The Earth’s Magnetic Field

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THE EARTH’S MAGNETIC FIELD

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Abstract

The Earth’s magnetic field is essential for life on Earth, as we know it, to exist. It forms a magnetic shield around the planet, protecting it from high energy particles and radiation from the Sun, which can cause damage to life, power systems, orbiting satellites, astronauts and spacecrafts. This report contains a general overview of the Earth’s magnetic field. The different sources that contribute to the total magnetic field are presented and the diverse variations in the field are described. Finally, measurements of the Earth’s magnetic field are introduced and the applied instruments and procedures described, with an emphasis on Leirvogur magnetic observatory.
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1 Introduction

The Earth’s magnetic field has existed since the planet’s early days - long before life on Earth began. Without the magnetic field, life on Earth, as we know it, could not even exist. That is because the Earth’s magnetic field protects the near-Earth environment from dangerous radiation and high energy plasma from the Sun by partially blocking it out. However, there is always some part of the plasma and radiation that is able to penetrate through the magnetic field barrier, and this often leads to magnetic disturbances around the Earth. Before humans started seeking out into space, with orbiting satellites and space voyages, and before electricity became a basic need of a daily life, the magnetic disturbances did not have a large effect on life on Earth. In modern times, the need to understand the behavior of the Earth’s magnetic field, its origin and variations, is becoming ever more important. With intense research, focused on both the surface of Earth, and the Sun-Earth environment, the goal is to be able to predict the behavior of the magnetic field, just as meteorologists predict the weather.

The intention of this report is to give a general overview of the Earth’s magnetic field by specifying its main properties, the characteristics of the magnetic field variations and their causes. Included is also an introduction to the measurements of the Earth’s magnetic field, which are mainly carried out at magnetic observatories, and the instruments used for measuring it. This report is hopefully suitable, not only, for someone stepping their first steps in the field of geomagnetism, but also for someone working within the field, wanting to refresh their memory on the basics.

To begin with, the geomagnetic field is introduced by describing the different sources of the observed magnetic field. One chapter is dedicated to variations in the magnetic field and their origins. From there, the emphasis moves to the subject of measuring the magnetic field, describing the involved instruments and standard geomagnetic observatories, where the measurements are carried out.
2 Properties and Origin

The Earth has a magnetic field which is mainly produced within its interior and forms a protecting shield around the planet, called the magnetosphere. The magnetosphere protects Earth’s inhabitants from the highly energetic and dangerous particles from the sun and it prevents the Earth’s atmosphere from being blown away by the solar wind.\textsuperscript{1} The geomagnetic field is thought to have existed much further back in time than life on Earth. According to magnetized rocks as old as 4 billion years, found in Greenland and Australia, the magnetic field already existed in the beginning of Earth’s history [8].

2.1 Contributions to the Total Magnetic Field

The observed magnetic field on Earth is called the geomagnetic field. It is a superposition of magnetic fields generated by different sources. The field generated by a magnetic dynamo in the Earth’s liquid core is called the main field and is by far the most dominant one. The main field and the crustal field, which is generated by magnetized rocks on the Earth’s crust, are categorized as the internal field and are rather stable in time compared to the external field. The external field is composed of the field produced by the electric currents in the ionosphere and magnetosphere. There is also a magnetic field produced by the induced currents in the crust, mantle and oceans which adds to the total geomagnetic field. To summarize, the geomagnetic field components which make up the total geomagnetic field are: The main field, the external field and the induced field.

A detailed global coverage of the magnetic field, for a long period of time, is needed for it to be possible to separate and identify the different contributions of each part of the field to the total geomagnetic field. As mentioned above, the main field is produced in the interior of the Earth by a self sustaining dynamo which is explained by dynamo theory [8, 15]. Without going into details about the origin of the main field, the following basic description is made. The Earth’s core is believed to be mainly composed of iron (and to a smaller proportion, nickel) and consist of a solid

\textsuperscript{1}It is thought that Mars had a denser atmosphere when it had a stronger magnetic field and that it got blown away by the solar wind when the magnetic field became weak.
inner core and liquid outer core. The magnetic field is produced by circulating electric currents in the the highly conductive outer core. The complex motion of the currents (or the fluid conducting material), is driven by convection and the rotation of the Earth. The energy needed to sustain this dynamo is believed to be produced by solidification; when the heaviest elements in the fluid core freeze onto the solid core and lighter elements are released, causing a convection [8]. The non-homogeneity of the electric currents causes regional magnetic anomalies on the Earth’s surface and variations in the main field are connected to variations in the fluid velocity [8]. Before the theory of a non-trivial dynamo in Earth’s interior, there were suggestions that the Earth’s magnetic field was produced by a strong permanent magnet located in its center. However, ferromagnetic material, like iron in the Earth’s core, cannot be magnetized above their Curie temperature, the temperature at which a magnet loses its magnetism. The temperature in the Earth’s core is much higher than the Curie temperature of iron. In fact, at the depth of a few tens of km the temperature is higher than the Curie temperature of almost all known ferromagnetic materials [5]. According to this there must be electric currents in Earth’s interior producing the field. At present time, the dynamo theory has come quite far. Computer models of the dynamo in the Earth’s core are being developed, and are able to simulate phenomena like the reversal of the poles, where north and south magnetic poles are interchanged, and the slow changes in the main field.

2.2 The Elements of the Earth’s Magnetic Field

The geomagnetic field has a direction and magnitude at every point in space and thus is a vector field. Three parameters are needed to describe the magnetic field vector, \( \mathbf{F} \), at a point, \( P \), on the Earth’s surface. The point, \( P \), is assumed to be situated at the center of a Cartesian coordinate system with the \( x \)-axis directed towards the geographic north, the \( y \)-axis directed to the east and the \( z \)-axis vertical and positive downwards into the Earth. The north, east and downward components of the field vector are denoted \( X \), \( Y \) and \( Z \). Figure 2.1 describes the geometry. A common way to describe the magnetic field is to specify the total intensity of the field \( F \), the inclination, \( I \), and the declination, \( D \). The inclination is the angle from the horizontal plane to the vector and the declination is the angle from the \( x \)-axis to the component \( H \), which is the horizontal component of \( \mathbf{F} \). Another way of describing the magnetic field, is by specifying \( H \), the vertical component \( Z \), and the declination, \( D \). Figure 2.1 shows the orientation of each component and the

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2 The relatively fast rotation of the Earth is believed to play a key role in the Earth’s magnetic field and the dynamo in its core because the neighboring planet Venus has a similar core structure but a slower rotation and thus doesn’t have a magnetic field [6].

3 For comparison, the radius of Earth is approximately 6400 km.
The following equations describe their internal relations:

\[ F = \sqrt{X^2 + Y^2 + Z^2}, \quad H = \sqrt{X^2 + Y^2}, \quad I = \arctan \frac{Z}{H}, \quad D = \arctan \frac{Y}{X}. \]

The components in Cartesian coordinates are:

\[ X = H \cos D, \quad Y = H \sin D, \quad Z = F \sin I. \]

The geomagnetic field components X, Y, Z and H are given in nano teslas (nT)

![Diagram of the Earth's magnetic field](image)

**Figure 2.1:** The elements of the Earth’s magnetic field. F is the total intensity of the field, H is the horizontal component of F, D is the declination and I the inclination [8].

while the I and D are presented in degrees, arcminutes and arcseconds which are all units of angular measurements, an arcminute being 1/60 of one degree and an arcsecond is 1/60 of one arcminute. A common way for magnetic observatories to present their daily measurements is through a magnetogram. A magnetogram is a plot of the magnetic field versus time. Most often, three field components are specified and the time interval used is 24 hours.

