1 Introduction

1.1. Motivation and Objectives

Ever since the ground breaking publication of Wegener (1915), who postulated continental drift based on geological evidence, and the following works of other authors on the topography (e.g. Hess, 1962) and magnetic structure of the ocean floor (Vine and Matthews, 1963; Pitman and Heirtzler, 1966), the theory of plate tectonics explains the history of formation, movement and distribution of the continents and tectonic plates. Wegener's theory can explain a wide and diverse range of topics such as fossil findings (Wegener, 1915) and migration patterns of turtles (Carr and Coleman, 1974).

McElhinny and McFadden (2000) consider the publications of Irving (1956) and Runcorn (1956) as key points for the paleomagnetic aspect of plate tectonics, as both authors showed that measurements of the direction of magnetization of rock samples allow for a reconstruction of the paths of continents. Since then, more than 9259 paleomagnetic poles have been published (Pisarevsky, 2005) and, together with other data, allow for a reconstruction of the history of the continents for more than the last 500 Ma (Gubbins and Herrero-Bervera, 2007). However, many questions remain to be answered, in particular about the oceanic plates: obtaining oriented rock samples for detailed paleomagnetic studies is challenging, and therefore their history of movement is not known in similar detail as of the continents (Sager, 2006). Most of the published data stems from measurements on only partially oriented core samples or inversion of ship-based magnetic field measurements (Beaman et al., 2007) and thus do not give a complete view of the magnetization.

Detailed information about plate movements in the past also serves as validation for large scale simulations of mantle flow (Steinberger and Calderwood, 2006). An observation of the processes within the mantle is only possible using seismic tomography (e.g. Boschi et al., 2009), which, due to the lower number of earthquakes and seismic observatories, is limited in resolution below the oceanic plates (French and Romanowicz, 2015). Additionally, seismic tomography can only give a present-day snapshot of the state of the mantle; here, the past movements of the plates determined from paleomagnetic measurements can be used to assess the validity of flow models.

Island chains of volcanic origin play an important role in the discussion about the dynamics of the upper mantle and the origin of intraplate volcanism: following the hypothesis that some of them are formed by narrow conduits of magma uprising from Earth's mantle, so-called mantle plumes (Morgan, 1971), the past movement of these island chains gives insight into the convection patterns of the mantle.

However, the problem of the paleomagnetic data coverage for oceanic plates remains, as many of the volcanoes that have formed the island chains are now extinct and the islands have submerged and are no longer easily accessible. Therefore, sampling these so-called seamounts is mostly limited to dredging and drilling. Dredged rock samples can only be used to estimate the strength of the paleomagnetic field and most drill core samples are azimuthally unoriented and only allow for a measurement of the inclination of magnetization (Gubbins and Herrero-Bervera, 2007).

Some solutions that allow for a determination of the declination are described by Morris et al. (2009), who for their part match structures both visible in oriented images of the borehole wall and visible on drill cores to reorient drill core samples from Integrated Ocean Drilling Program (IODP) Expedition 304/305. Fontana et al. (2010) give a comprehensive overview of core reorientation techniques using borehole images and state that image matching is impeded by the different diameters of borehole and core, which causes a difference in the vertical amplitude of structures. Hurst et al. (1994) (as cited by Morris et al., 2009) were the first to use a submersible (the Alvin) to gain oriented cores of the uppermost meter of the oceanic crust. Allerton and Tivey (2001) use a remotely operated drill system that can be lowered on a wireline to collect oriented samples of the uppermost meter of the crust. The orientation of the drill is determined using a magnetic compass, and technical measures assure that the core receives a continuous marking indicating its orientation. However, the compass is considerably affected by the metallic parts of the drill rig. A compensation is possible in parts, but the compass data becomes invalid when the slope of the seafloor is above 10° (Smith et al., 2001), which, in combination with the shallow depth of penetration, limits the use of the drill system.

The approach chosen in this thesis for the determination of the declination of magnetization is the measurement of the geographically oriented magnetic field vector within a borehole and the subsequent calculation of the vector of magnetization.

