# *Palaeozoic Geography and Palaeomagnetism of the Central European Variscan and Alpine Fold Belts*

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# Contents

# Glossary

## Abbreviations

| | |
|---|---|
| AF | alternating field |
| AM | Armorican Massif |
| AMS | anisotropy of magnetic susceptibility |
| APWP | apparent polar wander path |
| ATA | Armorican Terrane Assemblage |
| CAI | conodont colour alteration indices |
| CHRM | characteristic remanent magnetisation |
| CT | Catalan terrane |
| EEP | East European Platform |
| HCF | Holy Cross Fault |
| HCM | Holy Cross Mountains |
| IM | Iberian Massif |
| IRM | isothermal remanent magnetisation |
| MD | multidomain |
| MGCR | Mid-German Crystalline Rise |
| MORB-type | mid-oceanic ridge basalts |
| NBT | Noric-Bosnian terrane |
| NRM | natural remanent magnetisation |
| PSD | pseudo single domain |
| SIRM | saturation isothermal remanent magnetisation |
| STB | Saxo-Thuringian basin |
| TB | Tepla-Barrandian |
| TESZ | Trans-European Suture Zone |
| VFB | Variscan fold belt |

## Symbols

| | |
|---|---|
| $\alpha_{95}$ [°] | radius of the 95% confidence circle of a palaeomagnetic direction |
| Dec/Inc [°/°] | declination/inclination of site means |
| dm/dp [°/°] | 95 % confidence limits of the latitude/longitude of a magnetic pole position |
| $D_o$ | observed declination |
| $D_r$ | reference declination |
| $S_o$ | observed strike |

$S_r$ ........ reference strike
k ........ precision parameter for the dispersion of palaeomagnetic directions
$P_J$ ........ anisotropy degree
R ........ correlation coefficient
T ........ shape parameter
$k_{max}/k_{int}/k_{min}$ ........ principal susceptibility anisotropy axes
N ........ number of samples measured
n ........ number of samples used in calculation of site mean
Lat/Long [°/°] ........ pole latitude and longitude
$T_C$ [°C] ........ Curie-temperature
$T_b$[°C] ........ blocking temperature
$T_{ub}$[°C] ........ unblocking temperature
$H_c$[T] ........ coercive force
$H_{cr}$[T] ........ remanence coercivity
$M_s$[$Am^2/kg$] ........ specific saturation magnetisation
$M_{rs}$[$Am^2/kg$] ........ specific saturation remanence

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# Zusammenfassung

(Summary in German)

Während durch *Kossmat* [1927] erstmals die Gliederung des variszischen Gebirges beschrieben wurde, so ist es heute allgemein bekannt, dass die dominierende geologische Struktur Mitteleuropas durch die komplexe Kollision einzelner Krustensegmente im Rahmen der caledonischen und variszischen Orogenese im Paläozoikum geprägt wurde. Mit Hilfe geowissenschaftlicher Methoden konnte in der Vergangenheit der Kenntnisstand hinsichtlich des komplexen Aufbaus und der Zusammensetzung Europas aus einer Vielzahl tectonostratigraphischer Einheiten vorangetrieben werden. Zum unmittelbaren Verständnis der orogenetischen Abläufe und seiner Entwicklung ist jedoch die geodynamische Geschichte der einzelnen geotektonischen Einheiten grundlegend.

Die strukturellen Grenzen der einzelnen tectonostratigraphischen Elemente Europas lassen sich auf der Basis ihres unterschiedlichen geologischen Aufbaus, der tektonischen Gestaltung sowie des variierenden Metamorphosegrades definieren. Die jeweilige geologische und strukturelle Gestaltung ist auf den unterschiedlichen Einfluss der aufeinanderfolgenden Orogenesen -der caledonischen, variszischen und der noch andauernden alpidischen Phase- zurückzuführen. Bezeichnend für die einzelnen Phasen der Kollision und Amalgation sind die jeweils spezifischen geodynamischen Verhältnisse der umgebenden Krustenelemente.

Mit dem Aufbrechen Rodinias im Proterozoikum ist das Paläozoikum von der Umstrukturierung der dominierenden Kontinentalplatten Baltica, Gondwana, Laurentia und Siberia geprägt, bis hin zur Bildung des Superkontinents Pangäa. Innerhalb dieser plattentektonischen Einheiten und der sich ausbildenden ozeanischen Bereiche trennten sich im Verlauf des Paläozoikums mehrere Mikroplatten -*terranes*- vom nördlichen Rand Gondwanas ab und kollidierten sukzessive mit den nördlich positionierten Kontinentalplatten. Diese Phase der Konvergenz und Deformation mit der aufeinanderfolgenden Kollision der einzelnen tektonischen Einheiten sowie der Schließung der sich dazwischen befindlichen ozeanischen Becken umfasste ca. 150 Ma. Diese Phase der Akkretion führte in Mitteleuropa zur Bildung des variszischen Gebirgsgürtels.

Zur Klärung der individuellen Driftgeschichte der Vielzahl der prae-variszischen Mikroplatten, die sich durch ihre spezifische und in Bezug auf die Großkontinente unterschiedliche faunistische und lithologische Entwicklung im Paläozoikum auszeichnen, ist die Disziplin des Paläomagnetismus ein probates Mittel, um unabhängige Datensätze zu erhalten.

Die Entwicklung des Faunenbestandes Avalonias und dessen paläobiogeographische Interpretationen weisen auf einen Wechsel zwischen peri-polaren und tropischen geographischen Breitenlagen zwischen Kambrium und Silur hin. Paläomagnetische Untersuchungen bestätigten

diese Annahmen zur Paläodrift Avalonias. Somit gilt die Driftgeschichte von Avalonia, das sich im frühen Ordovizium von Gondwana löste und im späten Ordoviz/frühen Silur mit Baltika kollidierte, als anerkannt. Hingegen ist die Entwicklung der ursprünglich als eine Einheit aufgefassten armorikanischen Mikroplatte noch nicht völlig geklärt. Durch die Summe der existierenden paläobiogeographischen und paläomagnetischen Daten konnte von *Tait* [1999] gezeigt werden, dass es sich vielmehr um ein Agglomerat einzelner, zum Teil unabhängiger plattentektonischer Segmente handelt. Demnach sind das Armorikanische und Iberische Massiv sowie das Catalan *terrane* im Westen neben dem östlich positionierten Tepla-Barrandium als getrennt zu betrachten. Insbesondere zum Böhmischen Massiv besteht aufgrund des dort dokumentierten Rotationsbetrags von 165° bezüglich seiner heutigen Orientierung eine signifikante tektonische Diskrepanz. Die Stellung des Saxothuringikums innerhalb des Armorikanischen *terrane* Agglomerats ist jedoch aufgrund fehlender paläomagnetischer Daten bislang fraglich. Außerdem ist die Paläogeographie und damit die Stellung paläozoischer Einheiten des alpinen Raumes innerhalb der variszischen Orogenese noch weitgehend ungeklärt. Wenngleich lithologische und paläoklimatische Daten auf ein unabhängiges Proto-Alpines *terrane* hinweisen, bieten die faunistischen und lithologischen Interpretationen genügend Spielraum, Affinitäten zum Armorikanischen *terrane* Agglomerat, aber auch zu Gondwana in Betracht zu ziehen. Die Paläogeographie der alpinen Einheiten spielt auch eine bedeutende Rolle im Hinblick auf die geodynamische Entwicklung Gondwanas.

Die geodynamische Einbindung von östlich der variszischen Front aufgeschlossenen tektonostratigraphischen Einheiten in das Szenario der sich im Paläozoikum vollziehenden Konvergenz ist ebenfalls weitgehend ungeklärt. So finden sich im Bereich der Trans-European-Suture-Zone (TESZ) kontinentale Fragmente, die im Nordosten durch den Osteuropäischen Kraton sowie im Westen und Süden von phanerozoischen orogenen Gürteln West- und Mitteleuropas eingerahmt sind. Basierend auf seismischen Studien und Bohrungen beschreibt die TESZ die Grenze zwischen präkambrischer Lithosphäre des osteuropäischen Kratons und jüngerer Lithosphäre unter neoproterozoischen und paläozoischen tektonostratigraphischen Einheiten von West- u. Mitteleuropa. Zum Verständnis der Positionierung der hier vorliegenden konträren Krustenstrukturen trägt in großem Maße die Kenntnis über den Ablauf und die Dynamik der phanerozoischen Tektonik und somit die Driftgeschichte der einzelnen tektonostratigraphischen Einheiten bei. Lediglich im Bereich des Heiligkreuzgebirges liegen Aufschlussverhältnisse vor, die eine ungestörte Probennahme zulassen. Geochronologische, paläontologische, aber auch einige paläomagnetische Daten zeigen bisher kein einheitliches Bild der geodynamischen Zusammenhänge und der geotektonischen Zuordnung der hier aufgeschlossenen tektonostratigraphischen Einheiten.

Im Rahmen dieser Arbeit wurden für paläomagnetische Untersuchungen zur Bestimmung primärer charakteristischer remanenter Magnetisierungsrichtungen aus paläozoischen Gesteinen des Saxothuringikums, paläozoischen Einheiten der Alpen sowie im Bereich des Heiligkreuzgebirges insgesamt ca. 2000 Bohrkerne an über 260 Lokalitäten entnommen. Zur Analyse ihrer paläomagnetischen Richtungsinformationen wurden die Proben schrittweise thermisch oder im Wechselfeld bzw. in einer Kombination beider Verfahren entmagnetisiert. Die

Auswertung der Messergebnisse zeigte, dass die Gesteinsproben deutlich zu unterscheidende Magnetisierungskomponenten trugen. Die Auswertung der Richtungsdaten ließ erkennen, dass es sich beim größten Teil der identifizierten Magnetisierungen um Überprägungen bzw. um die Dokumentation von Remagnetisierungsereignissen handelt. Die Zuverlässigkeit der als primär identifizierten Daten konnte hingegen durch die Anwendung von *field tests* bestätigt werden.

Als Träger der Magnetisierungskomponenten kommen aufgrund ihres Blockungstemperatur- und Koerzitivkraftspektrums sowie der gesteinsmagnetischen Ergebnisse vorwiegend Goethit, Pyrrhotit und Magnetit in Frage. In den untersuchten Gesteinen liegen somit meist gemischte Fraktionen aus verschiedenen magnetischen Mineralen vor. Auch in der Korngrößenverteilung des Magnetit liegen Mischfraktionen aus PSD- und MD-Teilchen als Träger der Magnetisierungskomponenten vor.

Aus mittelordovizischen karbonatischen Gesteinen des Malopolska Massivs im Heiligkreuzgebirge konnte im Rahmen dieser Arbeit ein Paläopol mit 11°N und 47°E und eine daraus resultierende Paläobreite für den Probenort von etwa 44°S ermittelt werden. Geochronologische Daten detritischer Muskovite unterkambrischer Klastika aus dem Heiligkreuzgebirge stützen die Annahme, dass sich das Malopolska Massiv im frühen Kambrium -noch separiert von Baltica- am Nordrand Gondwanas befand. In der Folge zeigen sowohl faunistische Daten als auch Detritus aus den Svekofenniden, welche aus Sedimenten des mittleren und oberen Kambriums des Heiligkreuzgebirges gewonnen wurden, eine zunehmende Affinität zu Baltika an. Der für das Caradoc/Llandvirn ermittelte Paläopol liegt genau auf dem mittelordovizischen Segment der scheinbaren Polwanderkurve Baltikas. Somit kann mit den in in dieser Untersuchung präsentierten paläomagnetischen Daten gezeigt werden, dass spätestens im mittleren Ordovizium das Malopolska Massiv an Baltika akkretioniert war und der Transit von Gondwana im Zeitraum vom mittleren Kambrium bis zum unteren Ordovizium erfolgte.

