällströminkatu 2a,
on January 11th, 2013, at 12 o’clock noon.

Helsinki 2013
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Abstract

Meteorite impact cratering has played a key role in Earth's geological past and has left a dramatic effect on biological and geological records forming large volumes of igneous rocks and important mineral deposits. By studying terrestrial impact craters we can have valuable information of impact generated changes in rocks and minerals. Integrated geophysical study in this dissertation combines different methods to distinguish the meteorite impact related features from features caused by endogenic processes. Optical microscopy provides diagnostic evidence of shock produced deformations in minerals. Shocked and un-shocked rocks have contrasts in petrophysical properties, which cause anomalies to regional magnetic and gravity fields over meteorite impact structures. Magnetic minerals may re-magnetize during impact and thus provide information of the past geomagnetic field and the ancient paleoposition of impact site.

This dissertation focuses to the Finnish impact structure, Keurusselkä, which was discovered in 2003. The structure is situated in central Finland (62°08’ N, 24°37’ E) within the Central Finland Granitoid Complex, which formed 1890–1860 million years ago during the Svecofennian orogeny. Keurusselkä is deeply eroded remain of a complex crater, named after the Lake Keurusselkä, which is the dominant present day feature of the crater area.

For this study rock and drill core samples were collected from Keurusselkä region. The samples were chosen according to a sampling strategy, where samples were taken in and outside the impact region to investigate the impact related features and their radial distance from the centre of the impact. The main focus was to prove or disprove the impact origin of Keurusselkä.

First indication of impact was shatter cones, which are conical features in rocks formed by an impact shock and pressure. Shatter cones were found in a circular area interpreted as the central uplift of the original crater. In this dissertation petrographic analysis of thin sections was done to find evidence of deformational features in minerals. Diagnostic evidence for Keurusselkä was achieved when planar deformation features (PDFs) were found from quartz grains in shatter cones indicating impact pressures up to 20 GPa. Samples from 3 shallow drillings in southwest part of the Keurusselkä structure were also studied. Impact like monomictic breccia was found from one of the drill cores. Petrographic analysis revealed clay minerals (illite, smectite-group), which most likely have altered from impact glass.

Magnetic minerals were obtained for paleomagnetic purposes, i.e. to examine the ancient position and drift of Baltica. Paleomagnetic directions obtained from shatter cones indicate re-magnetization. The virtual geomagnetic pole implies that the impact event took place ~1120 million years ago during the formation of Rodinia supercontinent.

Petrophysical properties (density, susceptibility) of rocks were measured for differences between shocked and un-shocked rocks. The exposed bedrock in Keurusselkä was noticed to be fractured and damaged, which causes anomalies to the regional magnetic and gravity field. The geophysical signals were modelled along two profiles using measured physical properties of Keurusselkä rocks. The highly magnetized centre of
the structure forms an eroded circular central uplift with diameter of 6 km. Based on the gravity minimum around the structure an area of ~16 km in diameter and depth of ~1200 meters was modelled to explain the observed anomalies. The original size of the crater is estimated to be 24-27 km in diameter.
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It would not have been possible to complete this doctoral thesis without the help and support of many colleagues and friends around.

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Lastly, I offer my regards and thanks to all of those who supported and assisted me in any respect during the completion of this PhD project.
List of original publications

This thesis is based on the following publications, which are referred to in the text by their roman numerals:


Papers I and II are reprinted from *Meteoritics and Planetary Science* with permission from John Wiley & Sons Ltd. Paper III is reprinted from *Studia Geophysica et Geodaetica* with permission from Springer. Paper IV is reprinted from *Geophysica* with permission from the Geophysical Society of Finland.
Abbreviations

AM  Aeromagnetic
Ar  Argon
ChRM characteristic remanent magnetization
DEM Digital Elevation Model
Gal Galilean (unit of acceleration of gravity)
GTK Geological Survey of Finland
IGRF the International Geomagnetic Reference Field
Ma mega anno (million years)
NRM natural remanent magnetization
PEF present Earth’s magnetic field direction or intensity
SMOY Suomen Malmi OY (Finnish Mining Company)
TRM thermoremanent magnetization

κ  volume magnetic susceptibility ($10^{-6}$ SI)
ρ  density (kg/m$^3$)
φ  porosity (%)

$H_C$ coercive field (T)
$H_{CR}$ coercivity of remanence (T)
$M_{RS}$ mass normalized saturation remanent magnetization (mAm$^2$/kg)
$M_S$ mass normalized saturation magnetization (mAm$^2$/kg)
$T_C$ Curie temperature (°C)

$a_{05}$ radius of the circle of 95% confidence of the mean paleomagnetic pole
$A_{05}$ radius of the circle of 95% confidence of the paleomagnetic pole
Dec. declination (°)
Inc. inclination (°)
dp, dm Semi-axis of the 95% confidence oval of the paleomagnetic pole
$k$ Fisherian precision parameter (Fisher, 1953)
$P_{lat}$, $P_{lon}$ latitude and longitude of the paleomagnetic pole (°)
1 Introduction

A meteorite impact crater is a structure created by a meteorite, which is an extra-terrestrial object originated in space hitting Earth. A glimpse to the Moon and its craters reveal us the violent past of meteor impacts, which play a key role in the evolution of our Solar System. A terrestrial impact crater is formed when a falling meteorite, large enough to survive the atmospheric entry, hits the Earth’s surface. Despite the growing research interest towards terrestrial meteorite craters the impact cratering does not perhaps get the attention it deserves in geoscience community and in the school education. The ability to identify meteorite impact structures from endogenic geological processes will expand our knowledge to understand the complete geological evolution of Earth and how impacts have affected and reshaped our planet through the times. Identification of terrestrial impact structures will also give a better understanding of environmental effects of impact events and their role in formation of life on Earth. The link between 65 Ma old Chicxulub impact structure in Mexico (e.g. Sharpton et al. 1991; Sharpton et al. 1992; Alvarez et al. 1995) and the K/T boundary (Cretaceous-Paleogene extinction event) shows the significance and consequences of an impact event. Statistical database of impact craters would provide an estimate for impact rate on Earth and possibly give explanation to e.g. large volumes of igneous rocks, important mineral and hydrocarbon deposits (e.g. Grieve and Masaitis 1994), and biological extinctions in geologic record (Koeberl and MacLeod 2002) as well as widespread perturbations in the Earth’s crust (e.g. Henkel and Reimold 1998).

Currently there are 182 confirmed impact structures on Earth: 59 in North America, 38 in Europe, 30 in Asia, 27 in Australia, 19 in Africa and 9 in South America (Earth impact database 2012) (Figure 1). The majority of structures have been found on continental shield areas and some even on ocean floor. Oldest and largest impact structure on Earth is Vredefort structure in South Africa. Vredefort impact event took place ~2.02 billion years ago and the crater has a diameter of ~200 km (e.g. Dietz 1961; Kamo et al. 1996; Turtle and Pierazzo 1998). Latest impact event on Earth happened in September 2007 when a chondritic (H4-5) meteorite hit the ground in Peru creating a 14 meters wide and 4.5 meters deep crater called Carangas (e.g. Kenkmann et al. 2009). Impact cratering is, thus, a current and on-going phenomenon while large and catastrophic events, e.g. comet Shoemaker-Levy 9 impacting on Jupiter in 1994 (e.g. Zahnle and Low 1994), have become rare.

Precambrian shield areas (3800-544 Ma) are ideal for discovering traces of impacts as the bedrock is mostly exposed (Figure 1). Majority of the oldest impact structures have been found on these shield areas, although, many of the Precambrian rocks are eroded or metamorphosed and sometimes deeply buried (Levin, 2006). Another reason for high quantity of impact craters is that the impact crater rate on Earth during Precambrian times was higher.

Due to a high level research activity in Fennoscandia, it is one of the most densely mapped areas with impact craters on Earth (Dypvik et al. 2008) and 31 impact structures have been either confirmed or under investigation (Figure 2). Ages of Fennoscandian
impacts vary from 4000 years to 1800 million years. There are 11 proven impact structures in Finland. Most of the impacts have been found through ore and Kimberlite exploration, where strikingly circular geophysical anomalies have been investigated and drilled. In some cases like Suvasvesi N structure (Pesonen 1996) these anomalies were caused by impact generated lithology, such as melt layers or breccia. Diameters of Finnish impact structures vary from 1 km to 23 km and the ages 74–1800 million years. The relatively large number of impact structures found in Finland is not only a result of exploration activity, but also an output of a search strategy where ground surveys, petrographic investigations and remote sensing methods are integrated.

![World impact map with 182 confirmed impact structures](image)

**Figure 1** World impact map with 182 confirmed impact structures (Earth Impact Database 2012). Main shield areas from Levin (2006).

Meteorite impact craters can be studied in numerous ways. **Macroscopic studies** involve mapping of the impact area and sampling of impact produced rocks (melt and breccia), interpreting geophysical anomalies related to the impact and using remote sensing methods to obtain crater dimensions (e.g. Henkel and Reimold 1998; Plado et al. 1999). **Microscopic studies** concentrate on impact generated mineralogical features and geochemical changes in rocks (e.g. French and Koeberl 2010; Deutch and Lagenhorst 1998). Some features related to meteorite impact structures are not unique and can be products of tectonic deformation, volcanic eruptions or igneous activity (Kimberlite piping). All these processes produce circular forms and patterns of deformation, extensive fracturing and brecciation of the target rock, circular geophysical anomalies and large units of igneous rocks. Therefore, a careful petrographic analysis of sampled rocks combined with geophysical and geochemical features is needed. In this dissertation some
of these methods are used to show the impact origin for Keurusselkä and to explain the geophysical anomalies related to the structure.