The capabilities of oriented magnetic field measurements have long been recognised, and were initially focussed on the detection of off-borehole bodies (Levanto, 1959; Silva and Hohmann, 1981). Here, for example, they can be used to unambiguously determine the location of pipes or other metallic objects (Ehmann, 2010).

Bosum and Scott (1988) were the first to deploy an oriented downhole magnetometer in order to determine the direction of magnetization of oceanic crust during Ocean Drilling Program (ODP) Leg 102 (see also Shipboard Scientific Party, 1986; Bosum et al., 1988, for more detail). Further measurements were supposed to be made during ODP Leg 106/109, but the magnetometer got damaged and could only measure the vertical magnetic field (Shipboard Scientific Party, 1988; Bosum and Kopietz, 1990). A second japanese magnetometer used during Leg 106/109 was only able to measure the total magnetic field (Shipboard Scientific Party, 1988; Hamano and Kinoshita, 1990).

During ODP Leg 118, an oriented magnetometer built by the USGS (Scott and Olson, 1985) was used to conduct measurements of the magnetic field (Shipboard Scientific Party, 1989; Pariso et al., 1991). A german oriented magnetometer was supposed to be run during ODP Leg 148, but the mechanical gyro was not delivered in time, so only the vertical and horizontal components could be measured (Shipboard Scientific Party, 1993).

The Göttingen Borehole Magnetometer (GBM, Leven, 1997), which was used for the measurements during IODP Expedition 330, was introduced to ODP during Leg 197 (Gaillot et al., 2004). It was also deployed during IODP Expedition 304/305 (Expedition 304/305 Scientists, 2006) and IODP Expedition 351. The GBM is, to my knowledge, the only scientific borehole magnetometer that uses fiber optic gyros for orientation, which is a considerable advantage over mechanical gyros, whose function is negatively affected by the shocks and vibrations occurring within a borehole.

IODP Expedition 330 to the Louisville Seamount chain took place in the south-west Pacific Ocean from 13 December 2010 to 11 February 2011. A major goal of the Expedition is the determination of possible past movement of the Louisville Hotspot by means of paleomagnetic measurements. The GBM was deployed during the Expedition in order to aid with the determination of the direction of magnetization, in particular the declination. There are several challenges associated with oriented measurements of the magnetic field and the calculation of the magnetization vector:

- Reorientation of the magnetic field data
- Choosing the appropriate model for the data
- Estimating the errors

A key part of vector magnetic measurements is the reorientation of the magnetic field data from the internal reference frame of the tool to the geographic reference frame. Here, algorithms used by one tool can not necessarily be used by a different tool without modification, as the type, amount and data rate of sensors typically differ. There are several publications describing the reorientation methods used for GBM data (Klein, 2009; Ehmann, 2010; Virgil et al., 2010, 2011; Virgil, 2012; Virgil et al., 2015; Ehmann et al., 2015), but every new data set presents its own challenges. A key aspect of this thesis is therefore the enhancement of existing algorithms and the development of new algorithms, with a special focus on sensor-fusion methods for gyro and inclinometer data.

After reorientation of the magnetic field data, the information about the magnetization has to be extracted. Existing publications on the interpretation of downhole magnetic field measurements (both oriented and unoriented data) typically use horizontal layers intersected by a circular borehole as a representation of the subsurface (Bosum et al., 1988; Pozzi et al., 1988; Fieberg, 1994; Nogi et al., 1995; Williams, 2006). Other authors suggest inclined circular layers (Gallet and Courtillot, 1989) or rectangular layers (Hamano and Kinoshita, 1990). As the lateral extent of the drilled structures is not always known, typically an infinite outer radius is chosen. Virgil (2012) uses elliptical disks to interpret data from the Outokumpu formation and provides detailed information about the influence of the geometry and outer radius of the layers on the magnetic field. I am focussing on models using horizontal and inclined layers of infinite extent. In particular, I am introducing an approximation for the magnetic field of inclined layers that allows to separate data modeling in an inversion for horizontal layers and a subsequent consideration of possible dips and azimuths.