### 2.3 The Geometry of the Earth’s Magnetic Field

The absolute strength of the magnetic field on Earth’s surface varies from 23,000 nT to 62,000 nT [13]. It is strongest around the poles and weakest around the equator but between the two, the variations in the strength is not linear. The field both varies with space and time, and in this section the focus is on the spatial description and representation of the Earth’s magnetic field. Short and long term time variations are discussed in the next section. In the polar regions the magnetic field consists more
or less of a vertical component and at the equator the field lines lie in the horizontal plane. The location at which the field is purely vertical are called south and north magnetic poles and they are not the same as the geographical poles marked by the Earth’s rotation axis.\footnote{Another definition is the geomagnetic poles which are the locations where the magnetic poles would be if the Earth’s magnetic field was purely dipolar and produced solely by a bar magnet.} As implied, the Earth’s magnetic field is like the field of a magnetic dipole, but only to a first approximation.\footnote{More than 90% of the observed field can be approximated by the simple dipole model [5].} It is as if there were a giant bar magnet located at the center of the planet, tilted $11.5^\circ$ degrees from the Earth’s rotational axis. The hypothetical bar magnet should be oriented as shown in Figure 2.2 since a compass points north and opposite poles attract. The locations where the magnetic poles would be if the Earth’s magnetic field was purely dipolar and produced solely by a bar magnetic are called the geomagnetic poles. The location of the geomagnetic poles can be seen in figure 3.1. The simple and time independent bar magnet model does not take into account the many complex properties of the magnetic field, for instance magnetic anomalies, the field reversal, movement of the geomagnetic poles and other variations in the field, which will be introduced later. An example of the deviations from a dipole field is the fact that the magnetic poles are not located exactly opposite each other.

\begin{figure}[h]
\centering
\includegraphics[width=0.5\textwidth]{figure2.png}
\caption{To a first approximation, the Earth’s magnetic field is like the field of a magnetic dipole [18].}
\end{figure}

\section{2.3.1 Spherical Harmonic Description}

An analytical expression of the global magnetic field is achieved by the use of spherical harmonics which come from the solution of Laplace’s equation. Therefore it is relevant to start by deriving Laplace’s equation. The two Maxwell’s equations related to the magnet field are

\begin{equation}
\nabla \cdot \mathbf{B} = 0 ,
\end{equation}

and

\begin{equation}
\nabla \times \mathbf{H} = \mu (\mathbf{J} + \frac{\partial \mathbf{D}}{\partial t}) ,
\end{equation}
where $\mathbf{H}$ is the magnetic field, $\mathbf{B}$ is the magnetic induction, $\mathbf{J}$ is the electric current density and $\partial \mathbf{D}/\partial t$ is the electric displacement current density. Given the approximation that there are no electric currents on the Earth’s surface, equation 2.2 becomes $\nabla \times \mathbf{H} = 0$. As a consequence, the magnetic field, $\mathbf{H}$, is a conservative vector field and can be connected to a magnetic scalar potential, $V$, by

$$\mathbf{H} = -\nabla V. \quad (2.3)$$

At the Earth’s surface the magnetic induction is $\mathbf{B} = \mu_0 \mathbf{H}$, where $\mu_0$ is the permeability of free space. Now equation 2.3 and 2.1 together yield Laplace’s equation $\nabla^2 V = 0$, which in spherical coordinates is given by

$$\nabla^2 V = \frac{1}{r^2} \frac{\partial}{\partial r} (r^2 \frac{\partial V}{\partial r}) + \frac{1}{r^2 \sin \theta} \frac{\partial}{\partial \theta} (\sin \theta \frac{\partial V}{\partial \theta}) + \frac{1}{r^2 \sin^2 \theta} \frac{\partial^2 V}{\partial \lambda^2} = 0. \quad (2.4)$$

The components of the spherical coordinates are shown in figure 2.3. Here, $r$ is the radial distance of a point $P$, $\theta$ is the angle from the $z$-axis to the vector and $\lambda$ is the geographic longitude. Every function $V(r, \theta, \lambda)$ that satisfies equation 2.4 is a harmonic function. For this to apply to the magnetic potential on the surface of Earth, it has to be assumed that the Earth is a perfect sphere. It is worth emphasizing that a solution to Laplace’s equation leads to a mathematical description of the global magnetic field. Before presenting the solution to the above differential equation, the different components of the total field should be considered again. The total magnetic field at the Earth’s surface is the sum of all contributions from the internal and external sources. Thus, the total magnetic potential is also the sum of all internal and external potentials that contribute to the field. It turns out to be reasonable to approximate the total potential at the surface with internal contributions only, and assume that the external potential is negligible in comparison. Equation 2.4 is solved by separation of variables, and if only internal factors are taken into account,

\[\text{Examples of electric currents on the Earth’s surface are currents produced by a lightning.}\]
2 Properties and Origin

the solution is given by

\[ V = a \sum_{l=1}^{\infty} \sum_{m=0}^{l} \left( \frac{a}{r} \right)^{l+1} P_{l}^{m}(\cos \theta) (g_{l}^{m} \cos m\lambda + h_{l}^{m} \sin m\lambda) , \]  

(2.5)

where \( a \) is the radius of the Earth \( g_{l}^{m} \) and \( h_{l}^{m} \) are expansion coefficients called Gauss coefficients given in units of nT and \( P_{l}^{m}(\cos \theta) \) are the Schmidt functions which are often used in geomagnetism. Schmidt functions are equivalent to the better known Legendre functions, \( P_{l}^{m} \), but they have a different normalization factor.\(^7\) The integer \( l \) denotes the degree of \( P \), and \( m \) denotes the order. The intensity of the magnetic field of internal origin must decrease as \( r \) increases and go to zero at infinity. That is why the intensity depends on \( r \) as \((a/r)^{l+1}\). Now, the magnetic field is of more interest than the potential, since it can be measured, whereas the potential cannot. The \( X \), \( Y \) and \( Z \) elements of the magnetic field at the Earth’s surface are obtained by taking the gradient of the potential:

\[ X = \frac{\partial V}{r \partial \theta} \quad Y = \frac{-\partial V}{r \sin \theta \partial \lambda} \quad Z = \frac{\partial V}{\partial r} . \]

The two angular functions in equation 2.5, the Schmidt function and the Fourier series of \( \lambda \), represent the latitudinal and longitudinal magnetic field variations respectively. The product of the two makes up the so called spherical harmonics, \( Y_{l}^{m}(\theta, \lambda) \).

The Gauss coefficients are computed by using data about the magnetic field from satellite and ground based measurements. Individual terms in the equation can be interpreted in terms of hypothetical sources at the Earth’s center \([15]\). When only taking the terms associated with \( l = 1 \) in equation 2.5, the magnetic potential represents a geocentric magnetic dipole. The terms with \( l = 2 \) represent a geocentric quadrupole, terms with \( l = 3 \) an octupole and so on. In general, all \( l = 1 \) terms are said to make up the dipole field of Earth’s magnetic field, while the remaining terms represent the non-dipole field. If all the terms associated with \( l = 1 \) and \( l = 2 \) are used, the result is a potential of a tilted dipole (also called eccentric dipole), which gives a better description of the geomagnetic potential than if only the first degree terms were used. As expected, a higher \( l \), gives a better fit to the global representation of the Earth’s magnetic field. The potential sum converges and the higher \( l \) is, the less the corresponding terms contribute to the total field, and at around \( l = 14 \) the potential hardly changes at all with the addition of more terms \([8]\). It is possible to make the potential time dependent by adding time derivatives of

\(^7\)The Legendre polynomials for \( m \geq 0 \) can be expressed in the form

\[ P_{l}^{m} = \frac{(-1)^{m}}{2l!} (1 - x^{2})^{m/2} \frac{d^{l+m}}{dx^{l+m}} (x^{2} - 1)^{l} \]
the Gauss coefficients. In this way, slow changes can be represented by the magnetic potential.