Die Position der einzelnen Elemente des Armorikanischen *terrane* Agglomerates am Nordrand Gondwanas während des Kambriums und frühen Ordoviziums ist u.a. aufgrund des jeweils vorhandenen cadomischen Grundgebirges allgemein anerkannt. Die in dieser Arbeit präsentierten paläomagnetischen Daten für das Saxothuringikum zeigen mit einer Paläobreite von 63°S für das untere Ordovizium eine Lage nahe dem Nordrand Gondwanas an, wenngleich geochemische Daten für diesen Bereich auf beginnende Extension und *rifting* hinweisen. Für das obere Ordovizium und das obere Silur konnten Breitenlagen von 38°S und 21°S ermittelt werden. Somit wird für das Saxothuringikum ebenso eine fortschreitende Separation von Gondwana bestätigt, wie sie bereits für weitere Elemente des Armorikanischen *terrane* Agglomerates dokumentiert wurde. Signifikante Rotationen, wie sie bislang nur im Tepla-Barrandium erfasst wurden, konnten auch im Saxothuringikum nicht nachgewiesen werden. Es ist damit eine deutliche Trennung des Saxothuringikums vom Tepla-Barrandium anzunehmen. Eine Konsolidierung der amorikanischen Einheiten erfolgte nicht vor dem mittleren Devon mit der Schließung des Rheischen Ozeans und der fortschreitenden variszischen Orogenese.

Die Separation der Proto-Alpinen Einheiten von Gondwana wird durch einen im alpinen Raum für das Ordovizium dokumentierten bimodalen Vulkanismus angezeigt und fällt damit in die Phase der kambro-ordovizischen Extension am Nordrand Gondwanas. Für das Ober-Ostalpin und das Südalpin, die als das *Noric-Bosnian terrane* die südlichen Einheiten der Proto-Alpen bilden, konnte mit Hilfe der in dieser Arbeit ermittelten paläomagnetischen Daten eine fortschreitende Separation von Gondwana gezeigt werden. Für das obere Silur und das untere Devon konnten Paläobreiten von 47°S und 31°S nachgewiesen werden. Mittel- und spätdevonische Riffbildungen sowie lagunäre Sedimentationsverhältnisse und die damit vergesellschafteten Faunen weisen für diese alpinen Einheiten auf warme klimatische Verhältnisse hin. Die hier dokumentierten paläomagnetischen Daten für das mittlere Devon mit einer Paläobreite von 25°S bestätigen die paläobiogeographischen Annahmen. Auch signalisiert die Entwicklung der Faunen im Verlauf des Paläozoikums eine zunehmende Affinität zu nördlichen kontinentalen Einheiten wie den amorikanischen Elementen, Avalonia und Larussia einerseits, sowie deutliche Unterschiede zu Gondwana andererseits. Gleichzeitig postulieren die existierenden paläomagnetischen Daten eine Distanz von ca. 1000 km zwischen dem Armorikanischen *terrane* Agglomerat und dem *Noric-Bosnian terrane*. Außerdem zeigen die südlichen Proto-Alpinen Einheiten vom Ordovizium bis in das Karbon kontinuierliche Sedimentationsverhältnisse; prae-karbone Deformationen sind dort nicht nachgewiesen. Damit sind die hier präsentierten paläomagnetischen Daten der alpinen Einheiten u.a. ein Indiz für eine südliche Positionierung Gondwanas gemäß dem X-Modell der scheinbaren Polwanderkurve für Gondwana, welches für die Silur/Devon Grenze eine steile südliche Breitenlage für den Nordrand Gondwanas postuliert. Das konkurrierende Y-Modell, nach welchem der nördliche Rand Gondwanas in diesem Zeitraum eine äquatoriale Position einnimmt, hätte demgegenüber eine Kollision mit den alpinen Einheiten zur Folge. In Verbindung mit paläoklimatischen Modellen für Gondwana, welche -basierend auf lithologischen Interpretationen- Gondwana ebenso in hohen südlichen Breiten vermuten, wird mit den hier präsentierten paläomagnetischen Daten die Plausibilität des X-Pfades für die scheinbare Polwanderkurve Gondwanas bekräftigt.

Im unteren Karbon akkretionierten die südlichen Proto-Alpen mit den nördlich gelegenen, Äquator nahen und bereits konsolidierten kontinentalen Einheiten. Die abschließende Kollision Gondwanas mit der Schließung der Paläotethys führte im späten Karbon/frühen Perm zur Bildung des neuen Superkontinents Pangäa.

# Chapter 1 Introduction

## Chapter 1.1 Palaeomagnetism

In the second half of the 20th century new terms like sea floor spreading, plate tectonics and many others have been coined in order to describe the geodynamic processes of the mobile earth. The various disciplines of earth sciences contributed to the new geodynamic theory in different ways and intensity. Some of them could only provide qualitative data for the interpretation of the plate movements and for possible palaeogeographic scenarios. Among the geophysical methods palaeomagnetism was able to provide quantitative and independent data about the past locations and distributions of land masses and ocean basins yielding the first quantitative paleogeographic maps for the geological past. During the last 40 years palaeomagnetism became therefore a powerful tool to solve problems related to the movement of continents and continental fragments and could help to decipher the processes by which continents grow and mountain belts are formed.

The dominant advantage of the palaeomagnetic methods is the hypothesis that the earth magnetic field can be described as a geocentric axial dipole (GAD hypothesis). One of the most important features of this hypothesis is that there is a direct relationship between (palaeo)latitude and (palaeo)inclination. The remanent magnetisation of a rock unit, defined by the measured inclination, declination and the magnetic intensity, yields information about the palaeofield. This allows the reconstruction of palaeolatitudes using the palaeoinclination. Other palaeogeographic techniques, such as biogeography and climatology can provide useful information, but they are also influenced by a number of external factors such as oceanic currents, global climate etc. thus providing only qualitative information about ancient continental margins, relations and affinities between tectonostratigraphic units and their apparent palaeolatitudes. Furthermore, from the magnetic declination value, important information with regards to the amount and sense of rotation respectively to magnetic north of the sampled tectonic unit can be deduced. From geological methods alone it is difficult and uncertain to separate and calculate rotations of tectonic units in a comparable precision.

In addition, the geomagnetic field is not constant over geological periods. Apart from short term random variations of the orientation (secular variations) and variations in the magnitude of the dipole moment the dipolar geomagnetic field switch the polarity. The present configuration of the dipole field is referred to as normal polarity, the opposite configuration is defined as reversed polarity. This essential feature of the geocentric axial dipole with normal-polarity and reversed-polarity intervals yield to geochronologic applications of palaeomagnetism: magnetic polarity stratigraphy. This technique has been applied to a stratigraphic correlation and geochronologic calibration of rock sequences ranging in age from Pleistocene to Precambrian.

Magnetostratigraphy has developed into a major subdiscipline within palaeomagnetism and has joined stratigraphers and palaeontologists with palaeomagnetists to solve a wide variety of geochronologic problems and provide major refinement of stratigraphic correlations and geochronologic calibrations of both marine and nonmarine fossil zonations.

However, one major problem with palaeomagnetism if the often insufficient conservation of the ancient magnetic fields by the rocks over geological times. Mineralogical processes during the lithification of sediments and during the cooling of volcanic and plutonic rock sequences as well as postgenetic alteration of the primary minerals can affect the carriers of the remanent magnetization and thus modify the primary remanence acquired during the rock formation, often resulting in complete remagnetisation of the rocks. This is a particular problem when studying palaeozoic and older rocks of orogenic belts, or strongly weathered rockes where magnetic overprinting is common.

Palaeomagnetic techniques were first developed in the middle of last century Since then, the sensitivity and accuracy of instrumentation have improved such that even the most weakly magnetised rocks can now be measured. The increased understanding of magnetisation and remagnetisation processes themselves and the improvement in analytical techniques have led to significant improvements in the reliability and quality of data. The application of palaeomagnetic techniques is now very versatile and a reliable tool for solving complex geological problems.

Nevertheless, the chances of sampling remagnetised rocks is fairly high, although in the field they are apparently unmetamorphosed and undeformed. The high failure rates are very common in palaeomagnetic studies, as remagnetisation can often only be determined through detailed laboratory experiments.

For further details concerning the fundamental principals and analytical techniques involved in palaeomagnetism the reader is refered to following textbooks by [*Soffel*, 1991], [*Butler*, 1992], [*Tarling and Hrouda*, 1993], [*Van der Voo*, 1993], [*Opdyke and Channell*, 1996] and [*Dunlop and Ozdemir*, 1997] and references therein.

## Chapter 1.2 General Geodynamic and Tectonic Framework of Central Europe

Present day Europe consists of several palaeozoic and older blocks or massifs. This configuration is the result of successive convergence and amalgamation stages of terranes and microplates since Precambrian and Phanerozoic times. Figure 1-1 shows how Europe can be subdivided into a number of terranes and crustal domains. The marked boundaries between the individual terranes and domains are based on the contrast in surface and subsurface geology, tectonic environment and metamorphic stages. The geological and structural setting especially of Central Europe is attributed to the Caledonian, Variscan and Alpine orogenic events, the latter of which is ongoing. Each stage is the consequence of a significant geodynamic setting of crustal units.