The geometry chosen for the models of the subsurface has a large effect on the results. Based on a model of circular inclined layers given by Gallet and Courtillot (1989), Tivey et al. (2005) discuss the influence of the layer dip on the inclination calculated for ODP Leg 129 data and find that, for example, a layer dip of 50° can cause an error in the inclination of magnetization of 70°, but they do not consider the effect of the layer azimuth. Gallet and Courtillot (1989) discuss the effect of dip and azimuth of a layer, but mostly focus on details of the magnetic field and not on errors in the determination of the magnetization. Other authors (Steveling et al., 2003, for example) acknowledge the importance of the geometry of the layers for the interpretation of the data, but do not give an estimate of the error caused by a potentially wrong geometry. Here, my approximation for inclined layers can be used to efficiently assess the influence of uncertainties in the geometric parameters of a layer on the direction of magnetization. Without further knowledge about the magnetization and geometry of an inclined layer, a linear dependence of the error in the declination on the error in the azimuth of the layer should be assumed.

1.2. Thesis Structure

In Section 2.1, I give an introduction to the theory and history of Earth's magnetic field, followed by a summary of paleomagnetism (Section 2.2) and magnetic potential field theory (Section 2.3). The main purpose of these sections is to recall the most important aspects into the memory of the reader and to include all information and background necessary for an understanding of this thesis. The following Section 2.4 gives a historic overview of the methods used for magnetic field measurements in boreholes as well as details of the Göttingen Borehole Magnetometer, the tool that is used for the measurements described in this thesis. Chapter 3 then gives more detail about Expedition 330, its scientific background and the questions that it was supposed to answer.

The following chapters form the core of this thesis: they guide through the data processing steps necessary to transform raw data into interpretable values (Chapter 4), different methods of reorienting the magnetic field data (Chapter 5) as well as different methods of data interpretation (Chapter 6). After describing reorientation methods using one fiber optic gyro and inclinometers (Section 5.3), explaining the standard method for data reorientation developed during recent years (Section 5.4) and a new method for treating uncorrected offsets of the fiber optic gyros (Section 5.5), I am introducing two new reorientation methods based on a sensor fusion of gyro and inclinometer data by means of a Kalman filter (Section 5.6). The following reorientation of the Louisville magnetic field data is then conducted using the most appropriate algorithm for each data set.

As an interpretation using horizontal layers causes a discrepancy between the calculated magnetizations and the magnetization measured on drill core samples (Section 6.5), it becomes necessary to incorporate information about the geometry of the subsurface and to use inclined layers for the interpretation of the data (Section 6.7). Most importantly, I am introducing a new approximation for the magnetic field of inclined layers that is closely related to the horizontal layer interpretation, and which is faster than a complete, accurate calculation of the magnetic field. This approximation also enables us to quickly assess possible ambiguities during the inversion (Section 6.7.3 and following).

Finally, in an exemplary analysis of an igneous layer that is both identifiable in images of the borehole wall and in the magnetic field, I derive a value for the declination of magnetization that agrees with existing theories about the movement of the Louisville hotspot (Section 6.7.8).

In a concluding remark I demonstrate that there is at least a theoretical possibility to obtain some information about the declination of magnetization from unoriented measurements of the magnetic field, provided that high quality geometrical information and magnetization data from drill cores are available (Section 6.8).

2 Scientific Background

2.1. Earth's Magnetic Field

The use of the magnetic field of the Earth for navigation predates the understanding of its origin by several centuries. Chinese travellers have used magnetized needles floating on water for orientation as early as the beginning of the 10th century (Lowrie, 1997). Around 1600, William Gilbert was the first to describe that the magnetic field of the Earth resembles that of a magnetic rock approximately aligned with the axis of rotation of Earth (Telford et al., 1990). In 1838, Carl Friedrich Gauss, who was also the first to determine absolute values of the magnetic field intensity, used the spherical harmonic analysis he invented to show that the source of Earth's magnetic field is primarily within the Earth itself (Gubbins and Herrero-Bervera, 2007).