This method of spherical harmonic analysis, derived from Laplace’s equation, represents the geomagnetic field well, so the approximation of no electric currents on the surface turns out to be a good one. Electric currents crossing the Earth’s surface are small enough to not affect the coefficients of the spherical harmonic representation of the field [15]. However this approximation is of course not valid beneath the surface of Earth. Finally, the global representation of the total intensity, $I$, and the declination, $D$, of the magnetic field in the year 2010, are shown in Figure 2.4.
2 Properties and Origin

Figure 2.4: The total intensity, $I$, and the declination, $D$, of the main field in 2010 [17].
2.4 The Magnetosphere

The Earth and its magnetic field is highly affected by its neighboring star, the Sun. The Sun is continuously emitting ionized gas (consisting mainly of electrons, ionized hydrogen and helium) with an average density of around $7 \text{ ions cm}^{-3}$. This constant stream of plasma, moving at speeds of 250-2000 km/s [10], is called the solar wind. It’s always directed away from the Sun but it can change speeds and density with varying solar activity. The solar wind carries with it the solar magnetic field which fills up the interplanetary space and is called the interplanetary magnetic field (IMF). The low density plasma from the Sun and the IMF shape the Earth’s magnetic field, which is actually quite different from the dipolar field mentioned earlier. The area in which the Earth’s magnetic field dominates and controls the motion of particles is defined as the magnetosphere. Figure 2.5 shows the different characteristic regions within the magnetosphere and figure 2.6 shows how the solar wind shapes the magnetosphere. On the sunward side of the Earth, the magnetosphere is compressed due to the solar wind, and has a paraboloidal surface called the magnetopause. The magnetopause is the outer limit of the magnetosphere because there, at around $10 \text{ R}_e$ (Earth radii) [10], the solar particle pressure and the Earth’s magnetic pressure is balanced [8]. When the supersonic solar wind, moving with an average velocity of around 400 km s$^{-1}$ with respect to the Earth [15], collides with the magnetosphere, a bow shock is formed.

Figure 2.5: The Earth’s magnetosphere and its characteristic regions [8].
This happens at a distance of around $13 \text{ R}_e$. On the Earth’s dark side, the magnetosphere stretches far out into space, more than $100 \text{ R}_e$, forming the *magnetotail*. Within the tail there is a plasma sheet, often called the *neutral sheet* because of the weak magnetic field in the area. In this plasma sheet there are electric currents which affect the geometry of the magnetotail.

The magnetosphere shields the Earth from the incident solar wind but there is always some amount of electrons and ions, that are able to penetrate through the "shield" and after that become a part of the magnetosphere. The trajectory of charged particles in the magnetosphere is governed by the Lorentz force\(^8\) which makes the particles spiral around the magnetic field lines, moving back and forth between the poles. When the charged particles approach the polar regions, where the magnetic field lines are more dense and the field is stronger, they slow down as being repelled, and bounce back to the other hemisphere. When they reach the other hemisphere the process is repeated, and so on. These particles, surrounding the Earth and traveling back and forth between the hemispheres, form the Earth’s *radiation belts*. There are two regions in particular, where particles get trapped, and they are called the inner and outer *Van Allen radiation belts*. Some particles are energetic enough to be able to penetrate down to the Earth’s upper atmosphere. There, they collide with atoms and lose their energy. These collisions lead to the *auroras* because atoms in the Earth’s atmosphere get excited by the collision and emit

\(^8\)The Lorentz force $\mathbf{F}$ on a charged particle with velocity $v$ in an external magnetic field $\mathbf{B}$ is $\mathbf{F} = qv \times \mathbf{B}$. 
2.4 The Magnetosphere

electromagnetic radiation in the visible range when they de-excite. This radiation is the auroras.

The shape of the magnetosphere is dependent on the solar activity. When solar activity is high, more intense solar winds makes the solar side of the magnetosphere compress towards Earth and the tail stretch back. This also means that larger amounts of particles are able to get within the magnetosphere, causing complicated current systems and short term time disturbances in the magnetic field which are discussed in chapter 3.2.

Another way to describe the magnetosphere is in terms of its constituent electric currents shown in figure 2.7. The magnetospheric current systems has general features which can be separated into three smaller systems; The magnetopause current, the ring current and the magnetotail current system. The magnetospheric current system is quite complicated and will only be briefly described. The magnetopause current is defined by a drift of charged particles which make up an eastward current near the equatorial plane [10]. The magnetotail currents consist of a westward equatorial current sheet [10]. The ring current is located between two and seven Earth radii. It is caused by trapped particles in the radiation belts because ions drift westward and electrons eastwards. This longitudinal drift gives rise to a net westward equatorial current sheet [5].

![Electric currents in the magnetosphere represented by red lines and the blue lines represent magnetic fields](image)

Figure 2.7: Electric currents in the magnetosphere represented by red lines and the blue lines represent magnetic fields [10].

The lower boundary of the magnetosphere marks the upper part of the ionosphere. The ionosphere is localized at an height between around 50-600 km and it is the electrically conducting part of the upper atmosphere. The solar ultraviolet radiation
is what causes its ionization and thus, its properties vary with latitude, season, solar time, etc. The ionosphere is divided into different regions, depending on the height and ionization, they are called the $D$, $E$, and $F$ regions [10]. The ionization is greatest at about 300 km altitude, in the F region. Large electric currents can be produced in the ionosphere, especially in the F region where charged particles are subjected to electromagnetic forces [15]. Particles in the ionosphere are also affected by tidal forces and thermal effects which can cause further electric currents.
3 Time Variations

Geomagnetic field measurements and other geological studies show that the Earth’s magnetic field changes significantly with time. The time variations can be from the scale of seconds up to millions of years, they can be periodic or completely random, they can include changes in declination or inclination and the field’s total strength can vary from a few nT up to thousands of nT. The time variations are divided into two main groups: Long and short term variations. What distinguishes them, apart from the duration, is that long term variations come from the dynamics of the Earth’s interior, whereas short term variations have an external origin. The long term variations are on a scale of 5 years or more and are called secular variations [8]. They appear as main field variations. The short term variations can be on the scale of seconds or more but their duration hardly exceeds a year. They often cause intense variations in the field and are mainly produced by currents in the magnetosphere and ionosphere, but also by induced currents in the Earth’s crust and oceans. The rapidity and intensity of these variations makes them very obvious in magnetic observatory data. At times, when the intensity and direction of the geomagnetic field vector is constantly changing, the field is said to be magnetically active. Additionally, periodic variations due to the rotation and/or orbital motion of the Earth, Sun and Moon are included in the group of short term variations.

To describe the state of the magnetic field, a variable is needed and the most commonly used is the $K$-index. The $K$-index indicates the localized intensity of the magnetic activity at an observatory, where the indices are scaled from the observatory recording and presented by a one-digit number in the range 0 to 9 [5]. A value of zero means no magnetic activity and nine means significant variations in the field. The $K$-index is evaluated for each 3-hour interval and is measured from the disturbances in the horizontal components, $H$ and $D$, where the larger of the scaled amplitudes determines the index [5]. The amplitude is obtained by calculating the difference between the minimum and the um values in the relevant 3-hour interval. Observatories at various latitudes have different ranges but all of them are multiples of the standard scale in table 1.\footnote{The $K$-index range is dependent on the observatory because of an attempt to normalize the frequency of occurrence of the different sizes of disturbances [5].}
Table 3.1: Standard scale for conversion between maximum intensity in the horizontal components, \( H \) and \( D \), and the \( K \)-indices \([5]\).