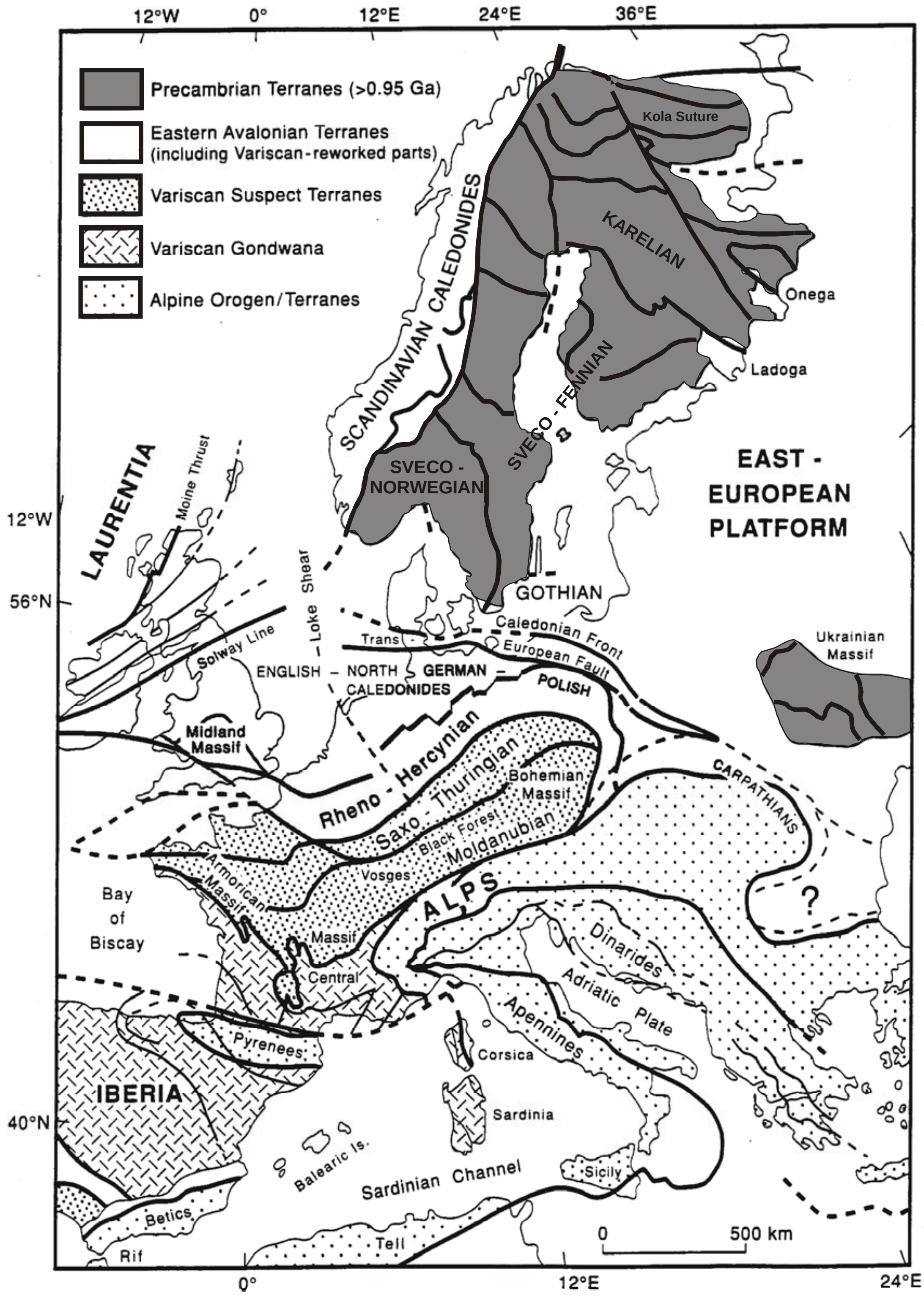



Figure 1-1: A simplified sketch map of the terrane collage of Precambrian and Phanerozoic Europe after [*Berthelsen*, 1992]. Sutures and orogenic fronts are shown as bold lines.

Since the Proterozoic the crustal evolution was dominated by the breakup and later reassemblage of crustal fragments. At around 1.1 Ga to 1.0 Ga it is assumed that the Rodinia supercontinent configuration was formed (Fig. 1-2) [*Meert et al.*, 1994] [*Smethurst et al.*, 1998b] [*Dalziel*, 1991]

[*Hoffman*, 1991]. From approximately 750 Ma to early Cambrian times Rodinia underwent a period of desintegration [*Torsvik et al.*, 1996]. This resulted in separation and formation of a number of large continental plates such as Baltica, Laurentia, Gondwana and Siberia which played key roles in Palaeozoic palaeogeography. Gondwana was finally amalganated in the Pan-African-Brasiliano orogeny which culminated approximately 550 Ma. With the increasing faunal evolution and diversification from Vendian and Cambrian times, the now independent crustal units are marked by their specific and individual faunal evolution. These data are used as an indicator for palaeobiogeographic interpretations.

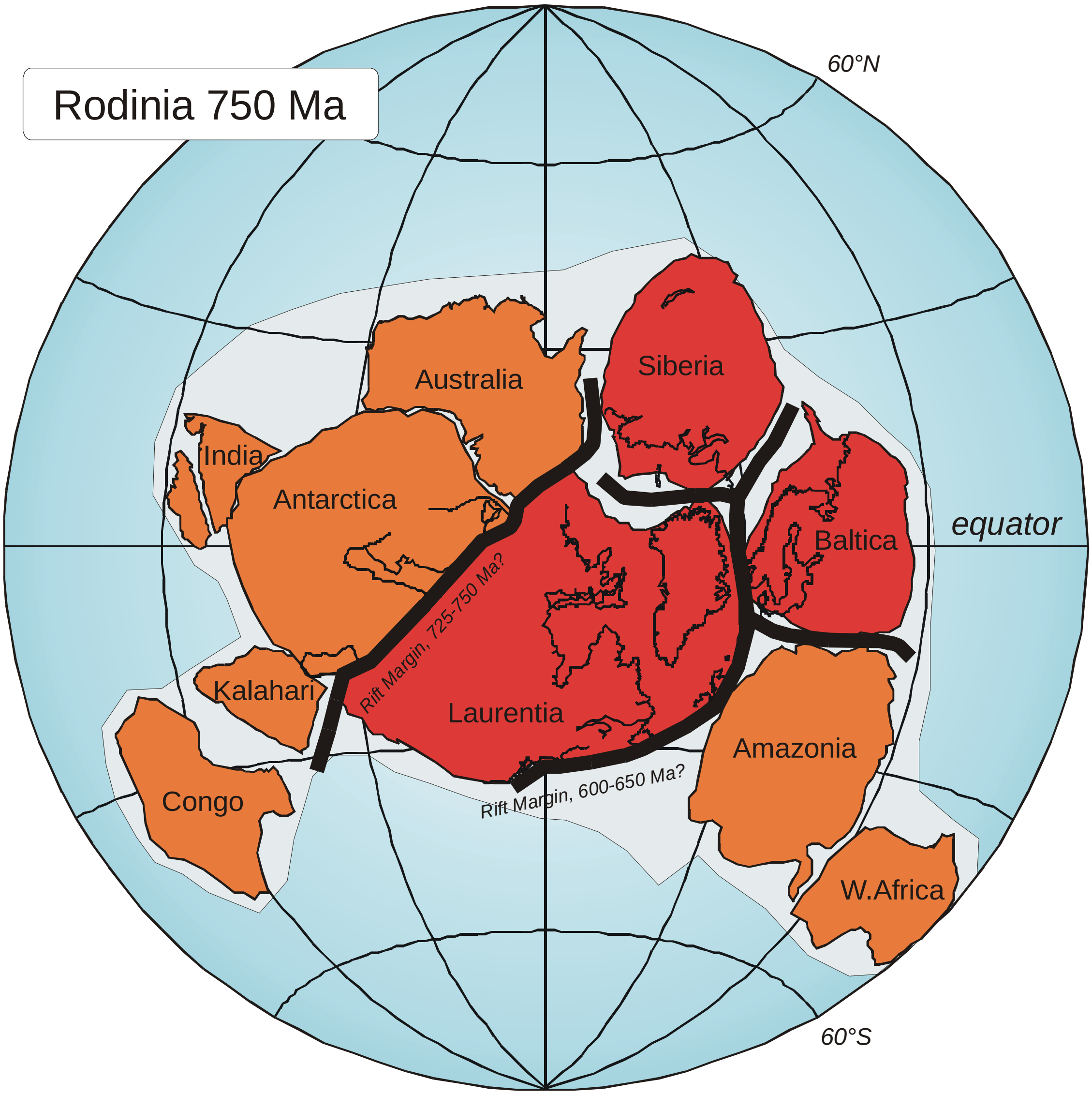



Figure 1-2: Rodinia configuration of 750 Ma after [*Smethurst et al.*, 1998b].

The main phase for the amalgamation of Central Europe occurred during the Palaeozoic with the predominant structure of the European Variscan fold belt (VFB), a complex orogenic system, which developed during Caledonian and Variscan orogenies. Convergence of Baltica, Laurentia and Gondwana and intracontinental deformation lasted for some 150 Ma and the subsequent collision of these major continental plates resulted in closure of at least four Palaeozoic oceanic basins – the Iapetus, Tornquist, Rheic and Galicia-Massif-Central. In this continental configuration of Gondwana, Baltica and Laurentia a number of pre-Variscan suspect terranes or microplates are incorporated. Most of these terranes have Cadomian basement, indicating Late Precambrian/Cambrian affinity with the northern margin of Gondwana which was at high southerly palaeoalatitudes in early Palaeozoic times. However many of these pre-Variscan terranes of the VFB e.g. Avalonia and Armorica are characterized by their specific and different faunal and lithological evolution for the Palaeozoic with respect to the major continents. The Alpine realm of Europe also contains Palaeozoic units, which are generally accepted as having been part of the northern margin of Gondwana throughout the Cambro-Ordovician. Geological and faunal evidence, however, points to the existence of a Palaeozoic Proto-Alpine terrane.

Geological, palaeobiogeographical and palaeomagnetic advances have led to significant improvments in understanding the geodynamic connections within the late Palaeozoic Variscan fold belt. However, many details regarding palaeolatitudinal drift histories and the affinities between different terranes and continents remain unclear.

## MEMBERS OF THE FRAMEWORK OF CENTRAL EUROPE

With regards to the apparent polar wander paths (APWP) of the major plates controlling the Early Palaeozioc palaeogeography and the development of the Caledonian and Variscan belts, those of Baltica and Laurentia are now well established [*Torsvik et al.*, 1992] [*MacNiocaill and Smethurst*, 1994] and their palaeogeographic positions since the Palaeozoic are now fairly clear. On the other hand the scenario of Gondwana remains a matter of controversy. Palaeomagnetic data from Gondwana are still disputed and essentially two different models for the APWP have been proposed. A detailed consideration of this problem is beyound the scope of this thesis, but discussions are given in [*Bachtadse and Briden*, 1991], [*Van der Voo*, 1993] and [*Tait et al.*, 2000a]. The generally adopted model which is used in the following palaeogeographic reconstructions is based on the simpler and more straight pattern of the Palaeozoic APWP of Gondwana, involving a gradual and continuous northward drift of Gondwana from Ordovician through Carboniferious times.

According to [*Cocks and Fortey*, 1982] Avalonia (including Nova Scotia, New Brunswick, Southern Britain, Ardennes and the Rhenohercynian zone) was defined as an independent microplate based on interpretations of faunal assemblages which show a change from peri-polar to tropical latitudes from Cambrian to Silurian times. Palaeomagnetic studies [*Trench and Torsvik*, 1991] [*Torsvik et al.*, 1993] confirmed this interpretation and solved the detachment, drift and attachment history of Avalonia from Gondwana to Baltica in more detail and the palaeodrift of Avalonia during the Palaeozoic is now well constrained [*Cocks et al.*, 1997].

The Amorican microplate hypothesis was suggested by [*Van der Voo*, 1982] based on palaeomagnetic data from the Armorican Massif, which indicate its separation from northern

Gondwana in Early Devonian times. In the beginning it was not clear whether it was a single independent microplate or a composition of autonomous tectonostratigraphic units including the Armorican massif, the Bohemian Massif, the Iberian Massif and the Saxo-Thuringian basin. In contrast to the model of Van der Voo palaeobiogeographic reconstructions prefered a position close to the Northern margin of Gondwana [*Scotese and McKerrow*, 1990] [*McKerrow et al.*, 1991] [*McKerrow*, 2000] throughout the Palaeozoic. However detailed geological investigations [*Franke et al.*, 1995a] [*Franke*, 2000] [*Matte*, 1991] do not support the hypothesis of a single Amorican microplate and document the presence of a number of small oceanic basins which opened and closed in Palaeozoic times. Since then extensive palaeomagnetic studies in several tectonostratigraphic units were carried out and the geodynamic connections within Armorica could be documented. Tait ([*Tait et al.*, 1994a] [*Tait et al.*, 1994b] [*Tait et al.*, 1995] [*Tait*, 1999] and [*Tait et al.*, 2000b]) was able to show that the tectonic coherence of the Gondwana derived Armorican terrane must be questioned. The results of these studies demonstrated that Armorica comprises an assemblage of terranes or microblocks with a major tectonic discontinuity between Bohemia (Tepla-Barrandian) and the more westerly Iberian Massif, Armorican Massif and the Catalan terrane. The term 'Armorican Terrane Assemblage' (ATA) was suggested by [*Tait*, 1999] to describe the relations of these Palaeozoic elements.