For a long time, the origin of the magnetic field was thought to be magnetized rocks within the Earth, but that theory could neither explain the time varying behaviour of the magnetic field nor that the expected temperature of Earth's core and mantle is in large parts above the Curie temperature of all ferromagnetic materials (see Section 2.2.1). Only in 1939, Walter Elsasser published his first ideas on a dynamo mechanism within the electrically conductive fluid outer core of the Earth that is now the accepted theory for the origin of Earth's magnetic field. (Gubbins and Herrero-Bervera, 2007)

The magnetic field at a particular point on Earth is commonly described using either its geographic components, i.e. north, east and vertical downward (B_N, B_E and B_V), or its total value and two angles: inclination I and declination D. The inclination is the angle of the magnetic field respective to the horizontal plane; the declination is the angle between the projection of the magnetic field on the horizontal plane and geographic north. The total intensity of the magnetic field is given by:

$$B = \left|\vec{B}\right| = \sqrt{B_N^2 + B_E^2 + B_V^2} \tag{2.1}$$

Figure 2.1b shows the global coordinate system of the Earth, given by longitude and latitude in comparison to the local definition of the coordinate system used for the magnetic field (Fig. 2.1a). Also shown is the horizontal component of the magnetic field B_H, defined by:

$$B_H = \sqrt{B_N^2 + B_E^2}. \tag{2.2}$$

The magnetic field is measured in units of Tesla ($1\,\mathrm{T} = 1\,\mathrm{Vs/m^2}$), but due to the relatively low strength of Earth's magnetic field, it is usually given in nT or µT.

Earth's magnetic field is mainly dipolar in nature and only about 25% are attributed to non-dipolar components. At the time of writing of this thesis, the axis of the dipole does

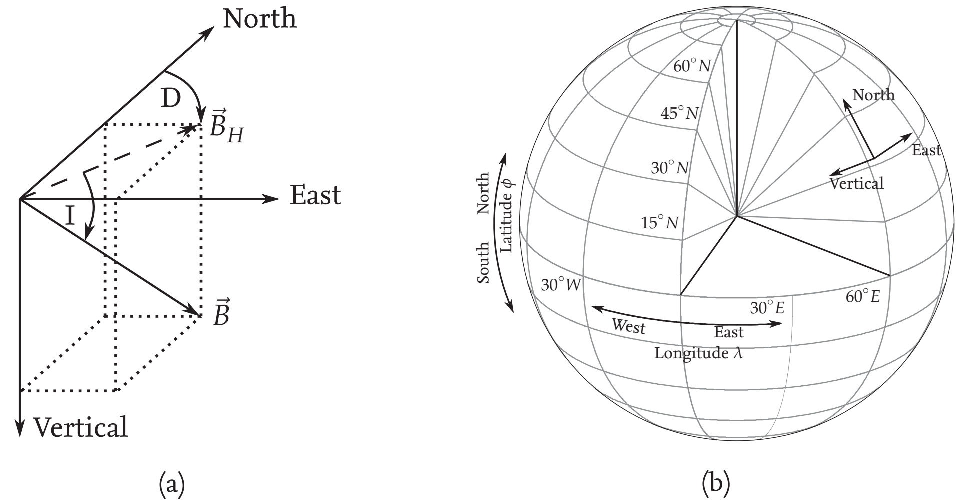

Figure 2.1.: a) The local reference frame for the magnetic field. I: Inclination, D: Declination, B_H: Horizontal Magnetic Field, $\vec{B}$: Magnetic Field Vector. (adapted from Merrill, 1996) b) The global coordinate system with longitude λ and latitude ϕ in comparison to the local reference frame, North, East and Vertical. (adapted from de Lange, 2013)

not coincide with the axis of rotation of the Earth but is inclined by about 10° and offset from the center of the Earth by 575 km (calculated using formulas given in Fraser-Smith, 1987 and values given in Finlay et al., 2010).