![Fennoscandian impact structure map](modified from Dypvik et al. 2008)

**Figure 2**  Fennoscandian impact structure map (modified from Dypvik et al. 2008).

The structure of this dissertation is as follows: first I will start by briefly introducing the impact cratering as a process, the crater types based on crater morphology and the distinct impact rocks related to meteorite impact structures. With addressing impact related features, which diagnose an impact structure, I will discuss the generally approved criteria to confirm an impact origin. Then, I will describe the methods and instrumentation used in this study. After that, I will give a short summary of the four original articles I-IV included in this dissertation with conclusions and an updated review of research results from the Keurusselkä structure.

### 1.1 Impact cratering process

Most of the objects entering Earth’s atmosphere originate from asteroid collisions and consist of stone, iron or a mixture of the two. Small particles burn up in the atmosphere, whereas larger ones can survive and hit the ground as meteorites. The formation of an
impact crater is a complex process dependent of several matters: (1) type, size and velocity of the projectile with sufficient kinetic energy to form a crater, (2) impact angle and (3) target material (crystalline, sediments, water or ice). Many details of cratering process are still uncertain and not well understood (see e.g. Melosh 1989; French 1998).

The impact event of a large body is extremely rapid process where the crater formation mechanisms are mainly triggered by extreme pressures (a few to hundreds of gigapascals) generating shock waves, which penetrate into a target material. A typical large stony projectile has a diameter between 0.5–10 km, mass of $10^9$–$10^{16}$ kg, velocity 20–40 kms$^{-1}$ and kinetic energy of $10^{15}$–$10^{20}$ J (French and Koeberl 2010).

The process producing an impact crater is divided into four physical stages: (1) contact and compression, (2) excavation, (3) modification and (4) collapse (Melosh 1989; French 1998; French and Koeberl 2010; Collins et al 2012). The entire process, including micro-scale deformation to mega-scale disruption and perturbations, is a rapid process lasting only a few minutes (Figure 3). The stages grade into one another, but the purpose for separating them is to comprehend the different impact produced features related to the process. The first stage starts when the projectile contacts the target material. A moving projectile penetrates at most 1-2 times of its own diameter into the target. At the impact point the projectile’s kinetic energy is transferred by shock waves, which spread radially in the target. This rapid contact and compression stage grades into excavation stage, which lasts during the time when the complex interactions and processes fracture and shatter the target material. The transient crater starts to form and material is ejected upwards and outside of the crater rim, which is uplifted around the excavating cavity. Some of the ejected material fall back into the crater and is mixed with un-shocked materials from the collapsing rim. At deeper levels of excavation the transient crater material is driven also downward. The transient cavity is typically 10– 20 times the diameter of the projectile (French and Koeberl 2010). After the shock and release wave energies have decayed into low-pressure elastic stress waves and do not excavate or displace material anymore the transient crater has reached its extent. Along the transient crater floor the shock pressures may exceed 25-30 GPa (French 1998). The pressures drop rapidly ~25 GPa every few kilometres down from the impact center (French 1998 and references therein). The formed crater is then modified by gravity and rock mechanics, which is the final stage in crater process. During this modification stage most of the impact-related changes occur such as the formation of central uplift and the crater collapse. The formed crater is filled with crater-fill deposits of allochthonous (allogenic) rocks (breccia and melt). The allochthonous impactite units consists a mixture of (1) fall-back ejecta, (2) small and large bodies of impact breccia and melt, (3) small and large fragments of unshocked target rocks from collapsed rim and (4) ejecta originally deposited on the rim, which slide during the rim collapse (French 1998). Although the cratering process is rapid, the post-shock processes have longer tendencies (e.g. temperature dependent hydrothermal alteration processes) (Naumov 2005 and impact structures e.g. Chixculub, Mexico - Abramov and Kring 2007; Bosumtwi, Ghana - Karikari et al. 2007; Kärdla, Estonia - Jõeleht et al. 2005; Charlevoix, Canada - Trepmann et al. 2005; Sudbury, Canada – Abramov and Kring 2004; Siljan, Sweden - Hode et al. 2003; Haughton, Canada - Osinski et al. 2001; Vredefort, South Africa - Gibson et al. 1998; Roter Kamm, Namibia - Koeberl et al. 1989).
Figure 3  The impact process of a simple and complex crater shown from numerical model by Dr. Kai Wünnemann (modified from Collins et al. 2012). In the case of a simple crater formation the projectile size is 50 m and in complex crater case 500 m. Both projectiles have the same velocity 12 kms$^{-1}$. The target rock and projectile material is quartzite. Figures (A) and (B) show the excavation stage, where the transient crater is developed, (C) the modification stage and (D) the final crater form.

The theoretical impact process model is typically described as a vertical impact to the target rock. However, an oblique impact is more probable. In oblique impacts the shock wave weakens with decreasing impact angle and produces an asymmetric form (Collins et al. 2012). This is most clearly seen in the distribution of ejecta pattern, which shape is a function of impact angle. Highly oblique impacts form e.g. butterfly patterns as seen around some of the craters on the Moon (Herrick and Forsberg-Taylor 2010). On Earth
ejecta patterns are rarely preserved. The final crater form on terrestrial structures can be elongated along the direction of the trajectory of the projectile, if the impact is sufficiently oblique. The structural asymmetries are largest in the central uplift area and a detailed structural analysis may reveal the impact direction (Scherler et al. 2006).

1.1.1 Simple, complex and multi-ring craters

The known impact structures on Earth range from a small circular pit to a large complex or multi-ring structures. All impact structures regardless of their final sizes are assumed to undergo the transient-crater stage. The range to which the transient crater is modified during the modification stage depends on its size.

The classic type of a crater is a bowl-shaped form called “simple crater”. These craters are usually small in size (less than few kilometres in diameter) and they have changed only little from the original transient crater. In larger craters modification and collapse stages can involve significant structural changes, such as the uplift of the central part of the crater floor and major collapse around the rim. Depending on the degree to which the crater is modified, larger craters are characterized as complex craters, with a central uplift or a peak-ring or multi-ring basins (Melos 1989; French 1998) (Figure 4). Simple, complex and peak-ring type structures are generally found on Earth while large multi-ring structures are found e.g. on the Moon surface.

![Figure 4](image-url) Simple line drawings showing the morphology of different crater types and sizes (modified after Spray 2002).
1.1.2 Impact rocks

Unusual rock types, such as breccias, melts, ejecta deposits, shatter cones and fractured bedrock, are formed during impact processes in and around the crater area by the high impact pressures and temperatures.

The stratigraphy of the impact rocks depends on the composition of target rocks (crystalline, sediments) and the target environment (land, sea or ice). The target environment affects the final crater morphology and especially the marine impacts have sedimentary features related to the resurge of water into the crater (Dypvik et al. 2004). Impacts on crystalline and sedimentary targets have equivalent stratigraphy in crater-fill deposits (Osinski et al. 2008). The impact rocks in all these cases are similar enough for classification.

![Figure 5](image)

**Figure 5** Complex crater form showing the impact structure with impact lithology (modified from Stöffler and Grieve 2007). The annular depression around central uplift is filled with lithic breccias, suevite, melt rock and fallback breccia. The crater basement is brecciated with intrusive melt veins and dykes. The bottom of the transient crater has monomictictly brecciated basement.

Allochthonous (Allogenic) rock units, such as breccias, melt rocks and breccia dykes fill the crater and form the ejecta while parautochthonous rocks are located beneath the crater floor (French 1998). Layered allochthonous impact rocks in the ejecta blanket can extend 2 to 3 crater radii (Stöffler and Grieve 2007). Allochthonous impact rocks are characterized by more diversified lithology with fragmental or melted character and a wide range of shock features. Parautochthonous impact rocks have experienced relatively lower shock pressures, which have limited the shock features into fracturing, brecciation and formation of shatter cones (Figure 6). Shatter cones are typically found below the crater floor in the central uplifts of complex impact structures where they generally point upward (French 1998).

High-pressure mineral deformation may occur in relatively small volume (French 1998). Tektites are a special type of impact rocks, which are glassy distal impactites.
During a hypervelocity impact some of the ejected material (melted and partly vaporized) cool to form mm-to-cm size Tektite bodies of molten material. Impact breccia and melt are very vulnerable to alteration and erosion unless they are buried under sediments. Therefore old (>500 Ma) impact structures on crystalline shield areas are usually deeply eroded lacking distinct impact lithology. Impact structures formed in water or marine environment are quickly deposited by sediments, which preserve the crater from erosion e.g. Mjølnir structure in Barents Sea (Dypvik et al. 2004).

Figure 6  (A) Keurusselkä shatter cone in metavolcanic rock boulder (photo by Dr. Viktor Hoffman, University of Tübingen), (B) and (C) upward pointing Keurusselkä shatter cones, (D) damaged and fractured bedrock in Keurusselkä, (E) Vilppula drill core samples and (F) monomictic breccia from Vilppula drill core (photos by Selen Raiskila).
Impact lithologies are classified into three groups (Stöffler and Grieve 2007): (1) shocked rock – defined as non-brecciated rocks, (2) impact melt rocks – divided subgroups according to the content of clasts and melt and (3) impact breccias – divided according to the degree of mixing various target lithologies and melt particles (Figures 7 and 8).