Based primarily on palaeomagnetic data, the palaeogeographic evolution of these plates in Palaeozoic times can be summarized as follows.

## EARLY ORDOVICIAN (FIG. 1-3)

From the existing palaeomagnetic database for Gondwana, in Early Ordovician times the south pole was situated in northwest Africa [*Corner and Henthorn*, 1978] [*Lanza and Tonarini*, 1998]. During this time period it is now fairly clear and well accepted that Avalonia, the various elements of the ATA and the Southern European Palaeozoic terranes were all adjacent to the northern margin of Gondwana and situated at high southerly palaeolatitudes[*Duff*, 1979] [*Perroud*, 1983] [*Cogne and Perroud*, 1988] [*Tait et al.*, 1997]. This is based on palaeomagnetic evidence for Avalonia [*Trench et al.*, 1992] [*Torsvik et al.*, 1993], faunal and lithological data from various massifs of Southern Europe and the fact that all the terranes have either Cadomian basement, or Gondwana derived detrital material. Avalonian faunas were coincident with those from Bohemia and the Armorican Massif, but in marked contrast to those of Baltica and Laurentia [*Cocks and Fortey*, 1982] [*Cocks et al.*, 1997]. In the Llanvirnian Avalonia was drifting northwards away from Gondwana and the Rheic Ocean opened in its South. Baltica was also at fairly high palaeolatitudes, rotated with respect to its present day orientation [*Torsvik et al.*, 1992] [*Torsvik et al.*, 1996] [*Perroud et al.*, 1992] and its Ordovician faunas are significantly different to those of Gondwana and Laurentia. Based on faunal, lithological and palaeomagnetic data Laurentia drifted northwards to the equator after the break-up of Rodinia, opening the Iapetus Ocean along its present-day eastern and southern margins. Throughout most of the Palaeozoic era it was straddling equatorial latitudes with the development of warm-water carbonates and faunas. Early Orodvician palaeomagnetic data from Tepla-Barrandian zone of the Bohemian Massif which indicate palaeolatitudes of 76±14° [*Tait et al.*, 1994b] demonstrate that the ATA remained adjacent to Gondwana probably until Mid-Ordovician times, when it started to move northwards in combination with beginning volcanic activity [*Tait et al.*, 1997]. Three reliable Early Ordovician palaeopoles are available for the Armorican Massif (reliable being those which pass the first three of the reliability criteria described by [*Van der Voo*, 1990]). These data are rather scattered and indicate southerly palaeolatitudes of 55±11° (Cap de Chevre, [*Cogne et al.*, 1991]), 72±6° (Moulin de Chateaupanne, [*Perroud et al.*, 1986]), and 62±6° (Pont Rean, [*Cogné*, 1988]). No reliable data are as yet available for the Saxothuringian or for the Iberian Massif.

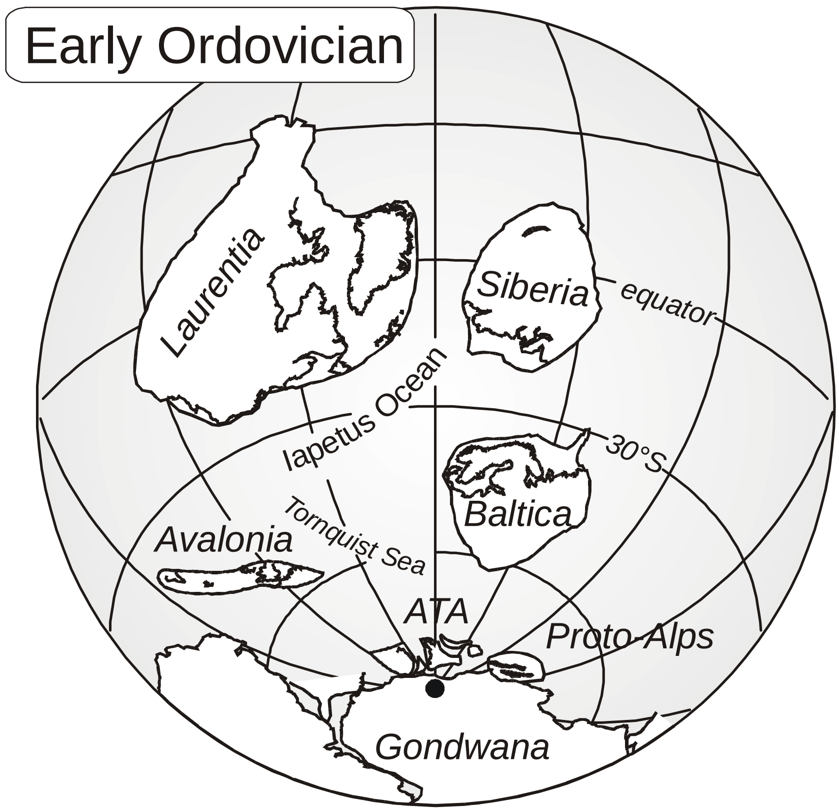



Figure 1-3: Early Ordovician palaeogeographic reconstruction based on palaeomagnetic data; for ATA see text and [*Tait et al.*, 1994b]; for Avalonia see data compilation of [*Torsvik et al.*, 1993]; for Baltica see review by [*Torsvik et al.*, 1992] and data compilation of [*Smethurst et al.*, 1998a]; for Laurentia see [*MacNiocaill and Smethurst*, 1994]; for Siberia see review by [*Smethurst et al.*, 1998b]; for Gondwana see [*Corner and Henthorn*, 1978] [*Lanza and Tonarini*, 1998].

LATE ORDOVICIAN (FIG. 1-4)

During the Ordovician, Baltica rotated about 90° counter-clockwise and drifted northwards towards the equator [*Torsvik et al.*, 1996], and by the Late Ordovician Its northern margin was in equatorial latitudes, in agreement with biogeographical indicators [*Owen et al.*, 1991]. By the Caradoc the Iapetus Ocean had narrowed significantly [*Torsvik et al.*, 1996]. Avalonia also drifted northwards and palaeomagnetic data demonstrate a palaeolatitude of 45°S [*McCabe and Channell*, 1990]. Thus Avalonia was separated from Laurentia by the southern Iapetus Ocean and from Baltica by the narrowing Tornquist Sea. The timing of closure of the Tornquist Sea is not clear from palaeomagnetic evidence alone. However, from geological evidence, it is thought to have closed in Late Ordovician to Early Silurian times [*Berdan*, 1990] and the ostracod assemblages were similar [*Pharaoh et al.*, 1993] [*Berdan*, 1990] [*Cocks et al.*, 1997].

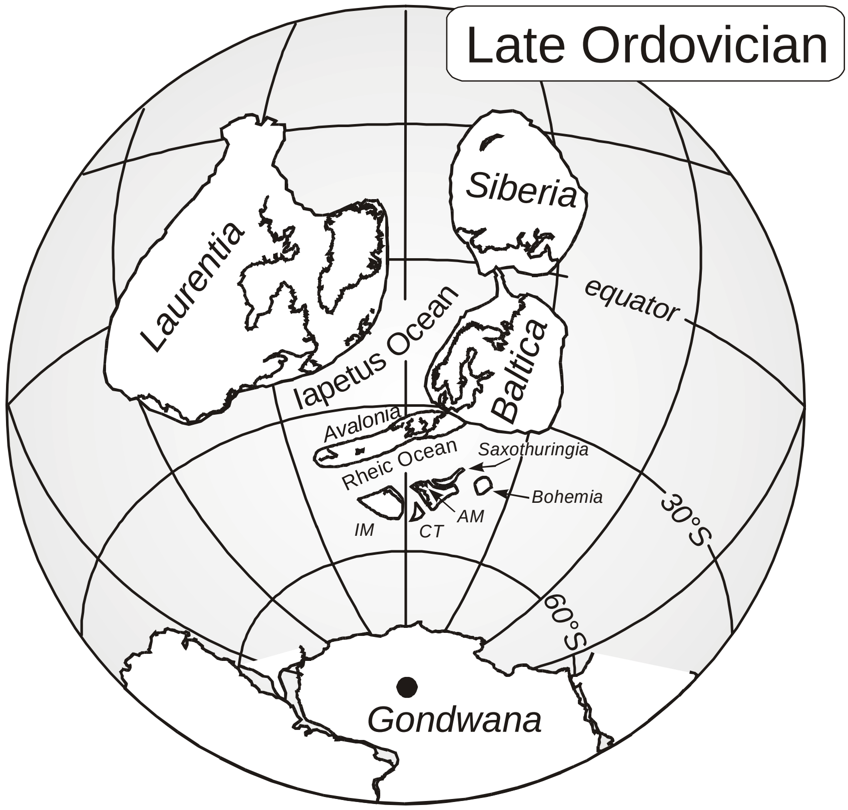



Figure 1-4: Late Ordovician palaeogeographic reconstruction based on palaeomagnetic data; for ATA see text and [*Tait et al.*, 1995]; for Avalonia see data compilation of [*Torsvik et al.*, 1993]; for Baltica see review by [*Torsvik et al.*, 1992] and data compilation of [*Smethurst et al.*, 1998a]; for Laurentia see [*MacNiocaill and Smethurst*, 1994]; for Siberia see review by [*Smethurst et al.*, 1998b]; for Gondwana see [*Schmidt and Embleton*, 1990].

It is assumed that during the Mid Ordovician the ATA began to rift from the northern margin of Gondwana and drifted northwards. Late Ordovician palaeomagnetic data of the Tepla-Barrandian point to an independent palaeogeographic position with respect to Gondwana [*Tait et al.*, 1995] and the Rheic Ocean between Avalonia and the ATA was also closing. No unambiguous palaeomagnetic data are available for the Late Ordovician of the Armorican and the Iberian Massifs [*Perroud*, 1983] [*Perroud and Van der Voo*, 1985] [*Tait et al.*, 1997]. However, strong faunal and lithological similarities in Ordovician to Devonian successions of the Armorican, Iberian and Bohemian Massifs indicate similar geological conditions and demonstrate that they were part of the same domain [*Kriz and Paris*, 1982] [*Robardet*, 1996]. Only two sets of palaeomagnetic data which pass the above reliability criteria have been published for the Armorican terrane assemblage, and these provide contrasting results. Data from Barrandia [*Tait et al.*, 1995] show intermediate palaeolatitudes (approx. 40°S), indicating separation from the northern margin of Gondwana sometimes in the early-mid Ordovician. Data from the Vendée province of the Armorican Massif, however, indicate continued peri-polar palaeolatitudes (approx. 76°S; [*Perroud and Van der Voo*, 1985]). From Ordovician data alone, it remains equivocal as to whether they, and the Saxo-Thuringian unit, remained adjacent to northern Gondwana in Late Ordovician times, or if they rifted away in Mid Ordivician times with the Tepla-Barrandian. However, faunal and lithological similarities of this tectonostratigraphic units would support a common drift history. However, the Barrandian palaeomagnetic data indicate a major tectonic discontinuity in the form of major rotations between Bohemian Massif and the remaining tectonostratigraphic units of the ATA, demonstrating that they did not form a coherent microplate.