Measurements of the direction of the magnetization of rock samples and models of magnetic field anomalies of the seafloor show that the direction of the magnetic field of the Earth has reversed several times within history. The mechanism of those so-called pole reversals is chaotic in nature and not yet completely understood. The current state of the magnetic field with a magnetic south pole near the geographic north pole is referred to as the normal state of the magnetic field, in contrast to the reversed state. (Gubbins and Herrero-Bervera, 2007)

Depending on the duration of a particular state, the history of the magnetic field is divided into polarity chrons, with durations from 50 kA to 5 Ma, subchrons, with durations from 20 kA to 50 kA, and short lived (i.e. about 10 kA) magnetic excursions, during which the magnetic pole starts to wander in the direction of the equator, but does not reverse completely. The formal definition of a magnetic excursion requires that the magnetic pole is at an angular distance of more than 45° from the geographic pole. As long as the magnetic pole is closer than that, the variation of the magnetic field is counted as part of the regular, long-term variation of the magnetic field, the so-called secular variation. (Gubbins and Herrero-Bervera, 2007)

An important point for the differentiation between magnetic excursions and a full reversal is the time that it takes for the magnetic field of Earth's solid inner core to reverse, which is believed to be in the order of 9.2 kyr. This magnetic "inertia" has to be overcome for a complete reversal of the magnetic field (Gubbins, 1999; Gubbins et al., 2015). Currently, the magnetic field on average reverses approximately each 250 kA, but, for example, the so-called Cretacious Normal Polarity Superchron that started about 121 MA ago had lasted for 38 MA (Lowrie, 1997). It is currently not possible to reliably predict the time of the next reversal of the magnetic field.

The magnetic field is also subject to variations of shorter timescales which are of external origin and are mainly caused by interactions of the solar wind with the magnetosphere, the magnetospheric ring current and ionospheric current systems. The periods of those variations vary between several years, e.g. variations related to the solar cycle with a period of 11 years and typical amplitudes of 10 nT to 20 nT, daily variations with typical amplitudes of 20 nT to 100 nT, and so-called pulsations with periods between 0.2 s and 600 s. The strongest variations can occur during solar storms with amplitudes of several hundred nT and periods of several hours to weeks (Gubbins and Herrero-Bervera, 2007).

Approximately 180 geomagnetic observatories continuously monitor the temporal and spatial variability of the magnetic field. In conjunction with satellite measurements, these observatory data are used to generate models of Earths magnetic field that can be used to estimate strength and direction of the magnetic field anywhere on Earth. The most commonly used models are the World Magnetic Model (WMM) and the International Geomagnetic Reference Field (IGRF). These models are limited in so far that they only include the magnetic field originating in the fluid outer core and their accuracy depends on the density of measurements within a certain region. Crustal magnetization and other sources of magnetic fields can cause differences of several hundreds of nT between the modeled field and the actual value of the magnetic field at a particular location (Gubbins and Herrero-Bervera, 2007).

Figure 2.2 shows the total value of the magnetic field (see Eq. 2.1) calculated for January 1st, 2011, at an altitude of 0 m above the reference ellipsoid, using the International Geomagnetic Reference Field (IGRF, Finlay et al., 2010). Typical strengths of the magnetic field are in the order of 30000 nT in equatorial and 50000 nT in polar regions. Figure 2.3 shows the distribution of the inclination of the magnetic field and Fig. 2.4 the distribution of the declination. The maps use a Mercator projection, which is frequently used for navigation purposes but which diverges at the poles, and are therefore truncated at latitudes of $\pm 75°$. When the magnetic field is used for navigation, especially the spatial and temporal variability of the declination has to be taken into account, as this can otherwise cause significant errors in the determination of the direction of travel. Consequently, the magnetic declination for the area in question is commonly shown on navigational maps, and those maps are updated frequently to account for temporal variation of the declination.