Figure 7  Vredefort structure, South Africa: (A) Shatter cones in a rock boulder, (B) In-situ shatter cone, (C) Impact melt breccia and (D) Pseudotachylite vein. Ries structure, Germany: (E) Megablock in Suevite and (F) Breccia sample in ZERIN (photos by Selen Raiskila).
Figure 8  Classification of impactites (French 1998.)
1.1.3 Shock metamorphism

To identify meteorite impact structures and to distinguish them from structures formed by endogenic geological processes diagnostic criteria of impact produced features are essential (French and Koeberl 2010). The positive identification of an impact structure is done from petrographic or geochemical analysis of shock-metamorphic effects in minerals, which can preserve in target material for $10^6 – 10^9$ years (e.g. Deutch and Lagenhorst 1998; French 1998; French and Koeberl 2010).

Most of the rocks in impact structures show only non-diagnostic, low level deformation features, while distinctive and diagnostic shock effects are restricted to relatively small specific areas of the structure (Table 1). Near the impact point, initial shock pressure conditions can exceed 100 GPa, which are an order of magnitude higher than in normal geological processes. End results of such pressures are melting (>60 GPa pressures with post-shock temperatures exceeding 2000°C) and vaporization of target rock material.

Table 1  Shock-produced effects for meteorite impact structures (after French and Koeberl, 2010). Features related to Keurusselkä impact structure are marked with *.

<table>
<thead>
<tr>
<th>A. Diagnostic indicators for shock metamorphism</th>
</tr>
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<tbody>
<tr>
<td>1. Preserved meteorite fragments</td>
</tr>
<tr>
<td>2. Chemical and isotopic projectile signatures</td>
</tr>
<tr>
<td>3. Shatter cones (* Article I, II and IV)</td>
</tr>
<tr>
<td>4. High-pressure (diaplectic) mineral glasses</td>
</tr>
<tr>
<td>5. High-pressure mineral phases</td>
</tr>
<tr>
<td>6. High-temperature glasses and melts (* Article III)</td>
</tr>
<tr>
<td>7. Planar features (PFs) in quartz (* Article I)</td>
</tr>
<tr>
<td>8. Planar deformation features (PDFs) in quartz (* Article I)</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>B. Non-diagnostic features produced by an meteorite impact or by endogenic processes</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Circular morphology (* Article IV)</td>
</tr>
<tr>
<td>2. Circular structural deformation</td>
</tr>
<tr>
<td>3. Circular geophysical anomalies (* Article II-IV)</td>
</tr>
<tr>
<td>4. Fracturing and brecciation (* Article III and IV)</td>
</tr>
<tr>
<td>5. Kink banding in micas</td>
</tr>
<tr>
<td>6. Mosaicism in crystals</td>
</tr>
<tr>
<td>7. Pseudotachylite and pseudotachylitic breccias (* Schmieder et al. 2009)</td>
</tr>
<tr>
<td>8. Igneous rocks and glasses</td>
</tr>
<tr>
<td>9. Spherules and microspherules</td>
</tr>
<tr>
<td>10. Other problematic criteria (* Article II-IV)</td>
</tr>
</tbody>
</table>
According to French and Koeberl (2010) lower pressures produce also distinct shock-metamorphic deformations in minerals, such as selective mineral melting at 40–60 GPa, diaplectic glass phase at 30–45 GPa, high-pressure minerals at 12–30 GPa (e.g. stishovite, coesite), planar deformation features in quartz at 10–25 GPa, multiple fracturing and basal Brazil twinning at 5–10 GPa and shatter cones with rock fracturing at 2–5 GPa. Beyond the eventual crater rim, the shock waves decay to lower pressures (<1 GPa), which transform into normal elastic (seismic) waves. As a result, search for rock types which contain above mentioned diagnostic indicators can be most successfully found in (1) crater-fill breccias in well-preserved craters: discrete inclusions of shocked rock and melt and (2) near-surface zone beneath the crater floor in more deeply eroded craters: shattered sub-crater parautochthonous breccias.

1.2 Aims of this study

The Keurusselkä impact structure was found in 2003 (Hietala and Moilanen 2004; 2007). The first recognized impact features were shatter cones, which are conical striated surfaces within target rocks (Figure 9). The shatter cones appear in a circular area with a diameter of ca. 10 km, indicating a central uplift of a complex crater structure. The bedrock in this area is also thoroughly fractured and damaged compared to the bedrock outside the impact area. The crater formation was soon noticed to be deeply eroded as no crater rim, melt deposits or breccia lens were found. Preliminary age estimate suggested an age between 1000 Ma (erosion state of the structure) and 1880 Ma (crystallization age of the target rock granitoids). Original crater diameter was proposed to be 10 to 30 km.

An integrated geophysical research project was established in 2007 to study preserved impact features of the Keurusselkä structure, of which this dissertation is a part. The challenge in this study was to prove or disprove whether Keurusselkä is an impact structure or only an endogenic formation. The structure seemed to be deeply eroded lacking impact lithology. Even though the quantity of shatter cones was abundant and their appearance remarkable, no other features supported the impact origin. Petrophysical properties of sampled rocks, remote-sensing observations and geophysical studies were considered to investigate the impact dimensions. Although this dissertation is based on investigation of geophysical features some geological observations (petrographic analyses) are included. Identification of an impact crater must be based on a set of shock-metamorphic features that are unique and distinguishable from endogenic geological features; e.g. planar deformation features (PDF’s) in quartz grains are unambiguous features in minerals caused by a meteorite impact. Therefore, the first aim of this dissertation was to investigate the shock-metamorphic features in Keurusselkä rocks.
Figure 9  (A) Shatter cone sample and (B) Location of the Keurusselkä impact structure. In-situ shatter cones appear in the central uplift (sites marked with black stars). Sampling sites are marked with white dots. Sites with arrow are sampled along the shear zones outside the map region (Article II).

The second objective was to collect rock samples and to establish a petrophysical database for Keurusselkä impact rocks and various target rocks to investigate whether their physical properties differ from rock samples gathered well outside the impact area. In order to date the impact event and to obtain a pole for Baltica paleomagnetic measurements and component analysis of remanent magnetization directions were planned.

The third objective was to investigate the three drill cores to look for impact related features and lithology. The drill cores are located nearby the centre of the crater and they are drilled during the 1968 ore exploration campaign in Keurusselkä area.

The fourth objective was to model the geophysical anomalies (gravity and aeromagnetic) encountered in the Keurusselkä region and to describe their connection to the impact event. The model was aimed to base on the petrophysical properties of rocks combined with seismic velocities of rock samples. Illustration of a FIRE 2 seismic profile interpretation (Kukkonen and Laitinen 2006) was used to shape the modelled source bodies beneath the surface to fit the calculated anomaly signal to the observed one. The aim of estimating the size of the original crater diameter was to study the crater morphology and the geophysical modelling results.
2 Sampling and methods

2.1 Earth’s magnetic field

In this dissertation, the magnetic properties of rocks are examined to obtain knowledge of magnetic minerals and their magnetization origin. Rocks acquire a magnetization in the Earth’s main magnetic field, which is driven by a dynamo processes in the liquid outer core. The dipole field is the most important field at the surface of the Earth for paleomagnetic purposes. The non-dipole field is about 5% of the total field (Lowrie 2002). The point where the Earth’s magnetic field is vertical is called magnetic dip pole. The tilt between the rotation axis of Earth and the magnetic dip pole is about 11.5°. The magnetic pole wanders, due to secular variation of intensity and direction of the field, around the geographic North Pole with a speed of few degrees in a century. The secular variation when averaged out over geologic time scale (in millions of years) leads to a model of geocentric axial dipole (GAD) shown in Figure 10. It describes a magnetic dipole M at the center of the Earth aligned with the Earth’s rotation axis (Lowrie 2002). The paleolatitude \( \lambda \) of the sampling site can be calculated from the magnetic inclination \( I \) using the basic dipole equation:

\[
\tan I = 2 \tan \lambda
\]

The GAD has a fundamental application in paleomagnetism since measurement of the inclination \( I \) of the remanent magnetization of rocks provide an estimate of the ancient latitude (paleolatitude \( \lambda \)) (see section 2.3). Declination in the GAD model is 0° everywhere due to axisymmetry of the GAD (e.g. Irving and Ward 1964).

![Geocentric Axial Dipole Model](image)

**Figure 10** Geocentric axial dipole model (GAD). The geographic North Pole is N, total intensity of the magnetic field at sampling site is H at the latitude \( \lambda \), the corresponding inclination is I and the Earth’s mean radius is \( r_e \) (Butler 1992).
2.1 Sampling strategy

Samples examined in this dissertation have been taken progressively during the years 2004 and 2011. First field excursion was done in 2004 after the impact structure was found (Hietala and Moilanen 2004). The very first shatter cone samples were collected from the central uplift area. These were the initial hand samples, which assured our research team to establish a research project.