<table>
<thead>
<tr>
<th>K-value</th>
<th>0</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
<th>8</th>
<th>9</th>
</tr>
</thead>
<tbody>
<tr>
<td>Range [nT]</td>
<td>0</td>
<td>5</td>
<td>10</td>
<td>20</td>
<td>40</td>
<td>70</td>
<td>120</td>
<td>200</td>
<td>300</td>
<td>500&lt;</td>
</tr>
</tbody>
</table>

### 3.1 Long Term Variations

The existence of the secular variation was discovered in 1634, when geomagnetic measurements showed that the declination of the Earth’s magnetic field was not only a function of position, but also of time \([5]\). Today, the intensity and the inclination of the main field are known to vary with time as well. These changes are rather small but still well recognizable when looking at geomagnetic data ranging over several years. The total intensity of the Earth’s magnetic field can vary from a few nT per year to tens of nT per year and the declination and inclination can vary from a few minutes per year to several minutes per year. The secular variation and the main field both originate from the same source within the Earth’s interior. Thus, the secular variation gives important information about the dynamics of the conductive fluid in the Earth’s core and the origin of the geomagnetic field itself. One of the main applications of research on the secular variation, is to improve and construct new theoretical models that simulate the current system which generates the geomagnetic field. A constant measure of the secular variation is also important for updating magnetic charts. A crucial factor regarding the study of the secular variation is having magnetometers with an accuracy smaller or equal to the field variations (\( \approx 1 \) nT) and obtaining continuous geomagnetic field data over a very long time interval. The time interval of direct measurement data is limited by the existence of the magnetometers. Luckily, however, there is another way of getting information about the Earth’s magnetic history than direct measurements: By examining magnetized rocks.

*Paleomagnetism*, the study of the remanent magnetization in rocks, yields the direction and intensity of the Earth’s magnetism from its early days, to the present. Paleomagnetism’s most significant contribution to the field of geomagnetism, is the discovery of the so called field reversal, which is the exchange of position between the north and south magnetic poles. The field reversal means change in the declination by 180° and reversal of the sign of the inclination. Analysis of volcanic and sedimentary rocks have led to the conclusion that the last reversal took place around 780,000 years ago \([8]\). The transition itself is believed to take 5000 – 10,000 years and the poles remain constant for about 100,000 to one million years, but the rate at which the field reverses seems to be quite random \([8]\).

The location of the magnetic poles also changes with time. Figure 3.1, from the
3.1 Long Term Variations

Figure 3.1: The drift of the north (left) and south (right) magnetic and geomagnetic poles [11].

World Data Center for Geomagnetism in Kyoto, shows the drift in both the north and south magnetic poles and their predicted position in 2015, as well as the location of the geomagnetic poles. In the 20th century, the north magnetic pole moved with an average speed of 10 km per year and has lately accelerated to 40 km per year [19]. Currently the north magnetic pole is located north of Canada. It is heading towards Russia, where it will most likely be located in a few decades [19]. Since the position of the poles are important, in navigation for instance, their accurate location must be published every five years or so.

The secular variation of the Earth’s magnetic field is often divided into the time variations of the Earth’s dipole field and its non-dipole field, mentioned in section 2.3.1. Magnetic observatory data give evidence that the energy in the dipole field is decreasing while the energy in the non-dipole field is increasing. Still, the energy in the total field shows an overall decrease [8]. Contour maps of the declination worldwide, similar to the one in figure 2.4, from the years 1600 to 2000, show clear evidence of the so called westward drift. The westward drift is a clear drift of almost all declination contour lines towards the west and it is due to variations in the non-dipolar part of the field [8]. The variations in the non-dipole field are believed to originate from electric currents flowing at the core/mantle boundary while the dipole variations are attributed to sources in the core [8]. Another interesting phenomena which is thought to originate in the Earth’s core, is geomagnetic jerks. Jerks are rapid changes in the time derivative of magnetic field components and take place in one or two years. They are best observed in the $D$ or $Y$ component and they separate periods of rather steady secular variation patterns and thus describe the change in the secular variations.
3.2 Short Term Variations

Short term variations will here be divided into regular variations caused by the orbital motion and/or rotation of celestial bodies, and the diverse irregular variations. The latter will be discussed first. Short term variations in the geomagnetic field are related to variations in the external field, where the Sun is the most dominant factor. Irregular variations are what catches one’s eye when looking at a magnetogram because of the rapid time changes in the field (often of large intensities).

3.2.1 Irregular Variations

What causes irregular variations in the geomagnetic field, is the interaction between the magnetic field of the solar wind, and the magnetosphere. The interaction, involving transfer of plasma and energy, leads to time varying currents in the magnetosphere and ionosphere, which in turn cause induced currents in the mantle and oceans.

Sudden and repeated (but not regular) changes in the magnetosphere are called a geomagnetic storm. These magnetic disturbances can cause variations of $1 - 1000$ nT and they effect all the elements of the geomagnetic field; $Z$, $H$ and $D$. The most common variations in a magnetic storm are jumps of a few hundred nT. A geomagnetic storm occurs when a large amount of energy and plasma has been transferred from the solar wind to the magnetosphere. The magnetosphere rapidly changes in strength and shape, and energy is transfered back and forth between different areas making the magnetic field lines continually re-align [4]. The influence of a geomagnetic storm can be measured at all latitudes. The first sign of an imminent magnetic storm it the storm sudden commencement (SSD). It is a sudden impulse, observed as rapid increases and decreases in the field, and is caused by a change of pressure of the solar wind against the magnetosphere [5]. The duration of a magnetic storm can be a few hours to several days and the storms are scaled from $G1$ to $G5$, depending on the amplitude of the variations. $G1$ represents a minor storm and $G5$ is an extreme magnetic storm [10].

Magnetic storms are, in fact, usually a sum of several substorms, which are minor, localized disturbances in the field - mainly evident at high latitudes [8]. Substorms can occur individually and they are believed to form when energy builds up in the magnetotail causing a magnetic reconnection [10]. The process of magnetic reconnection is when two oppositely directed magnetic field lines are brought together and magnetic energy is converted into plasma kinetic energy, making it possible for plasma to flow along the field lines [14, 16]. When this takes place in the magnetotail, particles are accelerated towards the poles, along the Earth’s magnetic field.
3.2 Short Term Variations

lines. The collision between these accelerated particles and atoms in the upper atmosphere give rise to the auroras which are described in chapter 2.4. Because of this, there is a direct connection between auroral display and magnetic disturbances. During minor perturbations in the field, auroras can be observed at high latitudes but they can only be seen at lower latitudes during intense magnetic storms. As an example of a magnetic storm, figure 3.2 shows the measured magnetic field at Leirvogur observatory on March 7th 2012. The $Z$ and $H$ components in the figure are given in nT and the declination $D$ is in degrees (measured from north in the eastward direction through east, south and west). According to the magnetogram there was a geomagnetic storm that day causing disturbances in all three components. In less than half an hour the direction of the field varied by 3 degrees. Like most of the disturbances in the Earth’s magnetic field, the frequency of substorms and magnetic storms increase and decrease with solar activity.

Figure 3.2: Magnetogram from Leirvogur observatory showing a geomagnetic storm. The $Z$ and $H$ components are given in nT and the declination $D$ is in degrees (measured from north in the eastward direction through east, south and west) [22].

Now the question is: What triggers these rapid disturbances in the magnetosphere? The Sun is continuously blowing the solar wind towards Earth, but that, in itself, is
not enough to cause large variations in the field, although it can lead to minor disturbances. The rapid and intense disturbances occur when an unusually large amount of plasma and burst of radiation is ejected from the Sun. The magnetosphere then responds by deforming drastically, leading to the measured magnetic perturbations on Earth. The Sun has a magnetic activity cycle which reaches a maximum every 11 years or so. When the Sun is in its active phase, different kinds of eruptions occur frequently on its surface and in the corona. The eruptions, resembling explosions, cause a huge amount of material and burst of radiation to be blown away from the Sun, sometimes directed towards the Earth. In section 3.2.2, these eruptions will be discussed further.