If the magnetic field of Earth was purely dipolar, with a dipole in parallel to the axis of rotation, the isolines in the figure for the inclination would be in parallel to the meridians; the declination would be constantly zero everywhere on Earth.

The prominent region of low total magnetic field strength that overlaps parts of South America and the Atlantic is referred to as South Atlantic Anomaly (SAA). Within this region, with minimum values of down to 22500 nT, the magnetic field is weakened to such an extent that charged particles of solar origin can disturb satellites and, during solar storms, endanger astronauts (Heirtzler et al., 2002). The magnetic field in this region has been decreasing since more than 50 years and there are no signs that the decrease is slowing down (Finlay et al., 2010). The SAA is believed to be associated to a large low shear wave velocity province underneath southern Africa. Recent research suggests that it potentially is the origin and trigger of many of the past geomagnetic reversals (Tarduno et al., 2015).

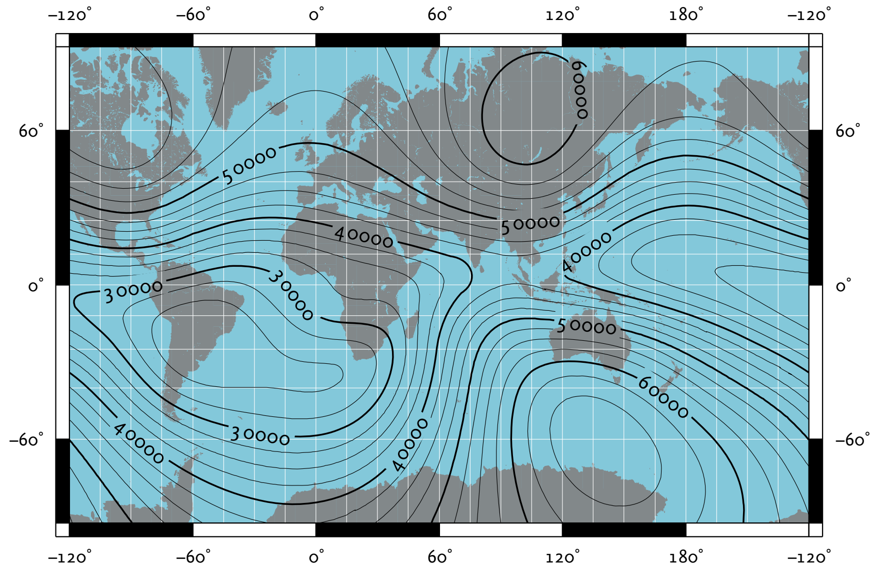

Figure 2.2.: Total value of the magnetic field according to the International Geomagnetic Reference Field (IGRF). Values are given in nT. Map generated using IGRF-11 data (Finlay et al., 2010).