The second field excursion was organized in the course of the novel Keurusselkä research project funded by the Academy of Finland (Figure 11). During 2007 majority of samples described in this dissertation were collected around the impact area for further studies (Article I and II). In 2009 and 2010 additional sites were sampled inside and around the impact area as well as the three shallow (vertical depth of 40-80 m) drill cores in the vicinity of the central uplift (Article III) (Figure 12). The cores, originally drilled by SMOY, were stored in GTK’s drill core facilities.

Author’s contribution to the sampling is described more thoroughly in section 3.

Paleomagnetic samples are usually collected to provide a set of samplings of the geomagnetic field direction at the time of rock formation (Butler 1992). For this study several widely separated localities around the impact area were chosen to investigate how far the impact features are spread. Sites were sampled using a portable field drill or taking hand samples. Samples were oriented using orienting device to measure the declination (magnetic or sun compass) and dip (inclinometer) of the sample core (Figure 13). Hand samples were oriented with a compass (declination and dip) and later on drilled in the laboratory. Paleomagnetic measurements were made with respect to specimen coordinate axes (Figure 13).

Figure 11  Our research team at the Keurusselkä structure in 2007. From right: Selen Raiskila, Professor Lauri J. Pesonen, Dr. Martti Lehtinen and student Jonna Poikolainen (University of Oulu).
Figure 12  (A) Sampling of the drill core bits at the GTK drillcore facilities in 2009. (B) Drilling rock samples with portable rock drill at Keurusselkä. (C) Dr. Martti Lehtinen (left) and Professor Lauri J. Pesonen (right) taking hand samples at the Lake Keurusselkä 2007.

Figure 13  (Left) Example of in-situ geographic orientation of drilled samples. The z axis is the core axis and the x axis is the vertical plane. Sample orientation is determined by (1) azimuth of the horizontal projection of the x axis and (2) Hade of the z axis (angle from vertical of the z axis, which is 90° – plunge) (modified from Butler 1998), (Right) Selen Raiskila orienting drilled sample cores with an orienting sun compass device.

Multiple specimens were prepared from a sample cylinder as standard specimen cylinders with diameter of 2.5 cm and length of 2.2 cm. Altogether 501 specimens were measured using various methods.

2.1 Petrophysical measurements

The foundation of basic petrophysical measurement consist of measuring density (ρ), porosity (ϕ), magnetic susceptibility (κ), intensity of NRM and Koenigsberger ratio (Q value). These give information about the strength (e.g. porosity) of rock material, indirect indication of mineralogical properties (e.g. density, susceptibility) and changes along with
the direction and stability of the natural remanent magnetization (NRM). Dry bulk density of each specimen was measured using Archimedean method (Kivekäs 1993). Porosities of water saturated and oven-dried specimens were determined similarly. Magnetic susceptibility was measured (Articles II and III) with a kappabridge (Risto-5 instrument set) functioning with frequency of 1025 Hz, DC field intensity of 48 A/m and sensitivity of 10×10⁻⁶ SI for a volume of 50 cm³. NRM was measured with a fluxgate device (Risto-5 instrument set). Specimen volumes ranged between 11 and 96 cm³. The Koenigsberger ratio is defined from NRM and magnetic susceptibility using equation:

\[
Q = \frac{\mu_0 NRM}{\chi B}
\]

The Q value describes ratio of remanent magnetization vs. induced magnetization in the Earth’s magnetic field. Specimens natural remanent magnetization is multiplied with the magnetic permeability in vacuum \(\mu_0=4\pi\times10⁻⁷ \text{ H/m}\) and divided by the multiply of specimens volume susceptibility \(\chi\) and geomagnetic reference field B of 50 μT, which was used as a representative value of the IGRF in northern Europe.

### 2.2 Rock magnetic measurements

Rock magnetism measures the magnetic properties (NRM) of magnetic minerals and rocks, which have been affected by Earth’s magnetic field (Carmichael 1989). The basic types of magnetization are: diamagnetism, paramagnetism, ferro-/antiferromagnetism, ferrimagnetism and superparamagnetism. For rock magnetic purposes the most relevant magnetic materials are ferromagnetic (e.g. Fe, Ni), antiferromagnetic (hematite \(\alpha\text{Fe}_2\text{O}_3\)) and ferrimagnetic (e.g. magnetite \(\text{Fe}_3\text{O}_4\) and maghemite \(\gamma\text{Fe}_2\text{O}_3\)) (Figure 14).
2.2.1 Thermomagnetic measurements of magnetic susceptibility

Temperature dependence of magnetic susceptibility was measured using a KLY-3S kappabridge, with an operating frequency of 875 Hz and 300 A/m field, in conjunction with a CS-3/CS-L furnace (Articles II and III). In high temperature thermomagnetic measurements the rock powder specimen was heated up to 700 °C contained in an Argon gas environment and cooled back to room temperature. Susceptibility of a particular magnetic mineral was monitored during the heating process to determine the Curie temperature ($T_C$), above which the ferrimagnetic mineral becomes paramagnetic and it cannot anymore hold a spontaneous magnetization (Table 2). Ferrimagnetic minerals, such as magnetite and pyrrhotite have their specific Curie temperature, which allows us to make a distinction between them (Figure 15; $\kappa$-T curve). However, impurities (i.e. titanium content) can lower Curie temperatures. Antiferromagnetic minerals (or parasitic ferromagnetic minerals), such as hematite, exhibit a Néel temperature, which is analogous to Curie temperature. Low temperature thermomagnetic measurement was performed in a cryostat furnace (KLY-3S) using liquid nitrogen to cool down the specimen to -192 °C (80 K) temperature. Susceptibility of the specimen was monitored during the warming process to 0 °C. The transitions of particular magnetic minerals in low temperatures are called Verwey (magnetite) and Morin (hematite) temperatures (Dunlop and Özdemir, 1997).
Table 2  Relevant ferrimagnetic minerals in this study; compositions and theoretical Curie temperatures ($T_c$) (Dunlop and Özdemir, 1997).

<table>
<thead>
<tr>
<th>Ferromagnetic mineral</th>
<th>Composition</th>
<th>$T_c$ (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Magnetite</td>
<td>Fe$_3$O$_4$</td>
<td>580</td>
</tr>
<tr>
<td>Maghemite</td>
<td>$\gamma$Fe$_2$O$_3$</td>
<td>590-675</td>
</tr>
<tr>
<td>Hematite</td>
<td>$\alpha$Fe$_2$O$_3$</td>
<td>675</td>
</tr>
<tr>
<td>Titanomagnetite</td>
<td>Fe$_{x}$O$_3$</td>
<td>150-540</td>
</tr>
<tr>
<td>Pyrrhotite</td>
<td>Fe$_7$S$_8$</td>
<td>320</td>
</tr>
</tbody>
</table>

Figure 15  Temperature vs. susceptibility ($\kappa$-T) curves of Keurusselkä shatter cones indicating Curie temperatures of 580°C and 320°C (Article II).

2.2.2 Magnetic hysteresis properties

A ferromagnetic material becomes magnetized when it is placed in external magnetic field. Magnetic moments (atomic dipoles) attempt to align the field and become saturated when alignment is complete. When the external field is removed the saturated magnetic material will hold part of the alignment, which is called ‘remanent magnetization’. Remanent magnetization can be removed back to zero when the material is heated or exposed to a magnetic field in the opposite direction. This is the basis of a magnetic hysteresis loop, which shows the behaviour of all the contributing particles in the specimen cycled through a magnetic field. Hysteresis loops are used to define the overall domain state of the magnetic minerals in the measured rock specimens.

Measurements of magnetic hysteresis were performed with a Vibrating Sample Magnetometer (VSM) by Princeton Measurements Co (Articles II and III). The instrument vibrates the measured specimen in a sinusoidal motion within applied field.
Hysteresis loops for rock specimens are determined at room temperature using a maximum field of 1 T. The maximum magnetization, which a specimen can achieve in an applied field, is called saturation magnetization ($M_S$). This value depends linearly on concentration of the ferromagnetic mineral (Butler 1998). The magnetization remaining in the specimen after the field is removed is called remanent saturation magnetization ($M_{RS}$). The ratio of $M_{RS}/M_S$ is a measure of efficiency in acquiring remanent magnetization and it depends on the size and shape of the grain. The coercive force ($H_c$) is the intensity of the applied field required to reduce the magnetization to zero after the magnetization of the sample is driven to saturation ($H_{cr}$). Thus, coercivity is the resistance of ferromagnetic material against demagnetization. Both these parameters are dependent on the domain state. Uniaxial single domain (SD) grains are very small (<40 nm) particles, which have high coercivity. In very small particles the magnetic spins are essentially lined up and the particle is uniformly magnetized. SD grains are considered to be the best recorders of Earth’s past magnetic field directions as they are efficient carriers for remanent magnetization. Uniaxial SD grains ideal ratio of $M_{RS}/M_S$ is 0.5 (typically between 0.3 and 0.5) and in cubic ones 0.87. Ratio of $H_{CR}/H_C$ is between 1 and 1.2 (Merrill et al. 1996).