Whether the ejected solar plasma is able to break through the Earth’s magnetic shield or not, depends on its velocity, the amount of plasma traveling, and the magnitude and orientation of the IMF with respect to Earth’s magnetic field [15]. Until recently, the following theory regarding the orientation of the IMF with respect to the Earth’s magnet field was thought to be true. If the magnetic field of the incoming solar wind is aligned with the Earth’s magnetic field, the magnetosphere was thought to act like a strong shield against it, and the solar particles would then slide around the magnetosphere. But with the opposite orientation it could more easily deposit magnetic energy and radiation into Earth’s environment [14, 8]. The way the energy would then be transferred into the magnetosphere is through the above mentioned magnetic reconnection; magnetic field lines from the solar plasma would connect up to the magnetosphere, opening a "door" into the magnetosphere. However, in 2008, the team of the THEMIS mission came to the opposite conclusion. According to the THEMIS satellites, twenty times more solar particles penetrate through the Earth’s magnetic shield when the two fields are aligned [24]. This came as an astonishing surprise, revealing that this subject obviously needed to be studied in great detail. But still today, the energy transfer into the magnetosphere is generally thought to occur through reconnection.

There are other issues to be mentioned regarding irregular variations in the Earth’s magnetic field. Disturbances in the geomagnetic field arise when the density and amount of particles is increased in the ring current, the ionosphere and the other current systems in the magnetosphere. A current system which has not yet been mentioned, is the current circling the polar cap area in the auroral zones. The strong magnetic fields followed by auroral events cause charges to move, giving rise to large ionospheric currents. The strongest is a high latitude current, called auroral electrojet [15], which is strongest on the night side and can have significant impact on the magnetic field. Deviations in the magnetic field, caused by an enhancement in

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2The corona is an extension of the Sun’s atmosphere and is made out of low density plasma [6].
3The IMF is the magnetic field that the solar plasma carries with it and was discussed in chapter 2.4.
4THEMIS is a NASA mission which consists of 5 satellites that measure and observe the behavior of the Earth’s magnetic field and the plasma in it.
the auroral electrojets, are called magnetic bays. They are typically a few hundred nT [15] and of rather regular shape (bay-like), lasting for one or two hours [5]. When looking at the shape of the magnetosphere in figure 2.5, one can intuitively conclude that the amount of plasma is more on the night side of Earth than the day side. It should thus not come as a surprise that disturbances during the night time occur far more often than during the day. Even when solar activity is low, there are often small perturbations in the field during the night time. In addition to the stream of particle radiation from sun, the solar X-ray and UV-radiation can amplify the ionospheric current systems so much that large affects are measured in the geomagnetic field [5]. It would be possible to go into much greater detail about these spontaneous short term variations, describing the currents in the ionosphere and magnetosphere, but since this is only a brief overview, this will have suffice.

3.2.2 The Active Sun

The events on the Sun that mainly lead to geomagnetic storms on Earth are the solar flares and coronal mass ejection (CME). A solar flare is an explosive event which produces a burst of radiation that can be observed in all regions of the electromagnetic spectrum. Flares are thought to emerge on the Sun’s surface from magnetic reconnection, the magnetic energy being converted into thermal energy and acceleration of solar particles, resulting in a bulk motion away from the Sun [16]. A CME is a major disruption which originates in the solar corona. The event is seen as a bubble coming out of the corona, breaking open and dumping its contents into the solar wind [16]. Figure 3.3 shows how enormous an CME event can be. Both solar flares and CME’s occur more frequently when the Sun is near maximum in activity than around the minimum. That, in itself, implies that they owe their existence to the Sun’s magnetic field. The direction of the eruptions, relative to the Earth, determines whether they effect on the Earth’s magnetosphere, or not. If it is not directed towards the Earth, it will not affect its near environment. Even if it is directed towards Earth, the dense package of plasma can diverge before reaching it.

Yet another solar phenomena associated with magnetic activity on Earth are coronal holes. The fastest solar wind is found to emerge from these holes [16]. Coronal holes are regions where the magnetic field lines in the corona are not closed but open into space. This means that the coronal material is free to stream out from the Sun [16]. An easy way to determine whether the Sun is active or not is by observing the number of sunspots it has. Sunspots are dark regions that appear and disappear on time scales ranging from days to years, and emerge due to a local, intense magnetic activity [16]. They are always associated with the solar activity, moving towards the Sun’s equator when its activity increases. One can thus predict geomagnetic disturbances by observing the sun in different ranges of the electromagnetic spectrum, looking for sunspots, CME’s, solar flares coronal holes.
3 Time Variations

Figure 3.3: Coronal Mass Ejection (CME). Picture taken by the SOHO satellite. Radiation from the Sun itself is blocked and the white circle shows the size of the Sun [23].

3.2.3 Regular Variations

Regular variations of the geomagnetic field are due to the Earth’s rotation and its orbital movement around the Sun, as well as the orbital motion of the Moon. The *diurnal variations*, also called solar daily variations, are the most prominent of the regular variations. It is believed that electric currents flowing in the E-layer of the dayside ionosphere are the main cause of the diurnal variations. However, electric currents in the Earth’s crust, produced by electromagnetic induction, also contribute to the variations [8]. During the daylight hours on Earth, the solar radiation ionizes the upper atmosphere, increasing the density of ions that make up the currents in the ionosphere. The current system in the ionosphere, which causes the diurnal variations, is driven by winds from the day-night temperature difference and electrically conducting tidal winds caused by the gravitational forces of the Moon and the Sun [10]. On the night side, the currents in the ionosphere are negligible. The amplitude of these slow modulations of the field are of the order of 10-100 nT and figure 3.4 shows their characteristic behavior and latitudinal dependence [5]. The regular diurnal variations depend on the time of year, solar activity and the geomagnetic latitude. Due to its relatively small intensities, the diurnal variation is only visible in magnetograms on days when there are no stronger magnetic disturbances in the magnetosphere, produced by other phenomena. Those
3.2 Short Term Variations

Figure 3.4: The diurnal variation of the magnetic field components $D$, $H$ and $Z$ on solar quiet days ($S_q$ variations), observed at different latitudes [8].

days are called solar quiet days, $S_q$. The corresponding days, when the magnetic field is active, are called solar disturbed days, or $S_D$ [8]. Observatories often choose the five most quiet days, and other five most disturbed days, in each month of the year to calculate the average diurnal variations. When the $K$-indices are evaluated, the average $S_q$ variations must be removed from the data.