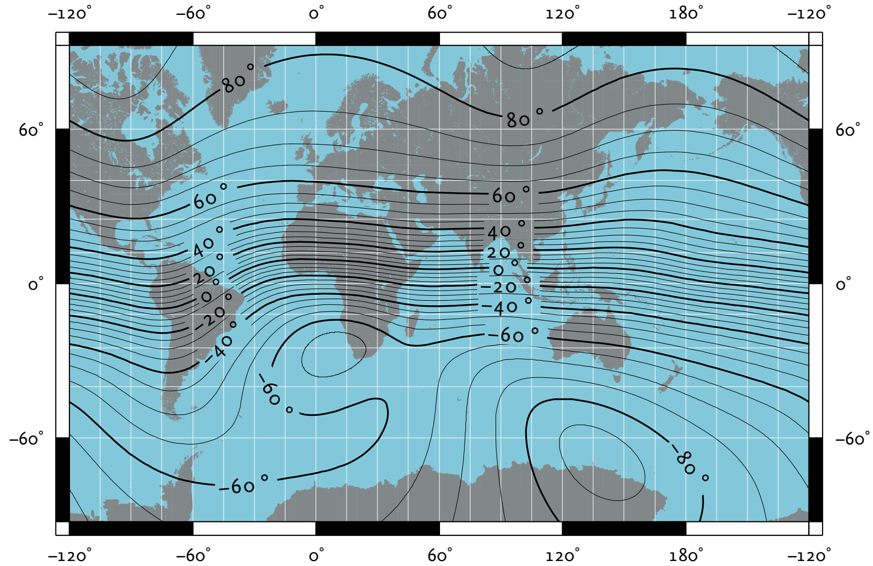

Figure 2.3.: Inclination of the magnetic field according to the International Geomagnetic Reference Field (IGRF). Map generated using IGRF-11 data (Finlay et al., 2010).

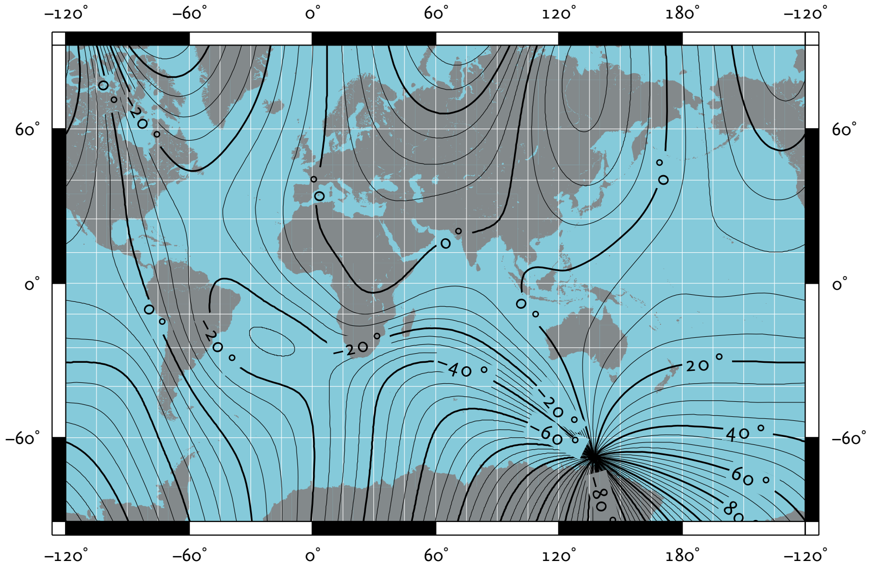

Figure 2.4.: Declination of the magnetic field according to the International Geomagnetic Reference Field (IGRF). Map generated using IGRF-11 data (Finlay et al., 2010).

2.2. Palaeomagnetism and Plate Tectonics

2.2.1. Magnetic Properties of Rocks

In order that be able to learn about the tectonic and magnetic history of Earth by sampling rocks and sediments, we need to understand how the physical properties of different materials influence their suitability for paleomagnetic studies. Depending on their magnetic properties, materials can be divided into several categories:

Diamagnetic: In diamagnetic materials, the induced magnetization opposes the inducing field, that is they have a negative magnetic susceptibility. All materials are, at least in parts, diamagnetic, but this is often masked by other effects. Examples of diamagnetic materials are salt, quartz and calcite. Their susceptibilities are low, mostly in the order of -10^{-6} (Lowrie, 1997).

Paramagnetic: In paramagnetic materials, existing magnetic moments of atoms are aligned in parallel to an external field. They have small positive temperature dependent susceptibilities in the order of 10^{-4} and lose their magnetization as soon as the external field is turned off. Examples are amphibole and olivine (Lowrie, 1997).

Ferromagnetic: In these materials, existing magnetic moments of atoms strongly interact and form so called magnetic domains of uniform magnetization, even without an external magnetic field. Ferromagnetic materials have high positive susceptibilities, which depend non-linearly on the external magnetic field and on their history of magnetization.When an external field is applied to a ferromagnetic material, the magnetization increases until all atomic moments are aligned in parallel to the external field. After the so-called saturation magnetization M_s is reached, an increase