With increasing grain size, the growth of domain walls will lead to multidomain MD grains, which do not have uniform magnetizations (Tauxe, 2005). The coercivity of MD grains is modest and magnetization tends to decay with time. MD grains are therefore less effective to carry remanent magnetization. For MD grains the $M_{RS}/M_S$ ratio is 0.05 or lower. The ratio of $H_{CR}/H_C$ is variable, but typically higher (~5) compared to SD grains (Merrill et al. 1996). The boundary between SD and MD grains is not clear. Pseudo-single-domain (PSD) grains have transition size in which both SD and MD behaviour and characteristics are present (Carmichael 1989). PSD grains have lower $M_{RS}/M_S$ ratios than SD grains, but are still stable to hold remanence. Particular case is the nanoscale SD grains, which can be in superparamagnetic (SP) condition where the paramagnetic moment is extremely large compared to the single paramagnetic atom (Dunlop and Özdemir, 1997). In SP grains the thermal energy dominates and they can come into equilibrium with regardless external field within seconds. SP grains have very high susceptibility and even a small fraction in a specimen will dominate the induced magnetization. Their hysteresis behaviour resembles to MD particles.

In reality rock material often have mixtures of magnetic particles with different grain sizes and magnetic domain states. The most useful hysteresis ratios are $M_{RS}/M_S$ and $H_{CR}/H_C$, which are responsive to domain state, grain size and shape as well as the the source of magnetic anisotropy (Tauxe 2005). The Day plot (Day et al. 1977) is used to illustrate these ratios to determine domain states and grain sizes(Figure 16). The diagram is divided into regions (SD, PSD and MD) using theoretical boundaries, but it has limitations, as it is basically suitable only for titanomagnetites (Dunlop 2002) and not for hematite or iron sulphides such as pyrrhotite. One difficulty is also hysteresis ratios of mixture particles (SD+MD) or interacting SD grains, which will plot in to the PSD range. This is the reason why most of the ordinary rock specimen hysteresis ratios are in usually the PSD range (Tauxe 2005).
**2.2.3 First Order Reversal Curves**

More advanced method to represent hysteresis data is First Order Reversal Curves (FORCs) (Roberts et al. 2000). A specimen undergoes a measurement cycle, where the external field changes between a certain low field $H_a$ and a saturating field $H_b$. A series of curves between these fields produces FORCs, which can be transformed into a FORC diagram. A FORC diagram is a contour plot of FORCs, which are transformed and gridded to a useful form and smoothed (Figure 17). FORC diagram can thus provide more information about the magnetic particles, their compositions, size distributions and interactions between them. In articles II and III the FORC data was processed using a program provided by Winklhofer (Winklhofer and Zimanyi 2006).
Figure 17  FORC diagrams of different target rock types and shatter cones with smoothing factors (SF). Target metavolcanic rock shows typical SD type grain as other samples show either MD/PSD type grains with interactions of different grain sizes. Shatter coned granodiorite sample shows weak and noisy signal of magnetic grains (Article II).

2.2.4 Optical microscopy, scanning electron microscopy and microprobe analysis

Petrography is the systematic description and classification of rocks where distinctive optical properties of different rock forming minerals can be detected using polarizing microscope (Carmichael 1989). Optical microscopy is a critical method on impact studies when looking for high-pressure impact features in minerals, such as PDFs in quartz using U-stage and a stereographic projection template for indexing PDF sets, or traces of impact glass. The petrographic analyses were done in co-operation with Dr Ludovic Ferrière (University of Western Ontario) (Article I) and Dr. Ulla Preeden (University of Tartu) (Article III).

Quartz grains develop regular planar microfractures (PF) under shock compression. They are either parallel open fissures with spacing of >15-20 µm (Ferrière et al. 2009) or planar deformation features (PDFs). PFs are not considered as shock-diagnostic features of an impact without accompanying PDFs, which commonly composed narrow amorphous materials, less than 2 µm thick, occurring in straight parallel sets spaced 2-10 µm. The orientation of PFs and PDFs can be described by rational crystallographic planes of indices. These Miller-Bravais indices represent the inverse plane intercepts along a1,3 and c axes, which identify the 3D orientation of planes in a crystal. Typical PDFs are parallel to \{10\overline{1}3\}, \{10\overline{1}2\}, (0001), \{10\overline{1}1\}, \{11\overline{2}2\}, \{11\overline{2}\}, \{2\overline{1}\overline{3}\}, \{5\overline{1}\overline{1}\}, \{1\overline{0}\overline{1}0\} and \{1\overline{1}\overline{2}0\}. To
measure large number of PDF sets and index them efficiently a universal stage (U-stage) method with stereographic projection template (Figure 18) is used.

![Figure 18](image)

**Figure 18** (Left) A new stereographic projection template by Ferrière et al. (2009) (Article I). (Right) Thin section photomicrographs of quartz grains from Keurusselkä shatter cones containing decorated PDF sets (Article I).

Magnetic minerals obtained from rock magnetic measurements can be identified using scanning electron microscope analysis. In this study (Article II and III) a Jeol JSM-5900LV scanning electron microscope equipped with EDS (Energy Dispersive Detector) of the Geological Survey of Finland was used to directly observe the magnetic minerals within the surface polished specimen cylinders. In back-scattered electron images the magnetic minerals are reflected, which makes it easy to recognize them (Figure 19). SEM analysis shows detailed images of magnetic particles and e.g. alteration is distinguishable.

To determine elemental composition of breccia (Figure 20; article III) a microprobe analysis was done with EPMA (Electron Probe Micro-Analyzer) (EDS; JEOL Superprobe JXA-8600 with an improved and updated analysis software) in co-operation with the department of Geosciences and Geography.
Figure 19  SEM photograph of magnetic minerals (magnetite and FE-oxide; pyrrhotite) in Keurusselkä shatter cone samples (Article II).

Figure 20  (A) SEM image of biotite gneiss breccias, (B) optical microscope image (plane polarized light) of heavily brecciated rock with different sizes of lithic and mineral clasts in fine-grained to submicroscopic matrix, (C) SEM image of a pyrite grain and relatively fresh biotite inside optically nearly isotropic clay rich area (smectite, illite-group) possible altered impact glass and (D) SEM image of K-(Mg-Fe)-Al-silicate (chlorite) rich isotropic area. (Chl – chlorite, Mz – monazite, Py – pyrite and Qz – quartz ) (Article III).
2.3 Paleomagnetic measurements

Iron-bearing magnetic minerals (e.g. magnetite, hematite and pyrrhotite) in igneous and metamorphic rocks contribute significantly to the remanence of rocks and may record past directions of the Earth’s magnetic field. They can acquire a natural remanent magnetization (NRM) parallel to the ambient magnetic field (Merrill et al. 1996). This is the magnetization found in a rock sample in its in-situ condition. Primary NRM is gained during rock formation, but a secondary NRM can be acquired subsequently and it can alter or obscure the primary one (Butler 1998). NRM can be divided into three basic forms (1) thermoremanent magnetization (TRM), (2) chemical remanent magnetization (CRM) and (3) detrital remanent magnetization, which is related to depositional sedimentary rocks.

This remanent magnetization in rocks and minerals is dependent on magnetic material (type, size and shape of the grains) and its geological history. TRM is acquired when the rock cools below Curie temperature in external magnetic field. It is also the form of remanent magnetization acquired by most igneous rocks. Particularly old rocks of Precambrian age can acquire several types of secondary magnetizations, e.g. thermochemical (TCRM), chemical (CRM), viscous (VRM) or shock (SRM) remanent magnetizations due to geological processes or meteorite impact related processes. Rocks acquire VRM gradually during time at low temperature and small external field as Earth’s magnetic field. This kind of remanence is weak and unstable, but is present in most of the rocks (Carmichael 1989). The NRM represents the vector sum of all these remanent magnetizations, which are acquired over time. The main reason for paleomagnetic analysis is to isolate the various remanence components, identify their origin, reliability and acquisition time. The characteristic remanent magnetization (ChRM) is commonly used if the origin of magnetization is uncertain.

2.3.1 Demagnetization techniques

Paleomagnetic demagnetization techniques are either alternating field (AF) or thermal (TH) techniques, both of which can be used to separate the remanent magnetization components. The goal is to find a characteristic remanent magnetization component (ChRM), which is either of primary origin or secondary overprint.

AF treatment relates to the coercivities of magnetic minerals. Magnetic grains having lower coercivity than the AF peak value will be demagnetized. In the demagnetization process the applied AF peak is increased progressively after each step and switched between negative-positive sign in cycles. In this dissertation (Article II) a 2G DC-SQUID magnetometer was used to demagnetize specimens in AF field in order to isolate the magnetic components. The 2G DC-SQUID equipment allowed demagnetization steps up to 160 mT.

Magnetic grains can be demagnetized also by thermal treatment, which involves heating-cooling cycles in zero-field. The temperature is increased in steps until the maximum unblocking temperature of the magnetic mineral carrying the remanence is reached. The remaining remanent magnetization is measured after each temperature step.
Specimens are heated to a particular temperature, which ranges from 200 °C up to 700 °C in steps of 50 °C or 100 °C. The magnetization of grains having unblocking temperature lower than the heated temperature step will be randomized and demagnetized. The advantage of using thermal treatment is that we can also obtain the magnetic mineral carrying the remanence by their unblocking temperature, which relates to the Curie temperature. The main disadvantage is that the magnetic grains might be oxidized during the heating process, which changes their magnetic properties and thus disturb the magnetic analysis. To demagnetize specimens thermally the Magnetic Measurements MMTD-60 oven was used to heat the specimens in zero field and KLY-3S for monitoring the susceptibility (Article II).