Other regular variations, which all show small magnetic effects, are the following: 27 days period which corresponds to the rotation of the Sun’s most active latitude zones. The lunar variations, which are small variations of only a few nT, occur because the gravitation of the Moon causes atmospheric tides [5]. The third is a yearly variations, with an amplitude of a few nT, possibly due to the Earth’s orbital motion around the Sun, and its tilting with respect to the plane of its orbit [5]. It is interesting to look further on the yearly variations by the help of observatory data. According to the geomagnetic observations in Leirvogur, the perturbation in the geomagnetic field varies with the seasons of the year. The field gets disturbed more often in the autumn and spring (close to the equinoxes) than at other times of the year. Figure 3.5 shows the total number of 3-hour time intervals with a $K$-index larger than or equal to 7 (indicating high magnetic activity) for each month of the year during the years 1972-2010. It is believed that the reason for this, is that the solar wind, which causes the field variations, has the most effect when the center of the Sun is in the same plane as the Earth’s equator. That happens when Earth is at the equinoxes [25]. At those times, it would be more likely that the solar plasma hits the Earth. Finally, it is worth stating that external field variations depend on the activity of the Sun and thus reaches a maximum in intensity every 11 years or so.
3 Time Variations

Figure 3.5: The total number of time intervals with $K$-index larger than or equal to 7 (indicating large magnetic disturbances) for each month during the years 1972-2010 [25].
4 Measurements

Measurements of the Earth’s magnetic field are informative in a number of ways, mostly already mentioned in the text. They make it possible to study the interior structure of the Earth and the complex current dynamo in its core which produces the geomagnetic field itself. The measurements give information on the currents in the ionosphere and magnetosphere along with the induced currents in the ocean and mantle. They reveal relations between magnetic disturbances and solar activity, as well as relations of magnetic activity to auroral display. Measurements of the magnetic field also make it possible to monitor magnetic storms, with the conceivable result of predicting their behavior in the future. Another application is the construction of magnetic charts which are constructed from the determination of the magnetic field vector measured over large areas. These magnetic measurements are carried out by enthusiastic scientists all over the world studying the different phenomenas. The most important factors regarding geomagnetic measurements are accuracy, temporal continuity and vast geographical coverage. Before going into details of the measurements themselves and the instruments, a short overview of the history of magnetic measurements is presented, followed by a general discussion on modern geomagnetic measurements.

The compass was the first instrument used to measure the geomagnetic field, employed to measure the field direction. According to ancient literature, the Chinese were the first to take advantage of the practical use of magnetism by using compasses for nautical navigation in the eleventh century A.D. This kind of application of the compass is not mentioned in European literature until about 1190 A.D., but by that time it was probably a well known instrument. The declination of Earth’s magnetic field, i.e. its deviation from geographical north, was most likely known in China long before Europeans discovered it, which was about 1450 [5]. The observation that the compass needle tilted with respect to the horizontal plane marked the discovery of inclination in about 1510. More than 200 years later, in 1785, Charles-Augustin de Coulomb found a way to measure the intensity of the horizontal component. But it was not until 1832, when Carl Fredrich Gauss discovered a way to measure the total intensity of the magnetic field, that techniques for measuring all the geomagnetic field components were available. At that time, regular measurements of the field components had been made and it was already known that all components varied with time. Alexander von Humbolt, a German geographer, naturalist and explorer, was very enthusiastic about the widespread establishment of magnetic observatories,
for then it was possible to get a broad view of variations in the geomagnetic field. In 1834, Gauss and Weber took part in von Humbolt’s project and in 1841, about fifty observatories were a part of the network \[5\]. This marked the beginning of a global coverage of geomagnetic measurements.

At present time, measurements of the Earth’s magnetic field vector are carried out at approximately 170 geomagnetic observatories around the world \[10\]. A geomagnetic observatory is a place where continuous measurements of the geomagnetic field are made for long stretches of time with the best possible accuracy. Modern observatories produce their data in digital form and depending on the quality of the observatory, one-minute or one-second-average digital data are reported with the accuracy of 1 nT, which corresponds to the angular accuracy of 5 arcseconds \[13\]. Note that the frequency of the data acquisition can be lower but that is unusual. Observatories typically consist of the following: A fluxgate variometer that measures the intensity of the field in \(X\), \(Y\) and \(Z\) direction, a fluxgate theodolite, which yields the declination and inclination of the field, and a proton precession magnetometer, which measures the total intensity. Absolute measurements have to be performed regularly (preferably once a week) at observatories. They are performed to correct for instrumental drift and other possible errors. Absolute measurements and the observatory devices are described in more detail in chapter 4.2. The location of an observatory needs to meet certain criteria, they cannot, for example, be close to any magnetic disturbances, but more on geomagnetic observatories in chapter 4.1.

In addition to the magnetic observatories, which have well defined operation rules, magnetic field measurements are also carried out at magnetic repeat stations. A repeat station is where the magnetic field is measured repeatedly every few years (ideally 2 years) \[13\]. Repeat stations are established with the aim of constructing a local map of the field or to get a better geographical coverage of magnetic field data. Most often they operate in connection with temporary research programs and only in use for a short time in comparison with the magnetic observatories. Figure 4.1 shows how repeat stations have been distributed over the world since 1900. In general, the same type of instruments are used at repeat stations and magnetic observatories. While repeat stations do improve the global coverage of geomagnetic data, a true global coverage demands the establishment of automatic observatories. Currently, there are no automatic observatories that reach the standards which a typical magnetic observatory fulfills. Such observatories would lead to great improvement in the field of geomagnetism. The magnetic measurements stated above are all ground based and do not tell the whole story. The magnetic field vectors can also be obtained by magnetic instruments in space, using orbiting satellites. The main difference between satellite and ground based magnetic measurements is that the former gives a global coverage with only one instrument while the latter needs hundreds. However, only ground based measurements will be described here.
4.1 Geomagnetic Observatories

The goal of every observatory is to have the highest possible accuracy and to avoid gaps in the recordings. To produce high quality data, the geomagnetic measurements need to be homogeneous over long time and have an accuracy of 1 nT. The method of measuring the field at an observatory needs to be very consistent so the exact same routines need to be followed year after year. One of the things that compose the regular routine of an observatory is a weekly calibration, which is obtained by doing an absolute measurement. The recordings of modern observatories should ideally be universally accessible, for instance on the internet (preferably showing the real-time vector field) and every year observatories should release their final data [13].

The site of an observatory needs to meet a number requirements for the observatory to satisfy certain criteria. It should be able to produce data within the desired accuracy and to be able to operate for a long time - hundreds of years even. Thus, the site has to be chosen carefully. Here are some of the things that need to be thought of:

The magnetic field in the observatory area should be homogeneous and, as a reference, the magnetic intensity at two points, separated by 10 meters, should not differ by more then a few nT [5]. The environment around the observatory, within a radius of several km distance, needs to be checked, because there cannot be any crustal magnetic anomalies close by, possible induced currents, nor other kinds of changes in magnetic properties. Factors like the high conductivity of seawater and man maid disturbances also need to be taken into account. For the observatory to be as independent of changes in the near environment as possible (for example growing residential areas), it is good to reserve a large area for the observatory [13].
4 Measurements

Magnetic observatories can be a part of a larger research complex, operating at the same location, or it can consist of only the basics needed for the magnetic field measurements. A typical magnetic observatory consists of five separate buildings or huts: the main building, a variation house, an absolute house, an electrical hut and a proton magnetometer house. The distance from the electricity hut to the sensors (located in the variation and absolute house) should not be less than 15 m for the sensors not to be disturbed by it. Variometers are designed to be extremely sensitive so the material in the buildings housing them has to be non-magnetic. The variometers are placed on very stable pillars. In case local conditions make the construction of stable pillars difficult, other solutions are available, such as suspended sensors (which are quite common), or the recording the tilt. As an example of the importance of stable variometers, a shift of only one arcsecond is can cause a 0.24 nT change in a field of 50,000 nT [13].

The temperature dependence of fluxgate variometers is 1 nT/°C so the temperature in the room (or box) where the variometer is placed, has to be very stable [5]. The temperature should be determined with an accuracy greater than 0.5 °C but it is common practice to change the temperature twice a year following the average temperature outside. The variation room is just a shelter for the variometer because nowadays they hardly need any service after instillation. Before, with the older instruments, they had to be visited on a daily basis.