## LATE SILURIAN (FIG. 1-5)

Laurentia remained at its equatorial position. With the collision of Baltica and Avalonia the Tornquist Sea was closed and final closure of the Iapetus Ocean between Baltica and Laurentia occured in Siluro-Devonian times, after which Baltica and Laurentia remained in equatorial palaeolatitudes until the end of the Palaeozoic era [*McKerrow et al.*, 1991] [*Torsvik et al.*, 1996] [*Van Staal et al.*, 1998].

Gondwana continued moving further northwards. Siluro-Devonian palaeomagnetic data from the Tepla-Barrandian and the Armorican Massif [*Tait et al.*, 1994a] [*Tait*, 1999] point to palaeolatitudes of between 20° and 30°S, thus showing a continuation of the movement towards the southern margin of Baltica and Avalonia. This is in good agreement with geological evidence for a gradual closure of the Rheic Ocean which separated the ATA and Avalonia in Ordovician and Silurian times [*Sommermann et al.*, 1992] [*Franke*, 2000]. The declination values between the Bohemian and Armorican data differ significantly, demonstrating still major tectonic discontinuity within the ATA. No reliable data are as yet available for the Iberian Massif, but palaeomagnetic data for the Siluro-Devonian of northeastern Spain, the Catalan terrane, indicate palaeolatitudes of 30°S [*Tait et al.*, 2000b]. While a correlation of the basement correlation between this region and the Iberian Massif is not clear, faunal evidence suggest that the Catalan terrane may have formed an independent tectonic unit in Palaeozoic times [*Robardet and Gutierrez Marco*, 1990]. However, the similarity of latest Silurian brachiopods and trilobite faunas of Iberia and Armorica and ostracode faunas of Bohemia and the Armorican Massif [*Kriz and Paris*, 1982] do not allow for any significant oceanic separation between these terranes of the ATA.

Therefore, the faunal connections support the interpretation based on palaeomagnetic studies that all these terranes of the ATA, including the Catalan terrane, had similar drift histories throughout Palaeozoic times and in Late Silurian they were located in the latitudinal zone of 20 –30°S, near to the southern Baltica-Avalonia margin.


Figure 1-5: Late Silurian palaeogeographic reconstruction based on palaeomagnetic data; for ATA see text and [*Tait et al.*, 1994a] [*Tait*, 1999] [*Tait et al.*, 2000b]; for Avalonia see data compilation of [*Torsvik et al.*, 1993]; for Baltica see review by [*Torsvik et al.*, 1992] and data compilation of [*Smethurst et al.*, 1998a]; for Laurentia see [*MacNiocaill and Smethurst*, 1994]; for Siberia see review by [*Smethurst et al.*, 1998b]; for Gondwana Late Silurian palaeopole is interpolated from Late Ordovician and Late Devonian [*Hurley and Van der Voo*, 1987] palaeopoles.

## MID DEVONIAN (FIG. 1-6)

The progressive collision between the ATA and the northern continents, as Laurentia and Avalonia which had consolidated during the Caledonian Orogeny, and closure of the Rheic ocean occurred up to the Mid-Devonian, with final consolidation in the Lower-Carboniferous [*Tait et al.*, 1997] [*Franke et al.*, 1995a]. According to available palaeomagnetic data for the ATA [*Bachtadse et al.*, 1983] [*Zwing and Bachtadse*, 2000] the independent units of the ATA were situated at slightly different latitudes but continously approached the northern continents. Based on palaeomagnetic data for Gondwana the palaeo-southpole lies in Central Africa in late Devonian times [*Hurley and Van der Voo*, 1987], thus indicating the ongoing northward drift of Gondwana. It also demonstrates a still major separation of Gondwana from Laurussia.

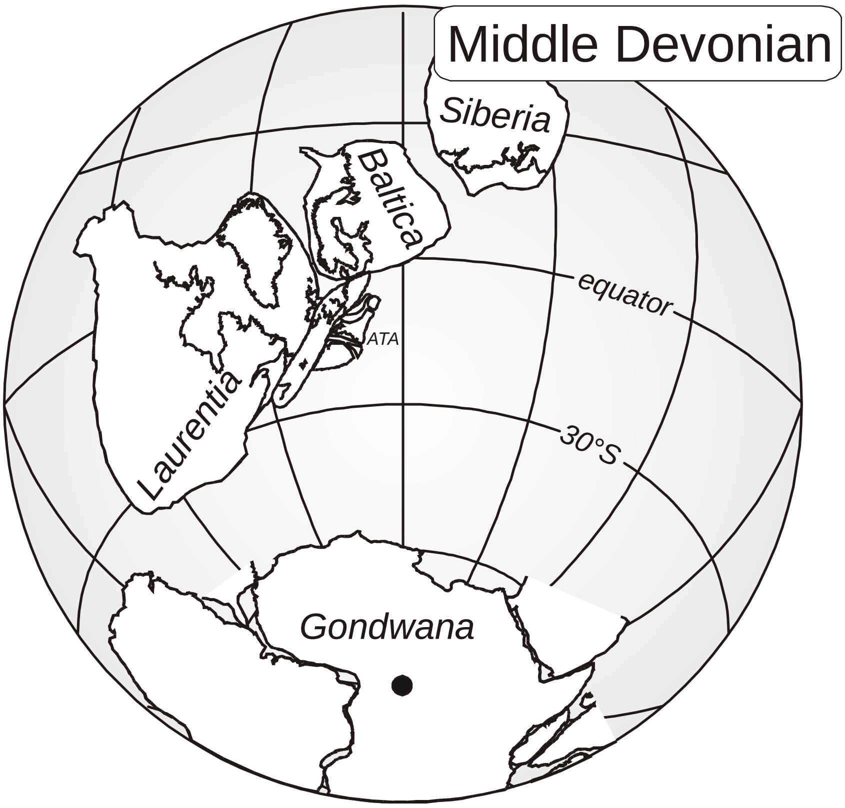



Figure 1-6: Mid Devonian palaeogeographic reconstruction based on palaeomagnetic data; for ATA see text and [*Bachtadse et al.*, 1983] [*Zwing and Bachtadse*, 2000]; for Avalonia see data compilation of [*Torsvik et al.*, 1993]; for Baltica see review by [*Torsvik et al.*, 1992] and data compilation of [*Smethurst et al.*, 1998a]; for Laurentia see [*MacNiocaill and Smethurst*, 1994]; for Siberia see review by [*Smethurst et al.*, 1998b]; for Gondwana Late Silurian palaeopole is interpolated from Late Ordovician and Late Devonian [*Hurley and Van der Voo*, 1987] palaeopoles.

## CARBONIFEROUS (FIG. 1-7)

The Carboniferous period witnessed the final approach of Gondwana and formation of Pangaea, with significant internal (intra-continental) deformation of the Variscan foldbelt and typical synorogenic sedimentation [*Franke and Engel*, 1986]. In relation to formation of Pangaea, a widespread Permo-Carboniferous remagnetisation event is ubiquitous throughout the Variscan fold belt. The evident similarity between several obtained palaeopoles from central Europe with palaeopoles from Baltica and Laurentia is ascribed to the widespread Permo-Carboniferous remagnetisation event. The systematic pattern of Carboniferous and Devonian declination data are indicative of oroclinal bending [*Bachtadse et al.*, 1983] [*Tait et al.*, 1996] have been identified as being directly related to the change in general strike of the orogenic belt. This also is a indication for indentation and profound deformation during the final stages of the collision between Gondwana and the northern continents.

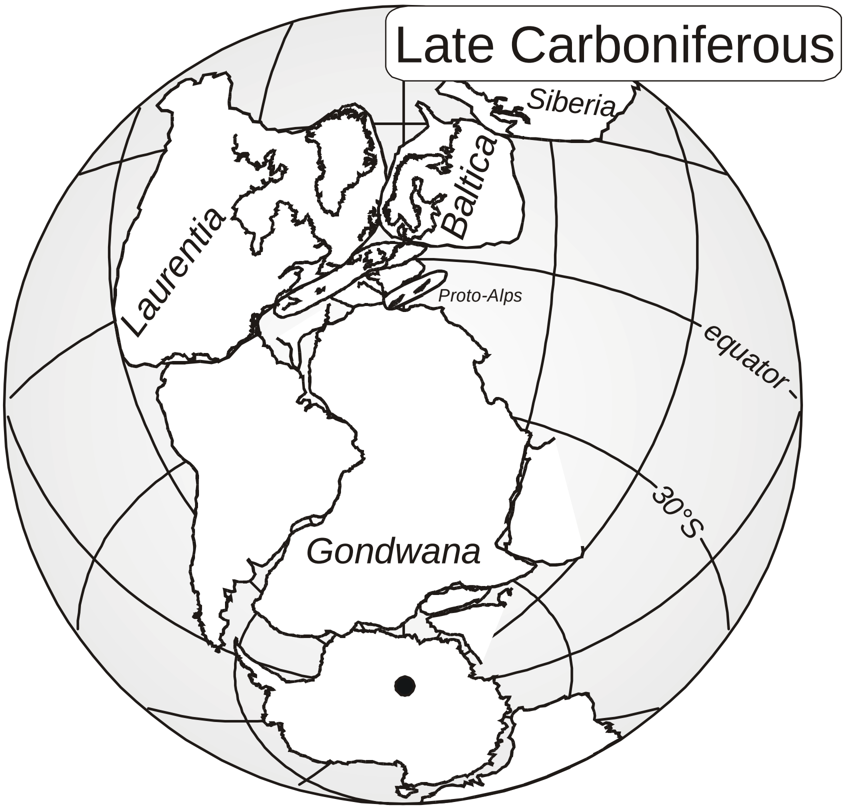



Figure 1-7: Carboniferous palaeogeographic reconstruction based on palaeomagnetic data; for ATA see text and [*Van der Voo*, 1993]; for Avalonia see data compilation of [*Torsvik et al.*, 1993]; for Baltica see review by [*Torsvik et al.*, 1992] and data compilation of [*Smethurst et al.*, 1998a]; for Laurentia see [*MacNiocaill and Smethurst*, 1994]; for Siberia see review by [*Smethurst et al.*, 1998b]; for Gondwana see [*Daly and Irving*, 1983]

## Chapter 1.3 Aim of this work

Despite recent geological, palaeobiogeographical and palaeomagnetic advances in our understanding of the geodynamic history of Palaeozoic palaeogeography and formation the Variscan and Alpine Orogenic Belts in Central Europe, many details regarding the palaeolatitudinal drift histories of suspect microplates as well as the tectonic affinities between different terranes are lacking. In order to solve this problem and to determine the palaeogeographic setting in more detail palaeomagnetic studies were undertaken in a variety of several palaeozoic sequences in central and southern Europe. In Figure 1-8 the tectonostratigraphic units building up the European Variscan and Alpine fold belt are shown.

The presence of Cadomian basement in all these tectonostratigraphic units (Fig. 1-8) shows that they are all Gondwana derived. So far, however it is not clear how long many of these units remained adjacent at the northern margin of Gondwana or if they had a possibly independent drift.