2.3.2 Analysis of remanent magnetization components

Results of demagnetising the selected specimens were analysed using demagnetisation decay curves, stereographic projections and orthogonal plots (Zijderveld 1967; Leino 1991) (Article II) (Figure 21). Observing the NRM intensity decay behaviour is a simple way to approximately specify the magnetic minerals and their coercivities. Low coercivity minerals (e.g. MD magnetite) tend to decay to zero rapidly while high coercivity minerals (e.g. hematite, goethite) will not reach zero magnetization even with the applied AF maximum field of 160 mT. In these cases the thermal treatment is more favourable. Decay curve also shows the vector nature of NRM where component decreases in intensity, but not in direction. Stereographic projections of an equal area plot of the NRM declination and inclination of every demagnetization step shows the progress of magnetic directions and components. The orthogonal plot called Zijderveld diagram (Zijderveld 1967) is used to visualise a three dimensional demagnetisation data on a set of two projections of the vectors, where magnetization $J_1$ is the vector sum of $J_A$ and $J_B$. Only the vector $J_A$ is reduced in intensity in every demagnetization step ($J_1$, $J_2$, $J_3$...). In this projection the solid dots refer to N (S) component plotted versus E (W) in horizontal plane and open dots refer to N (S) plotted versus Down (positive inclination) or Up (negative inclination) in vertical plane (Dunlop 1979; Butler 1998; Tauxe 2005). Consequently, the diagram is a representation of different one or several paleomagnetic components in a specimen. Principal component analysis (Kirschvink 1980) determines paleomagnetic components as a best-fit line through scattered observation points, usually minimum of 3 steps with angular deviation (MAD) less than 6°. Mean remanence directions of best-fit lines are then calculated using Fisher (1953) statistics for a one specimen. Site mean directions are achieved when minimum of 3 specimens has the same components. Calculations are continued to component level, where all the different sites show the same direction. Using this analysis method, different paleomagnetic components can be determined from demagnetization data.
2.3.3. Paleomagnetic direction and poles

A mean direction of NRM for a site is a record of the past geomagnetic field direction during time when it was acquired. The determined pole, through an analysis of specimens (Fisher 1953) from a certain site, is subsequently called a virtual geomagnetic pole (VGP). VGP is the position of pole of a geocentric dipole that can account for the observed magnetic field direction at one location and one point of time (Butler 1998) (Figure 22). For each continent, the apparent polar wander path (APWP) has sequential positions of paleomagnetic poles. Due to the fact that Earth’s magnetic field changes its polarity it is not possible to know whether the studied site area was in the northern or in the southern hemisphere. In addition there are no yet adequate APWP for Precambrian rocks to allow a smooth transition plot to Phanerozoic APWP (Figure 23).
Figure 22  Magnetic pole position is determined from the magnetic field direction at the sampled site; S is the geographic site location (latitude $\lambda_s$, longitude $\Phi_s$); $I_m$, $D_m$ are the site mean magnetic field direction (inclination and declination); M is the geocentric dipole for the observed magnetic field direction; P ($\lambda_p$, $\Phi_p$) is the determined magnetic pole; $p$ is the magnetic colatitude and $\beta$ is the longitudinal difference between the site and magnetic pole (Butler 1992).

Figure 23  (A) Precambrian key poles with ages and paleomagnetic poles A and B obtained from the demagnetization data from Keurusselkä rocks. Most of the target granitoids showed primary type magnetization (pole A), while shatter cone samples showed magnetization (pole B) similar to 1122 Ma old Salla diabase (Salminen et al. 2009) implying a Mesoproterozoic age for the impact. (B) The South Pole position for B and other poles (C and D) obtained from Keurusselkä rocks plotted in Phanerozoic APWP (Article II).
2.4 Geophysical database

2.4.1 Aeromagnetic data

The whole Keuruu area is covered by an aeromagnetic survey done in 2007 by the Geological Survey of Finland as a part of their national airborne geophysical mapping program (magnetic – Cs-magnetometer, EM – AEM-05 with 4 frequencies and gamma radiation – spectrometer with 265 channels) (Hautaniemi et al. 2005). The flight altitude of 30 meters and line spacing 200 meters was used for high resolution mapping. Aeromagnetic data from the Keurusselkä impact area shows prominent magnetic anomalies in the center of the structure and around it, with high-amplitude short wavelength anomalies up to 500 nT (Figure 24). A high-amplitude circular anomaly with diameter of ~6 km coincides with the in-situ shatter cones. The anomaly is clearly distinguished from the overall regional field, which is mostly covered by anomaly lows of circular granitic intrusions. The other high-amplitude anomalies, in particularly slightly to South East from the central magnetic anomaly, appears to be a distinct magnetic ring structure. This ring could represent the remains of the crater rim. However, it is partly missing in the west, which could imply an uneven erosion or feature unrelated to the impact structure.

2.4.2 Gravity data

Finnish Geodetic Institute made a gravity survey in 2005, which densify a national gravity network of 5 kilometer average station distance. The survey was measured using Scintrex CG-5 gravity meter in order to find observable geophysical signatures related to the impact structure. For height and coordinate determination a Leica SR 530 GPS receiver with Geotrim’s Virtual GPS Reference System together with digital geoid model of the FGI was used. Measurement points were located to cover profiles across the impact area. A circular shape negative gravity anomaly ca. 6 mGal was found to be associated with the central part of the Keurusselkä impact structure (Ruotsalainen et al., 2006) (Article IV) (Figure 24). Other smaller circular negative Bouguer anomalies in the area were also found. These circular shaped anomalies are, however, mainly caused by younger granitic intrusions (1860 Ma), which are recognized also in the geological map by Nironen (2003) and in aeromagnetic map as low amplitude regions.
2.4.3. Seismic profile FIRE2 and seismic velocities

Seismic profile data interpretation was used in the modeling (Article IV). The Finnish Reflection Experiment (FIRE) was measured 2001-2005 (Kukkonen and Laitinen, 2006) and it provides crustal scale reflection seismic data of the deep structures and improved understanding of the crust in the Finnish part Fennoscandian Shield. The seismic data has broad frequency band and signal penetration of 20 seconds (two-way travel time). Vertical resolution in the crust is few tens of meters.

The FIRE2 profile crosses eastern margin of the Keurusselkä structure from NS direction. Nironen et al. (2006) presented a geological interpretation of the upper part of the crust in Keurusselkä area. Seismic velocities from surface rock samples were measured (Article IV) along the FIRE2 using method described by Lassila et al. (2010) and Elbra et al. (2010). Selected specimens were measured under laboratory and crustal (1 MPa) conditions to demonstrate the seismic velocity in the upper crust. Both surface (laboratory) and crust velocities showed decreased values compared to the typical seismic velocities of granitic rocks in the Baltic shield. A tomography analysis of the seismic data and the velocity-model done in the Institute of Seismology, did not show any clear boundaries or features of the deeper parts of the impact structure (Institute of Seismology, University of Helsinki, unpublished data and personal communication with M.Sc. Marianne Malm and Dr. Anna-Kaisa Korja). However, several areas with slower seismic velocities were recognized to a depth of 570 meters.
3 Review of papers and the author’s contribution

3.1 Article I


The most important part of this dissertation was to prove or disprove the impact origin for Keurusselkä structure, as the origin was initially established only on the basis of the occurrence of shatter cones, which are not approved as a confirmation of an impact. Although several field trips to the impact site, no clear in-situ impactites other than shatter cones were found. Finally a preliminary petrographic analysis of in-situ shatter cones showed promising shock features. A detailed microscopic investigation of thin sections, from two relevant outcrops, was done in order to investigate the amount of shock features in shatter cones. This study allowed to estimate the peak shock pressures recorded by the shatter cones.

Field trips showed that the shatter cones were abundant and well developed, principally on the shorelines of the lake Keurusselkä. Alas, thick vegetation, the lake and several cropped fields mostly obscured the rock formations. Found in-situ shatter cones revealed planar fractures (PFs) and planar deformation features (PDFs) in quartz and feldspar grains. Crystallographic orientations were measured using a universal stage (U-stage) microscope. Altogether 372 sets of PDFs in 276 quartz grains were measured. The shatter cones had experienced peak shock pressures between 2 GPa up to ~20 GPa (orientation \{10\overline{1}3\}). Although, the significant differences between recorded shock pressures from the same outcrop and between studied sites, made it difficult to determine the modification processes involved in the formation of the Keurusselkä impact structure. Most of the PDFs were also decorated either with two-phase, liquid and vapour, or monophase, liquid, inclusions. These inclusions indicate that originally amorphous shock features are altered by post-impact processes. The absence of toasted quartz indicates that the impact temperatures were not high or did not last long enough to induce toasting. Toasted quartz grains have orange-brown to greyish-reddish-brown appearance, which is possibly related to metamorphism and post-shock temperatures. However, the formation mechanism for toasting is yet unsolved.

My contribution consisted of preparing the shatter cone cylinder specimens and choosing the specimens for thin-section analysis. I contributed to the writing of the introduction of Keurusselkä impact structure and the geological setting in the manuscript. I also took the photos in Figure 2 and provided the basemap with sample sites for Figure 1.
3.2 Article II


After confirming the impact origin to the Keurusselkä impact structure in article I the next focus was to measure the petrophysical and rock magnetic properties of impactites (shatter cones) and estimate an age for the impact event using paleomagnetic dating.