In the absolute house, only one non-magnetic pillar is needed and its stability is not as important as for the variometers [5]. An azimuth mark (a sighting mark with known azimuth), at a minimum distance of 100 m from the absolute house, needs to be in a line of sight from the pillar (seen out of a window), and the geographical direction of the mark has to be known. It is important for the instrument to have an accurate position on the pillar so there should be fixed slots for the instrument to be placed in. Displacements of the instruments to the sides causes an error in the declination.\(^1\) The further away the mark is located, the less sensitive the measurement becomes of the displacement. The proton precession house only needs to contain a pillar and the main building can store data loggers, computers, clocks and other devices. An uninterrupted power supply is very important, since the aim is to operate the observatory continuously without any gaps in the recordings. To avoid gaps, all systems should be supported by backup batteries in case of a power failure and run secondary systems in case of equipment failure. The secondary systems are often older systems that have been replaced. To be able to use the data from the secondary system, the recordings of the two types of systems should be compared at regular intervals [5]. To get an idea about the layout of a typical observatory, an overview of Leirvogur observatory is shown in figures 4.2 and 4.3.

\(^1\)Assuming the azimuth mark to be at a distance of 100 m distance, then a displacement of 1 cm of the instrument makes a change of 0.3 arcminutes in declination [5].
4.1 Geomagnetic Observatories

Figure 4.2: The layout of Leirvogur magnetic observatory [21].

Figure 4.3: Photograph of Leirvogur magnetic observatory, taken by Þorsteinn Sæmundsson. The picture is taken from the house Nýbaer in figure 4.2, and the houses in the picture are (from left) Móðabær, Flosabær, Miðbaer and Vesturbaer [22].

All objects close to the magnetometers need to be specially tested to make sure that
they are non-magnetic. In the variometer room, the magnetic field from an object, at a distance of 0.5 from the sensor, should be less than 1 nT [5]. Where the absolute measurements are performed, the sensitivity of the measurement is even higher.

The standards for geomagnetic observatories are decided by INTERMAGNET. INTERMAGNET is a global network of magnetic observatories which advocates magnetic observatory to produce high-quality data, and makes geomagnetic recordings easily accessible on the internet. For an observatory to be a part of this network it has to produce 1-minute data and fulfill other standard requirements set by INTERMAGNET. All data produced within this international organization are thoroughly checked by a team of experts before being published [13]. The number of participating observatories is constantly increasing, and in January 2010, 110 magnetic observatories worldwide were taking part [13]. The distribution of the observatories is shown in figure 4.4, where one can see vast areas without any observatories. This is a drawback which INTERMAGNET wants to improve, so they are always encouraging more observatories to participate in the project. Recently, INTERMAGNET has encouraged observatories to upgrade their data products to perform 1 Hz data (1 second data) and to constantly correct their data acquisition with a temporary baseline [13]. This is done to meet the modern needs of the scientific society.

![Global distribution of geomagnetic observatories that participated in INTERMAGNET as of January 2010 (dots). The circles show the recently established observatories located in remote areas. [13].](image)

### 4.2 Magnetometers

The role of a magnetic observatory is to continuously measure and report the geomagnetic field vector. The instruments used for measuring the field vector are typically a triaxial fluxgate variometer (here referred to as a variometer), a fluxgate
4.2 Magnetometers

The triaxial fluxgate variometer is an instrument that continuously records three vector components of the magnetic field ($X$, $Y$ and $Z$) and thus yields the total field vector. The proton magnetometer also operates continuously but he only measures the total intensity of the magnetic field vector. The total intensity values, obtained from the variometer and proton magnetometer, are constantly being compared because a change in the difference between them could mean a problem in one of the instruments.\(^2\) Finally, the fluxgate theodolite is used, together with the proton magnetometer, for a weekly calibration of the variometers. The fluxgate theodolite measurement, which is done manually, reveals the declination and inclination of the geomagnetic field vector, and simultaneous measurements, performed by the proton magnetometer, give the total intensity of the field. By combining the results from the two instruments, an absolute value of the geomagnetic field vector is obtained. This procedure is therefore called an absolute measurement.

The variometer is calibrated by simultaneously comparing the components of the total geomagnetic field vector, obtained from the absolute measurement to that from the fluxgate variometer recordings. The differences between the vector components are referred to as baseline values for the variometer, where one baseline value is obtained for each component of the field [13]. The final data, reported from the observatory, have the baseline values added to the values recorded by the variometer [5]. The reason why the absolute measurement needs to be performed on such a regular basis, is to correct for possible pillar instability due to tilting or rotation, temperature variation or drift of the instrument due to ageing [13]. This means, that each week a new baseline value is obtained for each of the three field components. The values are usually plotted as a function of time, and a line fitted through the points is called a baseline. The quality of an observatory data is directly connected to the homogeneity of the baseline. A continuously changing baseline indicates a variometer drift or other kind of instrumental instability, which of course is a bad thing. However, a clear step in the baseline may indicate some kind of an event, for instance a rotation of the variometer (done intentionally for it to be directed towards magnetic north), or the changing of the room temperature to follow the mean outdoor temperature. Figure 4.5 shows examples of the baseline values of $Z$, $H$ and $D$ components, and the adopted baselines, for the year 2010 in Leirvogur magnetic observatory. No obvious change in the baseline values can be observed, except for the two large steps, which are associated with the changing of the temperature in

\[ F = \sqrt{X^2 + Y^2 + Z^2}. \]
4 Measurements

the variometer house. Finally, it is worth mentioning the importance of all clocks connected to the different instruments showing the exact same time, to one second accuracy. This is also required by the simultaneous measurements mentioned earlier, where the values from different instruments are compared at a certain time.

![Graph showing adopted baselines at Leirvogur magnetic observatory in 2010][21]

**Figure 4.5**: The adopted baselines at Leirvogur magnetic observatory in 2010 [21].

4.2.1 Fluxgate Magnetometers

There are various types of fluxgate magnetometers available and they are all based on one or more fluxgate sensors. The principle of a fluxgate sensor is that its output is proportional to the component of the external field in the direction of the sensors axis [20]. Now, the focus here will be on the application of the instruments and not on how they work, so the following simplified explanation of the fluxgate sensor will have to suffice. The basic fluxgate sensor consists of a core made of soft magnetic material, which should be easily saturable and have high permeability.
4.2 Magnetometers

The core has two coil windings around it, the excitation coil and the pick-up coil [1]. Figure 4.6 shows an example of a fluxgate sensor. If the excitation coil carries an alternating current, with the frequency $f$, a magnetic field within the core will be induced [5]. This causes the magnet to saturate and unsaturate, back and forth, whereas its magnetization follows the hysteresis curve of the material. If there is no external field, the induced signal in the pick-up coil will have the same frequency as the excitation current. However, in the case of an external field, like the Earth's magnetic field, the pick-up coil will measure a signal of other harmonics, additional to those with the frequency $f$. The amplitude of these secondary harmonics, caused by the external field, is proportional to the intensity of the external field's component along the sensor [5]. Thus, the signal in the pick-up coil yields the intensity of the magnetic field component aligned with the sensor.

Figure 4.6: Fluxgate sensor consists of a soft magnetic core and two coil windings around it; the excitation and pick-up coil [5].

The instruments at modern magnetic observatories, based on fluxgate sensors, are the triaxial fluxgate variometer and the fluxgate theodolite. The triaxial fluxgate variometer consists of three orthogonal fluxgate sensors, mounted in a marble cube, where two lie in the horizontal plane and one in the vertical. There are two types of variometers available, one where the marble is suspended and one where it is not. By suspending the sensors, the instrument will show much smaller drift. The difference in drift between the two types can reach 100 nT per year [2]. A suspended fluxgate variometer is shown in figure 4.7, where the cube is suspended in two crossed phosphor-bronze strips [2]. The instrument is usually oriented in such a way that one of the horizontal sensors is directed towards the geomagnetic north and the vertical sensor is aligned with the Z-component [13]. The instrument is connected to digital data logger equipment where the signal from each sensor is converted and modified to give the total intensity of each field component, in units of nT [21].