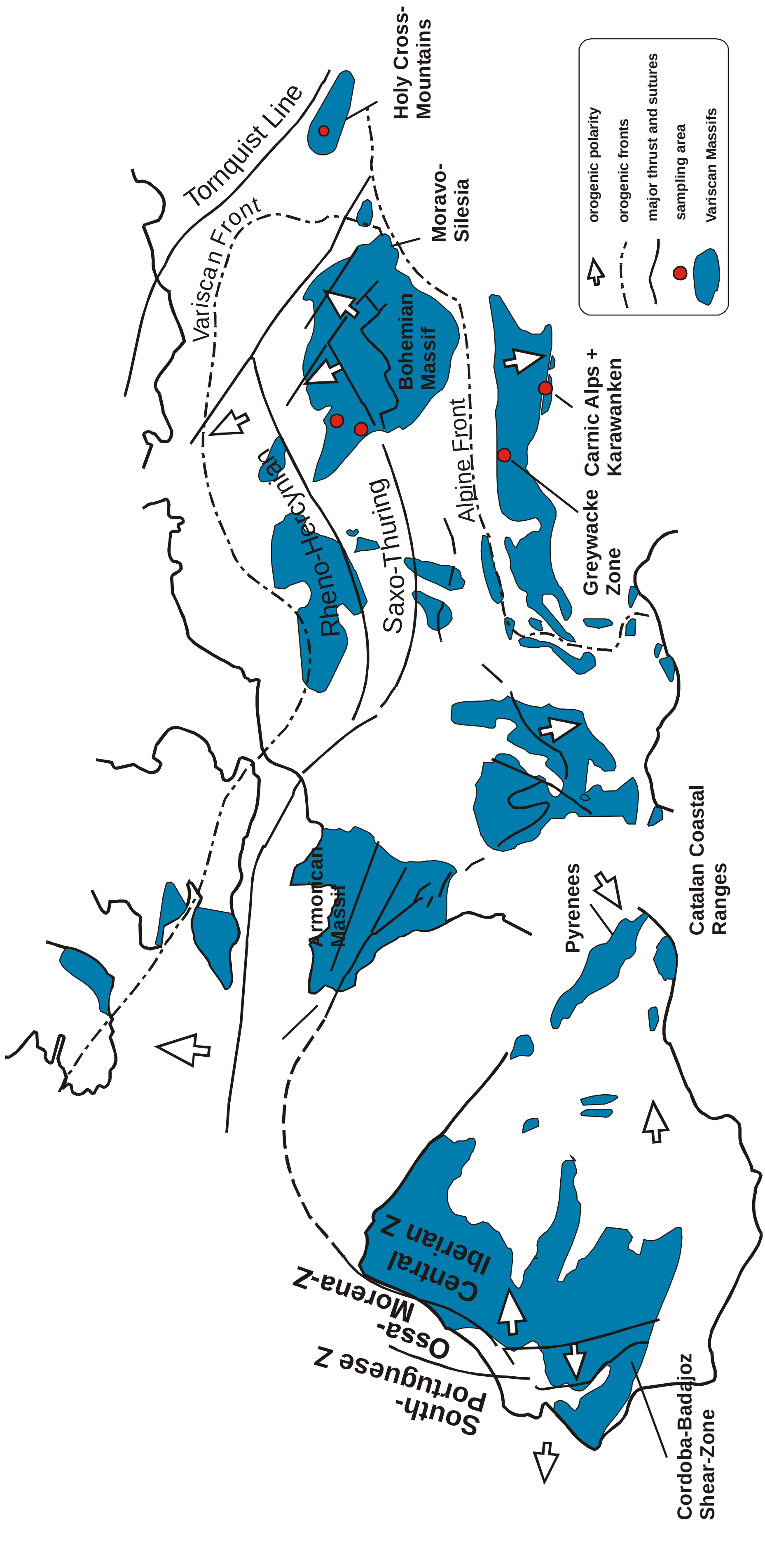



Figure 1-8: Structural map of the European Variscan fold belt after [*Franke et al.*, 1995a]. Filled dots represent the sampling areas discussed in this thesis.

The ATA comprises an assemblage of terranes or microblocks, with a major tectonic discontinuity between the Bohemian Massif and the Armorican and Iberian Massif. Especially for the Saxothuringian basin there is no palaeomagnetic evidence whether it was at similar palaeolatitudes as the ATA, and therefore in tectonic coherence with the Bohemian Massif. Thus new palaeomagnetic studies have been undertaken in the Saxothuringian unit (see Chapter 3).

The situation of the Palaeozoic basement of the Alpine realm, is complex and the palaeogeography of the various terranes remains unclear due to strong Alpine overprinting and deformation. Facies studies from the Pyrenees, Montagne Noire and the Northern Greywacke Zone suggest that from Ordovician until at least Mid-Devonian times, deposition took place in a passive margin environment, with a stable source area and a deepening basin [*Echtler*, 1990] [*Matte*, 1991] [*Heinisch et al.*, 1987]. In contrast, the data from the Carnic Alps and Karawanken show greater variations in lithology and the palaeoclimatic indicators point to a independent drift history during Palaeozoic times [*Schönlaub*, 1992]. However, there are similarities in their general faunal and lithofacial evolution and significant differences to Gondwana and the Bohemian assemblages. The palaeogeographic affinities of the Proto-Alpine units, juxtaposed in the Eastern and Southern Alps, in the Palaeozoic, remains unclear. Did they continue to be an integral part of Gondwana, did they form part of the Armorican terrane assemblage or must they be regarded as an independent unit? All these are open questions. In order to resolve these problems, palaeomagnetic studies in several palaeozoic sequences of units of the Eastern and Southern Alps has been carried out (see Chapter 4).

The structural and geodynamic history of the Malopolska Massif (Holy Cross Mountains – Poland), situated within close proximity of the Trans-European Suture Zone (TESZ) is still a matter of debate. Sedimentary provenance studies in combination with K-Ar ages of detrital muscovites indicate that Malopolska formed part of Baltica since Cambrian times [*Belka et al.*, 2000]. Existing palaeomagnetic data, however, are ambiguous. They indicate either close proximity and coherence to Baltica since Silurian times [*Nawrocki*, 2000] or a significant post Devonian dextral strike-slip displacement of Malopolska with respect to Baltica along the TESZ [*Lewandowski*, 1995]. In order to address this problem a detailed palaeomagnetic study of middle to upper Ordovician carbonate rocks from the Malopolska Massif has been undertaken (see Chapter 5).

# Chapter 1.4 Geological Setting of the selected Sampling Areas

The following sections give a short overview of the geological setting of the several study areas and sampling regions, for which palaeomagnetic results are discussed in this thesis.

## 1.4.1 SAXOTHURINGIAN BASIN

The Saxothuringian Basin of the Variscan Belt in central Europe (Fig. 1-9) is a north-easterly trending basin, bounded to the NW by the Mid-German Crystalline Rise (MGCR), and to the south-east by the north-west margin of the Barrandian. This latter boundary is marked by a major fault zone which is largely covered by Permo-Carboniferous deposits [*Matte et al.*, 1990]. Palaeobiogeographic indicators are rather sparse for the Saxothuringian, and the first palaeogeographically indicative fossils (ostracods) for Palaeozoic times are from the late Silurian Ockerkalk limestones. These show strong similarities to Barrandia [*Hansch*, 1995], thus indicating proximity of Barrandia, Saxothuringia and Armorica at least in late Silurian times [*Kriz and Paris*, 1982].

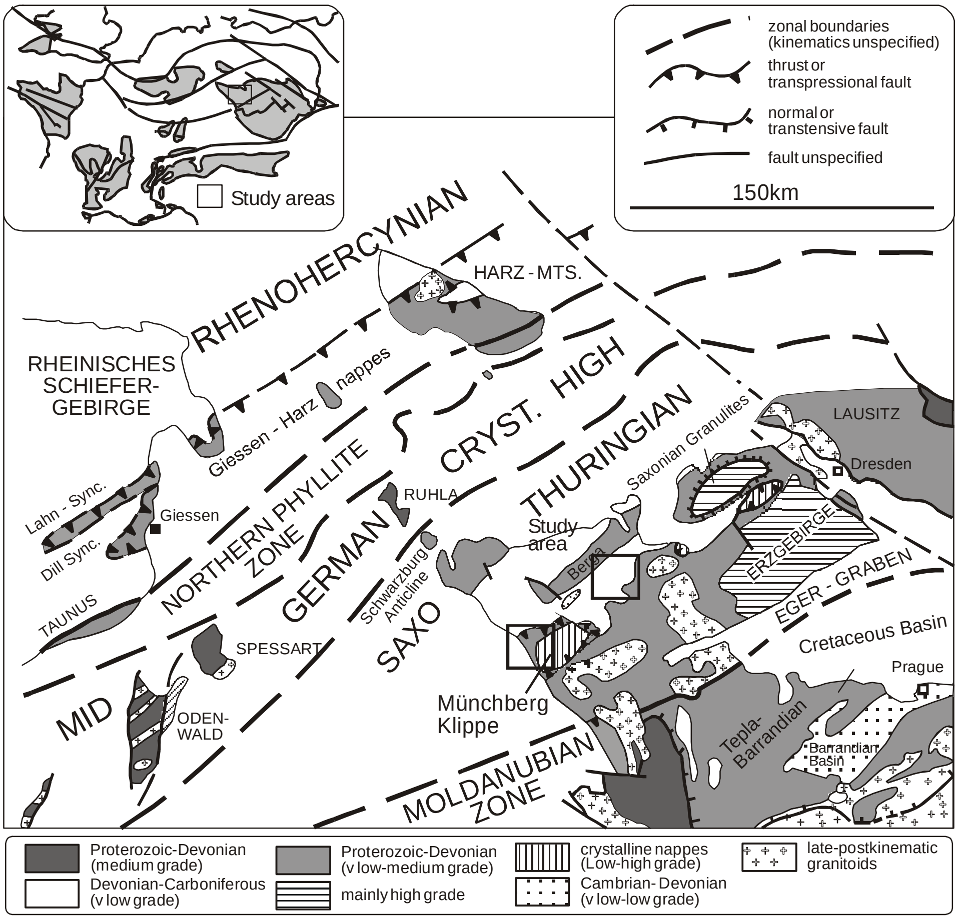



Figure 1-9: Simplified structural map of the main structural and lithological units within the Variscan fold belt of central Europe

The Saxothuringian Zone (Fig. 1-9) is thought to comprise at least two Early Palaeozoic palaeogeographic units, the Saxothuringian Terrane, which is considered part of the Armorican terrane assemblage, and the Erzgebirge which shows more affinities to Avalonia. Differentiation between these structural units is based on inherited zircon populations and provenance analyses [*Kröner et al.*, 1995] [*Kröner et al.*, 2000] [*Friedl et al.*, 2000]. The MGCR (Fig. 1-9) is generally thought to have formed the leading edge of the Saxothuringian terrane during its post-Ordovician evolution [*Reischmann et al.*, 2001] [*Anthes and Reischmann*, 2001], and was separated from the Rhenohercynian Zone of the Variscan belt (the southern margin of Avalonia) by the Rheic Ocean. The exact location of this suture zone is not clear, but is thought to be camouflaged within the Northern Phyllite Zone (Fig. 1-9, [*Franke and Onken*, 1995]). Episodic closure of the Rheic Ocean is documented by Silurian and Devonian calc-alkaline volcanism in the southern part of the Northern Phyllite Zone and the northern part of the MGCR [*Sommermann et al.*, 1992] [*Reischmann and Anthes*, 1996] [*Anthes and Reischmann*, 2001]. Timing of the final closure of the Rheic Ocean is not clear from geological evidence, but the presence of Silurian to Pragian age carbonate blocks, with Bohemian faunas in the Rhenohercynian suggest that by Early Devonian times these Armorican rocks were close to the southeast margin of Avalonia [*Franke and Onken*, 1995].