Physical properties of rocks with increased magnetization and susceptibilities implied a central uplift area with diameter of ~5 kilometers (possibly up to 10 kilometers). The main magnetic mineral identified from shatter cones was magnetite with occasional traces of pyrrhotite. The amount of magnetic minerals was increased at the central uplift area indicating shock induced formation of magnetic material in the rocks.

Four paleomagnetic components were obtained to record the regional and global Mesoproterozoic and Paleoproterozoic geological events. First component A was interpreted to represent the primary Svecofennian magnetization of granitic rocks (1890-1840 Ma). Two components, C and E, were clarified to be remagnetizations related to much younger geological events unrelated to the impact. The component B led up to be the most interesting of the obtained four components. The pole B, yielding a paleomagnetic age of ~1120 Ma, is in close agreement with the 1122±7 Ma virtual geomagnetic pole of the Salla diabase dyke (Salminen et al. 2009). Furthermore, this age agrees with the $^{40}$Ar/$^{39}$Ar age, obtained from a pseudotachylitic breccia vein with PFs in quartz, which indicated 1140±6 Ma age for the breccia (Schmieder et al. 2009). The main question was if B component truly is related to the impact event as similar directions has been found also from other localities in Finland. Preeden et al. (2009) and Mertanen et al. (2008) proposed that this component would represent a Phanerozoic pole age (265-230 Ma), caused by oxidizing fluids in reactivated shear zones, rather than a Mesoproterozoic age. But, in Keurusselkä case this component was clearly restricted to the central uplift area and shatter cones with peak pressures up to 20 GPa. The fracture and shear zones in the Keurusselkä area did not show similar magnetization directions with component B. Furthermore, the component B and its paleomagnetic pole resembles more closely to well dated Salla diabase (1122±7 Ma) than the poles obtained by Preeden et al. (2009) and Mertanen et al. (2008).

The observations suggest that the impact event took place ~1120-1140 Ma and the impact induced the formation of new magnetic minerals, especially in the rocks at the central uplift area.

I contributed to the work by collecting majority of the rock samples during field trips 2007-2008 and 2010. Other samples from 2004 were collected by Dr. Tiiu Elbra, Dr. Tomas Kohout and Professor Lauri J. Pesonen and from 2009 by Dr. Jüri Plado and Professor Lauri J. Pesonen. I prepared most of the cylinder samples, made all the petrophysical and magnetic measurements, processed them and was responsible of the analysis and interpretation of the data. I was responsible of writing the manuscript.
3.3 Article III


In this article the main interest was to study samples from three drill cores, which were drilled in the vicinity of the Keurusselkä central uplift in 1968 during an ore exploration survey by the SMOY. A shallow drilling was organized to study the aeromagnetic anomalies of schistose belts around the Keurusselkä as pyrite and copper rich boulders was found in the area. The analyses of ore content, however, were not promising and the drill core samples were stored in the Geological Survey of Finland’s drill core facility.

In 2009 the author and research team colleagues visited the drill core facilities with intention to study and sample the drill core bits. The official report of the drill cores (Hugg, 1968, unpublished results in internal SMOY report of investigation) did not mention anything special other than sections of fractured rock. Despite the unpromising report we decided to visit the facilities and we found monomictic breccia, which resembled greatly with impact breccia.

A total of 99 samples were collected for further analysis. The lithology of drill cores are dominated by mica schist (metagraywackes), gneisses and felsic metavolcanic rock. The first drill core is 102 m deep, second 124.5 m and third one only 55 m. All drill cores have a dip between 45° to 50°. Unfortunately, the drill cores were un-oriented and paleomagnetic measurements were not possible. The core samples showed typical petrophysical values for schists and gneisses: density 2644-2752 kgm⁻³, susceptibility 160-761 × 10⁻⁶ SI, NRM 3-306 mAm⁻¹ and Q ratio 0.1-10. Rock magnetic measurements (κ-T curve and hysteresis) showed mostly paramagnetic behaviour with low quantities of fine-grained pyrrhotite and magnetite. These minerals are not, however, abundant enough to explain the magnetic anomalies, but rather represent the damaged crater basement surrounding the central uplift.

Petrographic analysis of four representative samples was done in order to investigate the breccia found from core V-002 at two different depths (68-70 m and 100-110 m). The core, which penetrates a 10-m-thick vein of monomictic breccia at the depth of 100-110 m (60-70 m of vertical true depth), has a section that resembles most to impact breccia (vertical true depth of 68 meters). This breccia has low density (2538 kgm⁻³), moderate susceptibility (270 × 10⁻⁶ SI) and very low NRM. The rock forming mineral phases include plagioclase, K-feldspars and quartz with minor dark minerals like biotite. Hornblende is predominantly replaced by fine-grained sub-microscopic mass of secondary Al-Fe-Mg rich silicate, chlorite. The presence of monazite-(Ce) that forms lamellae and irregular aggregates within chlorite grains or fills voids indicates secondary hydrothermal alteration or metamorphic origin. Diagnostic impact features, such as micas with kink banding were not found, which might be due to wide-spread chloritization. No PDFs in quartz, clear melt or diaplectic glass were found either. However, few small optically isotropic areas usually less than 1 mm were discovered. These areas consist of very fine grained clay minerals ~2 μm (chlorite, illite and smectite-group). Element mapping
showed that the isotropic area is depleted in Na and enriched in Fe, Mg and K relative to the surrounding breccia matrix. However, Al and Si did not differ from the matrix. Based on the homogenous appearance of isotropic areas and their very fine grained clay mineral composition, it seems that they have originated from glass. Non-impact origin is also plausible for the clay minerals. However, we based our views on results reported by Dressler and Sharpton (1997) from Canadian Slate Island impact structure. It has similar target rock type with large quantity and variety of impact breccia. The breccia had different phases, mainly altered and chloritized, showing hydrothermal alteration, which have replaced impact glass with chlorite or smectite. Post-impact hydrothermal alteration of impact rocks is a common phenomenon and this kind of activity is known from over 60 terrestrial impact structures (Naumov 2005).

As a result, our observations suggest that the lithology penetrated by the shallow drill cores represents the damaged and brecciated crater basement of deeply eroded Keurusselkä structure without diagnostic shock features. However, the optically isotropic areas appear to have originated as melt patches, which have solidified into glass and subsequently crystallized in hydrothermal alteration into clay minerals. These glass clasts could have originated as tectonic pseudotachylitic melt or impact related-melt. Impact induced melting could have occurred either in-situ or impact glass fragments were transported from upper parts along down faulted blocks (Kenkmann and Dalwigk 2000).

My contribution to the paper consists of collecting and measuring the samples with Dr. Tiiu Elbra. I prepared, measured and processed all the samples for petrophysics and rock magnetic measurements. I was also responsible of writing the major part of the manuscript.

Dr. Ulla Preeden prepared the thin sections, analysed them and provided text for the manuscript and the microscope images. M.Sc. Pasi Heikkilä supervised microprobe analysis. He also gave instructive comments to the manuscript.

3.4 Article IV


The fourth article integrates all the previous results, petrophysical properties of Keurusselkä rocks and the knowledge of impact related features, into a first geophysical model of the Keurusselkä impact structure. Geophysical modeling of magnetic and gravity data allows the delineation and estimation of crater dimensions, even when the studied structure is deeply eroded. In Keurusselkä case a special interest was in the central uplift region, which is characterized by well-defined in-situ shatter cones. The central uplift is clearly recognized from the aeromagnetic data as a prominent (~500nT) circular anomaly (6 km in diameter). The central uplift is not, however, recognized from topography. This
implies that the Keurusselkä central uplift is eroded to the level of crater basement. The damaged and fractured basement produces a gravity low of 6 mGal around the center parts of the Keurusselkä structure and the gravity low is more prominent in the eastern part of the structure. The asymmetry of the gravity anomaly suggests that the eastern part of the structure might have collapsed. In addition, other circular gravity anomalies in the reduced Bouguer anomaly of Keurusselkä were obtained. By using least-squares method three rings could be fitted around the central uplift with diameters of 24.4 km, 75.6 km and 101 km. The origin of these larger rings and their relation to the Keurusselkä structure is yet to be solved.

The article introduces the potential field maps and two modeled profiles based on petrophysical and rock magnetic properties of impactites (shatter cones) and target rock. First profile (A-A’) is modeled following the seismic profile line (FIRE2) across the eastern part of the Lake Keurusselkä. The second profile (B-B’) is modeled through the central uplift following the felsic metavolcanite intrusion. The central uplift has a striking high-amplitude circular magnetic anomaly with an expanding ring shape related to the shatter cones. The felsic metavolcanic rock with shatter cones has high susceptibility $26,615 \pm 26,535 \times 10^6$ SI displaying the effect of an impact into magnetic properties of the crystalline target rock. In contrast, felsic metavolcanic rocks without shatter cones have susceptibilities of $365 \pm 245 \times 10^6$ SI. Magnetic susceptibility of target granites is $1925 \pm 1895 \times 10^6$ SI.