A soft magnetic material has the ability to get magnetized but does not stay magnetized like permanent magnets.
4 Measurements

Figure 4.7: Suspended triaxial fluxgate magnetometer. The fluxgate sensors (a) are mounted in a marble cube (b) and suspended with phosphor-bronze band (c) [13].

final data is obtained by adding the baseline values to the field components, after correcting for possible sensor misalignments.

The fluxgate theodolite is a non-magnetic theodolite with a fluxgate sensor attached along the theodolite’s telescope. The instrument is shown in figure 4.8. The telescope of the theodolite (and the attached fluxgate sensor), can be rotated in the horizontal and vertical plane. Its angles, in the two separate planes, can be read from an eyepiece just beside the eye-piece of the telescope [9]. By observing the output of the fluxgate sensor, the component of the total field vector along the sensors and telescope’s axis is found. Further properties on the fluxgate theodolite can be found in section 4.2.3.

Figure 4.8: DI fluxgate theodolite (left) and the scales of a theodolite (right) [3, 7].
4.2.2 Proton Precession Magnetometer

Proton precession magnetometers are absolute instruments that measure the total intensity in the magnetic field. The proton’s spin (its intrinsic angular momentum) tends to align with the external magnetic field, and the technique of these magnetometers is based on the precession frequency of the proton’s angular momentum, \( \mathbf{L} \), around an external magnetic field \( \mathbf{B} \). The angular frequency of the precession is given by \( \omega = \mu B \), and is called the Larmor frequency. The constant \( \mu \) is the gyromagnetic ratio which depends on a particle’s charge and mass only, and \( B \) is the magnitude of the external field, \( B = |\mathbf{B}| \). The gyromagnetic ratio for protons is known with high accuracy, so if the precession frequency of \( \mathbf{L} \) around \( \mathbf{B} \) can be measured, the total intensity of the external field can be calculated. A proton magnetometer consists of a bottle filled with proton rich fluid, like alcohol or water, and a coil system wound around it. A direct current running through the coil induces a magnetic field in the fluid which polarizes the protons in the direction of the field, causing a net magnetization [1]. When the artificial field is turned off, the proton’s angular momentum vector will start precessing around the Earth’s magnetic field with the Larmor frequency, proportional to the strength of the field. The protons’ precession induces an AC voltage of the same frequency, and by dividing the frequency with the gyromagnetic ratio, the strength of the Earth’s magnetic field can be deduced [13, 1]. The electronics have to be able to measure a frequency with an accuracy of \( 10^{-5} \). To give an example of the precession frequency, protons in a 45,000 nT field will precess with frequency close to 2 kHz [1].

4.2.3 Absolute Measurements

Once a week, an absolute measurement is carried out at magnetic observatories, according to a fixed routine. The inclination and declination of the geomagnetic field is found by the use of a DI-flux, which is the fluxgate theodolite shown in figure 4.8, while the simultaneous total intensity of the field is obtained by a proton magnetometer located in the area. For the absolute value of the field vector to be obtained, the value of the field’s intensity needs to be shifted from the location of the proton magnetometer to the absolute pillar for it to represent the same vector as the inclination and declination. This means, e.g. that before or after the \( I \) and \( D \) measurement, the difference in the total field intensity, between the two locations, needs to be evaluated. That is done by placing another proton magnetometer on the absolute pillar and comparing the field intensities obtained by the two magnetometers simultaneously.

The determination of the inclination and declination is based on finding the positions of the sensor, aligned with the telescope, where the output is close to zero. Since
the sensor’s output equals the magnetic field vector along its axis, those positions correspond to the sensor being almost perpendicular to the Earth’s magnetic field [9]. The procedure of the absolute measurement, as it is carried out at Leirvogur observatory, will be briefly describe here. Detailed derivation of the problem and its trigonometry is thoroughly described in [9]. For starters, the instrument needs to be levelled so that the horizontal circle (see figure 4.8) is horizontal, and the axis of rotation vertical. A mark with a known azimuth is sighted and the telescope’s horizontal angle of the sighting mark is noted. Thereby, the geographical north can be "marked" on the horizontal scale of the theodolite and the declination of the magnetic field determined.

The procedure of the measurement can be divided into two; the declination measurement and the inclination measurement of the field vector. First, the horizontal component of the magnetic field is found by adjusting the sensor to be horizontal and finding its position where the output is close to zero. Several positions of the sensor fulfill the desired condition, they occur when the sensor is on top and bottom
4.2 Magnetometers

of the telescope, and the telescope is directed approximately East and then West. For each of the positions, the horizontal angle and the output of the sensor is noted. From this information, the declination of the geomagnetic field vector can be found.

Figure 4.9 shows the scheme of an absolute measurements performed at Leirvogur observatory, where A, V, N, S correspond to East, West, North and South, and the signs ↓ and ↑ refer to the sensor being on top (↑) or under (↓) the telescope. The left side of the figure corresponds the declination measurement described above and the right side of the figure corresponds to the following inclination procedure. Note that the different positions of the sensor cancel the effect of the possible difference between the alignment of the axis of the telescope and the sensor’s axis [9].

In the latter part of the measurement, the inclination of the magnetic field is found by aligning the sensor with the magnetic meridian, whereas the magnetic meridian can be found from the earlier declination measurement. When in the magnetic meridian, the sensor is rotated about the horizontal axis, again placed at the position where the sensor’s output is close to zero, where the vertical angle and the output is noted. As can be seen in figure 4.9, four measurements are done with the sensor on top and telescope facing South and then other four where the telescope faces North. The simultaneous total intensity of the field, measured from the proton magnetometer is also noted for each measurement. At the bottom of the right side of figure 4.9, one can see that the difference in the total intensity of the magnetic field vector between the location of the absolute pillar and the proton magnetometer has been found. From these series of measurements, the inclination, declination and the total intensity of the geomagnetic field vector at the absolute pillar can be deduced.
5 Summary and Conclusions

The magnetic field observed at the surface of Earth is a superposition of magnetic fields generated by different sources. The most dominant source by far, is the complex current system in the Earth’s highly conducting and liquid outer core. The Earth’s magnetic field is a vector field so three components are needed to fully describe the field at a given point, usually the horizontal component, vertical component and the declination of the vector are specified. Spherical harmonics can be used to describe the magnetic field on the Earth’s surface, and are commonly applied in the construction of global magnetic charts. Direct measurements of the geomagnetic field and the study of paleomagnetism, reveal that the geomagnetic field changes with time. Long term variations, also called secular variation, which are on the scale of several years to a million years, are due to the dynamics within the Earth’s interior - those producing the main field. Variations on a shorter time scale are due to currents in the magnetosphere and ionosphere, which in turn cause induced currents in the Earth’s crust and oceans. These magneto- and ionospheric currents originate from the changing shape of the Earth’s magnetosphere, which is a response to the incoming the solar wind.

Continuous measurements of the Earth’s magnetic field are performed by magnetic observatories worldwide, which have the combined goal of recording magnetic data with the highest possible accuracy and to avoid gaps in the recordings. The observatories communicate with each other and exchange data, together making a global network of magnetic observatories. The main instruments of a modern geomagnetic observatory are the triaxial fluxgate variometer, the fluxgate theodolite and the proton precession magnetometer. The field components are continuously and automatically measured by the triaxial fluxgate variometer but a weekly calibration needs to be performed by hand, using the fluxgate theodolite and the proton precession magnetometer.

If the reader is interested, the field of space weather and satellite based magnetometers could be suitable for further reading. There are many exciting ongoing, and forthcoming missions, with the focus on examining the Earth’s magnetosphere and its interaction with the solar wind. That is something definitely worth following. It is good to keep in mind that each chapter is only a summarization of the main details of the corresponding subject. For further reading on the different subjects, the cited books and articles are recommended.
Bibliography

institute, 1979.