The Saxothuringian basin represents a Cambro-Ordovician rift basin, which developed on Cadomian crust [*Linnemann et al.*, 2000] [*Franke et al.*, 1995a]. Evidence for MORB-type basalts and ophiolites which may represent fragments of a Saxothuringian ocean floor, are now preserved as eclogites along the steeply dipping shear zone, or suture, separating this zone from the Moldanubian and Barrandian to the south-east [*Franke et al.*, 1995a] [*Matte*, 1991]. *Franke et al.* [1995b] used these eclogites to interpret the STB as a failed rift between North Armorica (basement of the Saxothuringian and the MGCR) and South Armorica (Tepla-Barrandian). Subduction of the Saxothuringian basement under the Tepla-Barrandian was completed by the late Devonian, as evidenced by Bohemian derived Famennian flysch deposits in the southern part of the Saxothuringian. Thus any ocean separating the Tepla-Barrandian source area from the Saxothuringian passive margin must have closed before the Famennian [*Franke and Engel*, 1986].

The basement rocks of the STB have recently been demonstrated to be Cadomian in origin, with identification of the Cadomian unconformity in two drill cores [*Linnemann and Buschmann*, 1995] [*Linnemann et al.*, 2000]. Two facies associations can be identified in the Saxothuringian Basin and were termed the Thuringian facies and the Bavarian facies by *Wurm* [1925]. As described by *Falk et al*, [1995], the Bavarian facies consists largely of hemipelagic and pelagic sediments which have subsequently been thrust over the (para)autochthonous sequences of the Thuringian facies. The Bavarian facies is exposed predominantly as isolated klippes and nappes of high grade metamorphic rocks, i.e. in present day in the Franconian forest and the surrounding areas of the Münchberg Klippe [*Franke*, 2000]. In the Kupferberg region, however, the Bavarian facies rocks are of very low metamorphic grade. In this area a subaqueous pyroclastic succession of Early Palaeozoic age from the Randschiefer Series and Vogtendorf Beds, is exposed near Kupferberg in Bavaria (Figs1-9 and 3-1). These rocks, which belong to the Bavarian facies of the

Saxothuringian Zone, are thought to be bi-modal in character, and have been described in detail by [*Martin et al.*, 1998]. The mean age of these radiometrically dated rocks obtained by single zircon evaporation technique is 478.2±1.8 Ma [*Schätz et al.*, 2002]. The geodynamic and palaeogeographic setting under which they were deposited can only be loosely inferred. The Randschiefer formation of the Bavarian facies is overlain by the Ashgillian Döbra sandstone [*Sannemann*, 1955], which can also be found only in the Bavarien facies (Fig. 1-10).

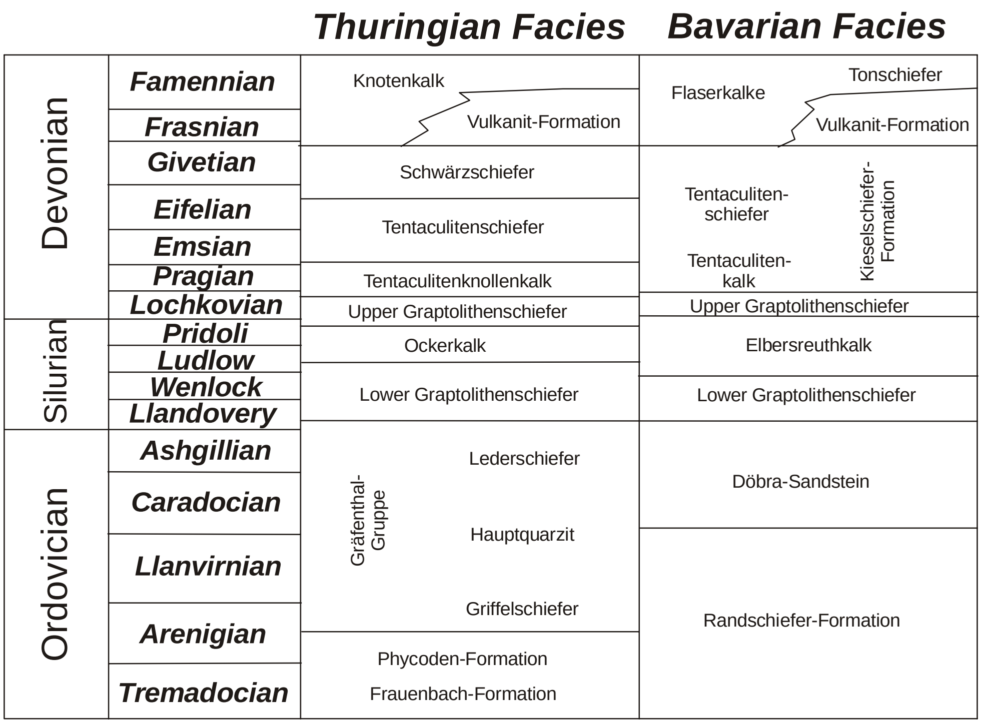



Figure 1-10: Stratigraphic column of the sampled sections of the Thuringian and Bavarian facies of the Saxothuringian basin

In contrast, the Thuringian facies rocks, exposed in the present day north-western part of Saxothuringia, are in general of very low metamorphic grade, and comprise neritic to hemipelagic shelf realms, i.e., the shallower water equivalents of the allochthonous units and represent the passive margin of a continental unit to the northwest (Fig. 1-10). Deformation of the STB has occurred during the Variscan uplift in the Dinant. Cambrian and Lower Ordovician sediments of the STB comprise a thick pile of quarzites and shale with hummocky cross bedding with trace fossils indicating a high energy, shallow water environment [*Luetzner and Mann*, 1988]. These neritic sediments are overlain by mid to Upper Ordovician quarzites, oolitic ironstone beds, and occasional limestone horizonts. The Asghillian sediments are represented by the Lederschiefer, a shaley shale-sandstone sequence dominated by glacio-marine tillites with occasional dropstones and faceted, striated clasts indicating a glacial origin [*Falk et al.*, 1995]. These sediments are the stratigraphic equivalent to the diamictites of the Barrandian Basin and the Iberian Massif. In general, the sedimentary record of the STB demonstrates a gradual deepening from the Cambro-Ordovician neritic clastics to Silurian age black graptolite shales, which are ubiquitous in the Variscan Belt. Upper Silurian deposits demonstrate a brief interruption of the anoxic environment with the deposition of Pridolian age pelagic limestones (Ockerkalk), which is likewise widespread in Europe [*Jaeger*, 1988]. Devonian sediments comprise black shales with intercalations of condensed, nodular, pelagic limestones (Tentakuliten-Knollenkalk) of Gedinnian through Emsian age. These are overlain by black shales until Late Frasnian times when a conspicuous sequence of sedimentary and volcanic rocks are interpreted as recording a brief period of renewed rifting [*Falk et al.*, 1995].

The subject of a part of this study are the Early Palaeozoic beds of the Bavarian facies which crop out along the northwestern margin of the Münchberg Gneiss massif (Figs. 1-9, 1-10 and 3-1 and see Chapter 3.1) and the Upper Silurian Ockerkalk of the Thuringian facies of the Vogtländischen Synclinorium (Figs. 1-9, 1-10 and 3-6 and see Chapter 3.2).

### 1.4.2 ALPINE REALM

The Palaeozoic palaeogeography and tectonic evolution of the Alpine realm is extremly complex due to the poly-metamorphic and poly-deformational events suffered by these rocks. From the geological record the Palaeozoic basement rocks may comprise an amalgamation of different terranes, the Proto-Alps as a whole, with distinction being made between the external Alps (the more northerly and westerly units in present-day coordinates) and the internal Alps which are the more southerly units, i.e. the Austroalpine Nappe complex and the Southern Alps. Nevertheless, a number of different Palaeozoic tectonostratigraphic terranes can be recognized, particularly in the Eastern Alps. These Palaeozoic basement units are widely exposed in the Southalpine, Austroalpine and Penninic unit of the Eastern Alps (Fig. 1-11). Metamorphic and deformational signatures vary greatly throughout the various basement units of the Alpine realm. The different characteristics of the pre-Alpine metamorphic evolution have been interpreted to result from stepwise accretion of these terranes onto the active Laurussian margin from Devonian to Permian times [*Frisch and Neubauer*, 1989] [*Neubauer et al.*, 1999] [*Schönlaub*, 1992]. The Austroalpine/Southalpine basement rocks with conformable Early Palaeozoic to Late Carboniferous sedimentary sequences are termed the Noric-Bosnian terrane (NBT) by *Neubauer and Raumer*, [1993]. No pre-Carboniferous orogenic events have been observed in the NBT where there was continuous sedimentation from Ordovician to Late Carboniferous times in a passive margin-type environment [*Schönlaub*, 1992] [*Stampfli*, 1996] [*Frey et al.*, 1999] [*Neubauer et al.*, 1999]. These units generally show a very low-grade Late Variscan metamorphic overprint [*Neubauer et al.*, 1999], however the influence and intensity of the Alpine orogeny is not irrelevant for palaeomagnetic studies, and is discussed in more detail in chapter 4.

The Cambro-Ordovician marks a phase of crustal extension and rifting along the northern margin of Gondwana and is documented in widespread sedimentary and volcanic sequences within the European Variscan fold belt [*Franke*, 1992]. Such basin-development can also be recognized in the Alpine region where these rocks form part of the basement units of the Alps [*Neubauer and Raumer*, 1993] [*v. Raumer*, 1998].

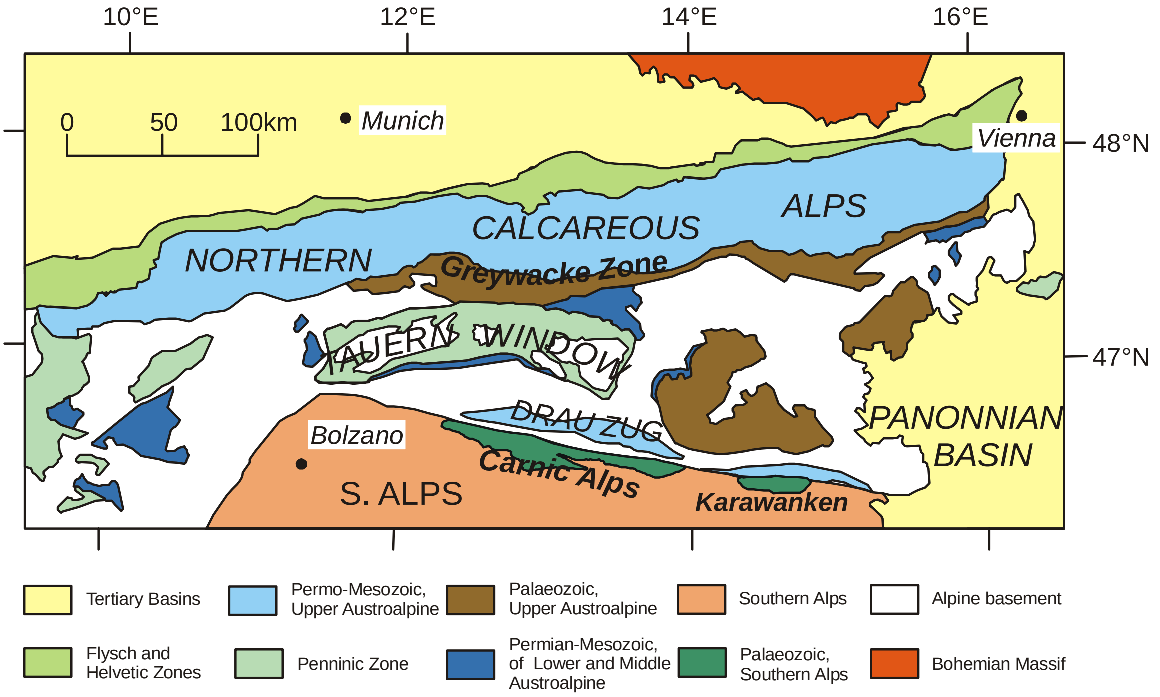



Figure 1-11: Simplified structural map of the pre-Triassic basement units in the Eastern Alps