Both models highlight a bowl shaped area of damaged and micro-fractured bedrock with lower density $(2532 \pm 187 \text{ kgm}^{-3})$, which produces the local gravity minimum. Seismic velocities in surface and crustal conditions were measured using instrumentation described by Lassila et al. 2010 and Elbra et al. 2011. Results highlight the fact that rocks in impact region have low seismic velocities indicating micro-fracturing. The measured average $V_{P0}$, $V_P$ and $V_S$ values for the granite and granodiorite are 4386, 5373 and 2688 ms$^{-1}$ and the $V_P/V_S$ ratios 1.83-2.17. In comparison, the average $V_P/V_S$ ratios, in the upper crust, range from 1.68 to 1.73 (Kuusisto et al. 2006). The impact damaged area reaches the depth of 1200 m beneath the present erosion level of the crater and has a diameter of $\sim$16 km. In comparison, the calculated theoretical parameters based on the diameter of central uplift (shatter cones) suggest diameter at least 19.4 km up to 27.3 km.

Geophysical features of Keurusselkä are not similar to the nearby located Lappajärvi structure (diameter 23 km, age $\sim$74 Ma) within the same target lithology. However, the resemblance to Siljan structure (diameter 52 km, age $\sim$377 Ma) is more evident. Siljan has a 10 km wide circular high-amplitude anomaly in the center of the structure similar to Keurusselkä. Therefore, the original rim-to-rim diameter for Keurusselkä is more likely larger than Lappajärvi and has been originally at least 24-27 km.

I contributed to the work by modelling and interpreting the data and the geophysical anomalies based on petrophysical properties of Keurusselkä rocks. I was responsible of writing the major part of the manuscript.
4 Discussion and updated conclusions

The increasing recognition of impact craters on Earth has extended our knowledge of our planet’s violent past and geological evolution. Understanding of impact cratering as a fundamental geological process on the bodies in the Solar system is essential, as it has contributed to surface rock formation on terrestrial planets, moons and asteroids.

This dissertation presents typical geophysical methods on studying impact craters. Keurusselkä structure offers a challenging view to study an eroded complex crater with limited impact features. The most important part of this dissertation was to find diagnostic proof of the origin of Keurusselkä structure and to integrate required methods needed to explain and interpret the observed impact related features.

The discovery of PDFs in quartz grains from three shatter cone sites made it clear that Keurusselkä structure is indeed originated from an impact event. Wider investigation of thin sections from all shatter cone sites would allow solving the overall occurrence of PDFs in Keurusselkä.

Even though several field trips were organized to the impact area, no clear in-situ impactites (melt, breccia) other than shatter cones and the one pseudotachylitic breccia vein were found. This disappointing aspect did not allow us to date the impact event using melt samples. However, the central uplift area where shatter cones were found showed high-amplitude magnetic anomalies indicating strong magnetizations. This fact made it possible to try paleomagnetic dating from shatter cones. The magnetization direction B (Dec=42.4°, Inc=64.1°, α95=8.4°) yielding a pole Plat= 61.0°, Plon= 129.1° and Α95=10.6° corresponds to 1122±7 Ma Salla diabase (Salminen et al. 2009) and provides important perspectives to different magnetization components of Fennoscandian Shield. The Keurusselkä paleomagnetic age (~1220 Ma) agrees also with the dating of a pseudotachylitic breccia vein by Schmieder et al. (2008), although the vein was never fully proved to be impact related.

Later on, a zircon dating was done in co-operation with Dr. Irmeli Mänttäri (GTK) from shatter cone samples, which showed highest impact pressures (20 GPa). Yet again, dating showed only ages of 1860±10 Ma (upper cutting age; crystallization age of the granitoids) and 310±28 Ma (lower cutting age) (personal communication with Dr. Irmeli Mänttäri). The lower cutting age corresponds with the paleomagnetic ages related to remagnetization of shear zones in southern Finland presented by Preeden et al. (2009) and Mertanen et al. (2008). This Phanerozoic age is typical in Finnish rocks and is related to the Caledonian uplift. The magnetization direction and the paleomagnetic South-pole of Keurusselkä B component indeed agrees with the Phaneroic APW path of Baltica, but its age is closer to 160 Ma than 310 Ma. A phanerozoic age would also be too young age for the Keurusselkä structure. Besides, no such B direction was found in fracture and shear zones in Keurusselkä area. These are the main reasons why the Keurusselkä component B was favoured to be Mesoproterozoic (1120-1140 Ma). So far, the paleomagnetic dating from shatter cones with diagnostic impact evidence is the best estimate for the age of Keurusselkä impact event.
The discovery of monomictic breccia from one of the drill cores located in the vicinity of the central uplift of Keurusselkä structure gave a chance to find possible impact melt. A petrographic analysis of thin sections revealed few areas (less than 1 mm), which appeared optically isotropic. Element mapping showed that the isotropic area is depleted in Na and enriched in Fe, Mg, and K relative to the surrounding breccia matrix. These isotropic areas consisted of very fine-grained clay minerals (chlorite, illite, smectite-group) likely to originate from glass. Other explanation could be that the isotropic areas represent only altered mineral grains without impact relation.

A geochemical analysis of the monomictic breccia was done in order to find meteoritic material within the samples (personal communication with Dr. Johanna Salminen). Alas, no such material was found. This brought us back to the question if isotropic areas are originated from impact glass or not. New drillings would be required to study thoroughly if there are other impact related breccia veins or if the highly magnetized central uplift contains some hidden features still to be discovered.

The final focus in this dissertation was to make a first geophysical model for Keurusselkä. Prominent magnetic anomalies distinguishable from the regional potential field are related to the central uplift and shatter cones. These high-amplitude (~500 nT) short wavelength anomalies occur at ~6 km wide area with in-situ shatter cones. The magnetic anomalies coincide also with the local gravity minimum, although the minimum is located slightly to the east of the shatter cones. Two profiles were modelled to interpret the anomalies and the rock sources producing them. The central magnetic anomaly appears to be originated by the effect of impact into the target rock material and represents the eroded central uplift. The magnetization could be due to uplifted rocks from deeper parts of the upper crust or the magnetic minerals are produced partly by the impact pressures and temperatures altering rock material. The modification of magnetic minerals occurs in high P-T conditions during the impact and post-impact processes where biotite decomposes into magnetite (Ugalde et al. 2005).

The gravity minimum related to the central uplift is caused by the damaged low density crater basement. The measured density values 2500-2550 kgm$^{-3}$ of the in-situ shatter cones and monomictic breccias from the drill cores were used in the model. The damaged and fractured region with low density has a diameter of ~16 kilometres and exceeds to the depth of 1200 metres. The crater dimensions were estimated based on the modelling parameters and theoretical equations. According to the model the minimum diameter of the original Keurusselkä crater has been at least 16 km, but most possibly between 24 and 27 km. Besides the main gravity minimum also other gravity anomalies were found around the central uplift. By fitting circles to the reduced Bouguer anomaly data using least-squares method it was observed that the circles are concentric and coinciding with the shatter cone. The three circles have diameters of 24.4 km, 75.6 km and 101 km. If these gravity anomalies are related to the Keurusselkä impact structure they will give interesting perspective to wider disturbances in the target rocks of Keurusselkä region. More information about these features could be gained by extending the geological field mapping in the Keurusselkä region.
The source body dimensions in the geophysical model of Keurusselkä structure was defined using a graphical interpretation of FIRE2 seismic profile line data. Had the numerical seismic profile data being available the model would have been much improved. A further improvement would also be the compilation of structural geological investigation. Drillings in the central uplift area would provide more information on the magnetic source and possible reveal new breccia veins.

Keurusselkä is an old and deeply eroded impact structure and the remains of the impact lithology is likely absent from the nearby glacial till formations deposited by the Fennoscandian ice ages. One possibility to find traces of distal ejecta is Mesoproterozoic sedimentary sections in Baltica or nearby located Rodinia continent, such as the 1170 Ma Stoer Group sediments in Scotland (Amor et al. 2008; Parnell et al. 2011); although, the Stoer group sediments are somewhat older (1170 Ma) compared to the age estimates for Keurusselkä structure (~1120-1140 Ma). We note that there is not yet found an impact crater, other than Keurusselkä, within a suitable age and size to match with the Stoer group ejecta. Alternative known craters within Mesoproterozoic age are Finnish structures Suvasvesi N (age <1000 Ma, diameter 4 km), Lumparn (age ~1000 Ma, diameter ~9 km) and Iso-Naakkima (age >1000 Ma, diameter ~3 km). Also Santa Fe structure in New Mexico has Mesoproterozoic age (<1200 Ma, diameter ~6-13 km), although it has originally situated too far to be linked with the Stoer group in NENA (Northern Europe – North America; Gower et al. 1990) configuration, where west Baltica (East European craton) and east Laurentia (North America and Greenland) coexisted from 1800 to ~1200 Ma.

The scientific work presented in this dissertation provides impact evidence for Keurusselkä structure. The diagnostic planar deformation features in shatter cones state that Keurusselkä originates from an impact. However, it is deeply eroded and post-impact hydrothermal activity has altered possible traces of impact melt into clay minerals. Geophysical signatures imply that the rim-to-rim diameter of Keurusselkä structure has been 24-27 km. But, the lack of melt breccia or any other impact features other than non-diagnostic ones, thus, prohibit the ground truth interpretation of the original crater size and the projectile type. The obtained results highlight the ultimate challenge to study ancient impact structures and their features in respect of impact cratering in the Earth’s geological history.
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