In the Eastern Alps (Fig. 1-11) metamorphic and plutonic Austroalpine and Penninic basement units, exhibiting a complex Phanerozoic history, are well exposed. The Austroalpine units are classically divided into three division, the Lower, Middle and Upper Austroalpine units relating to Late Mesozoic nappe stacking and being indicative of their relative structural position. Whilst the Lower and Middle units tend to have fairly high metamorphic grades, non- or very low grade-metamorphic Palaeozoic sedimentary sequences are exposed within in the Upper Austroalpine-Nappe Complex (in the Western and Eastern Greywacke Zone, the Gurktal and Graz Nappe Complexes [*Tollmann*, 1963]). The Southern Alpine units (Carnic and Karawanken Alps) are exposed along the Austrian-Italian border along the Periadriatic fault, and comprise similarly low to very low grade metamorphic successions. *Dallmeyer and Neubauer*, [1994] obtained 650-600 Ma $^{40}Ar/^{39}Ar$ apparent ages for detrital white micas within Ordovician sandstones of the Carnic Alps. Similar $^{40}Ar/^{39}Ar$ apparent ages were also determined for detrital white micas in Ordovician sandstones from the Gurktal Nappe Complex (570-550 Ma) [*Antonitsch et al.*, 1994] and from the Eastern Greywacke zone (600-560 Ma) [*Handler et al.*, 1997], both of the Upper Austroalpine units. First geochronological data from the Western Greywacke Zone give similar age distributions (616-568 Ma) [*Panwitz et al.*, 2000] from locations close to the palaeomagnetic sites described in this thesis. These data indicate Cadomian source rocks, and also demonstrate the low thermal effects (i.e., below the closure temperature for white mica in the $^{40}Ar/^{39}Ar$ isotopic-system) of the Variscan and Alpine deformation in these regions.

The Ordovician to Silurian in the Proto-Alps is characterized by tectono-thermal and volcanic activity and simultaneous clastic sedimentation [*Schönlaub*, 1988] [*Schönlaub and Heinisch*, 1993] which is compatible with crustal-thinning in a back arc tectonic setting [*Frisch and Neubauer*, 1989]. In the Late Silurian to Devonian, however, widespread volcano-sedimentary successions are mostly of intraplate geochemistry and character [*Loeschke*, 1989] [*Schlaegel*, 1988] [*Fritz and Neubauer*, 1988] [*Schlaegel-Blaut*, 1990] [*Loeschke and Heinisch*, 1993] indicating differentially subsiding mobile basins affected by extensional tectonics. Devonian sediments are characterized by variable interfingering facies which range from condensed pelagic cephalopod limestones, deep sea shales and coastal sediments to carbonate buildups. From the mid Devonian on, reef development occurred in several regions ([*Heinisch*, 1988] [*Schönlaub and Heinisch*, 1993] and references therein). This sedimentation is interpreted as having developed in a passive margin environment.

In order to constrain the palaeogeographic position of the Proto-Alps in the Palaeozoic samples were taken from biostratigraphic well dated sections of the Western Greywacke Zone and the Carnic Alps (Figs. 1-11 and 1-12).

## WESTERN GREYWACKE ZONE

The Kitzbühel Alps belong to the Western part of the Northern Greywacke Zone which forms the Palaeozoic basement of the unconformably overlying Mesozoic sediments of the Northern Calcareous Alps. Both the Northern Greywacke Zone and its transgressive cover form part of the allochthonous Upper Austroalpine-Nappe complex [*Tollmann*, 1963] in the Eastern Alps (Fig. 1-11). Detailed mapping of the Kitzbühel-Saalbach area (Fig. 4-2) demonstrate the presence of two distinct facies associations, the Wildseeloder Unit and the Glemmtal Unit [*Heinisch*, 1986] [*Heinisch*, 1988] [*Heinisch et al.*, 1987] [*Schlaegel*, 1988]. These units are separated by the complex Hochhörndler imbricate shear zone (Fig. 4-2) [*Heinisch*, 1988].

The Glemmtal Unit (Fig. 1-12) represents a marginal basin dominated by a thick sequence (up to 2000m) of Ordovician to Devonian siliciclastic rocks, predominantly sandstones and siltstones. Sedimentary structures give evidence for turbiditic deposition with submarine fan systems; proximal channel facies are preserved as well as distal facies, displaying various types of Bouma sequences [*Heinisch*, 1988]. The mineralogical composition of the sandstones indicates a continental source area without any influence of an active continental margin. Palynological data place the onset of clastic sedimentation in the earliest Ordovician [*Reitz and Höll*, 1989] [*Reitz and Höll*, 1991]. The turbidite basin persisted through Silurian into Devonian times. From the Middle Devonian onward, a generally coarsening upward trend is observed. The onset of orogenic activity is marked by the deposition of flysch sediments in the Carboniferous [*Heinisch et al.*, 1987]. The monotonous pre-Carboniferous siliciclastic sequences are intercalated by the Klingler Kar Formation (KK), which is up to 5m thick and comprises condensed carbonates, marls, black shales and lydites. Conodont biostratigraphy indicates an Upper Silurian (Ludlovian/Pridolian) to uppermost Lower Devonian (Emsian) age for the Klingler Kar [*Heinisch et al.*, 1987]. They are overlain by Devonian age basalts which, in their lower part, are intercalated with fossiliferous carbonate beds. Volcanism in this region started in the Emsian and

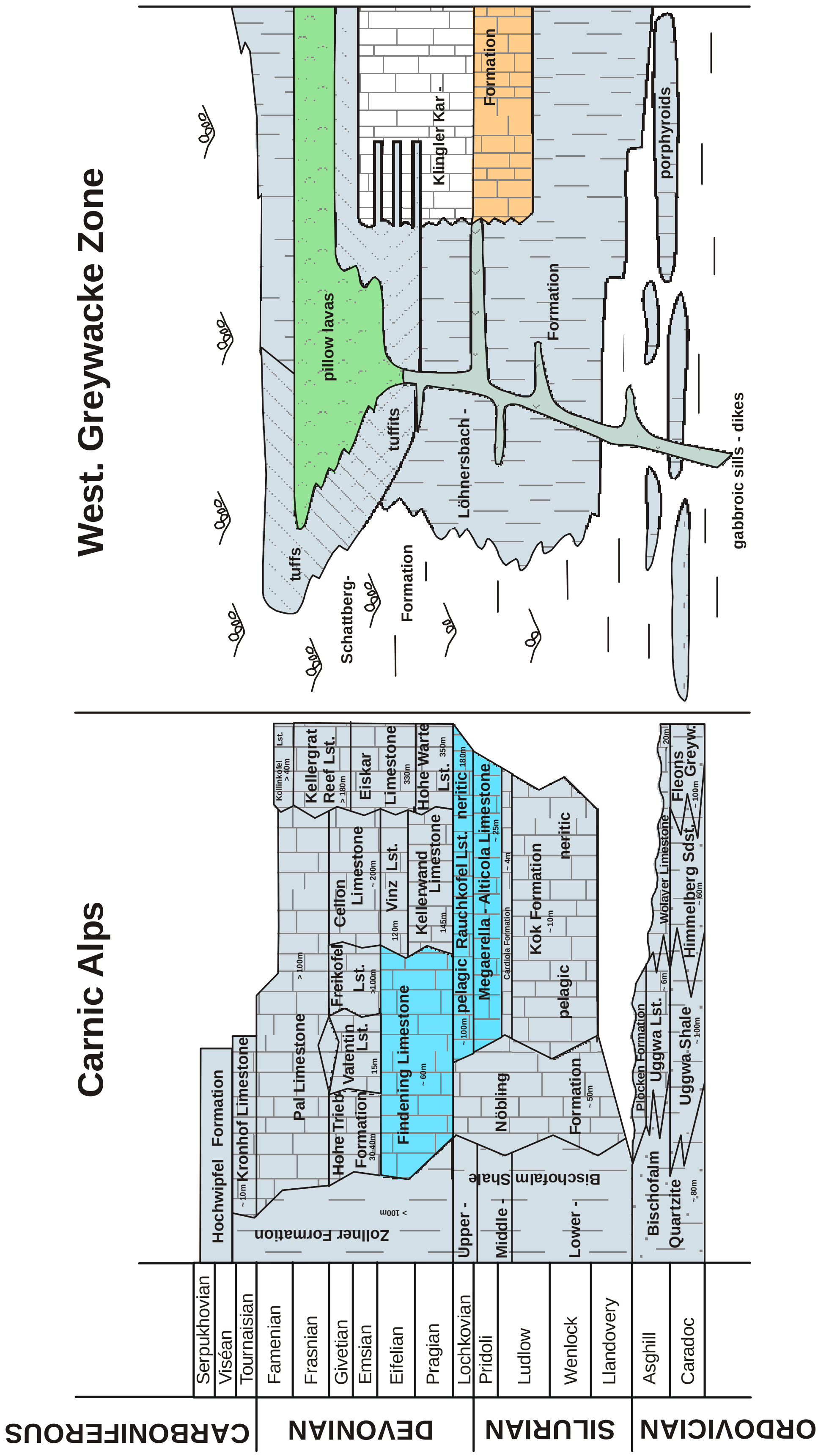



Figure 1-12: Stratigraphic columne the sampled sectionof the Carnic Alps near Kötschach-Mauthen after [*Schönlaub et al.*, 1992] and of the Glemmtal Unit of the Western Greywacke Zone at the Kitzbüheler Alps after [*Heinisch*, 1988].

persisted until at least Mid Devonian times. The basaltic seamounts produced pillow lavas, sheet flows, sills and pyroclastic layers up to 300m thick [*Schlaegel-Blaut*, 1990]. Based on geochemical analyses, the basaltic volcanism is of an intraplate type (transitional basalts to alkali basalts), and volcanological evidence shows that that these seamounts were active in a shallow marine environment adjacent to a passive continental margin [*Schlaegel-Blaut*, 1990] [*Loeschke and Heinisch*, 1993]. The continuous sedimentation of clastic, continent-derived detritus, including boulders of gneisses, mica schists and granites, indicates a palaeogeographic link to a continental landmass which was uplifted and eroded during the whole time span of sedimentation.

Within the Wildseeloder Unit of the Kitzbühler Alps, Upper Ordovician porphyry rocks are overlain by Silurian sediments. The sedimentary rocks range from black shales to cherts, siliceous pelagic limestones, condensed cephalopod limestones and even to dolomitic rocks. The Late Silurian rocks show the onset of carbonate platform development which persisted until the Early Upper Devonian. It comprises shallow water lagoonal dolomites with local reef development containing pelagic limestones of Frasnian age [*Mostler*, 1970] [*Heinisch*, 1988] [*Schönlaub and Heinisch*, 1993].

CARNIC ALPS

The Palaeozoic sequence of the Carnic Alps represents a strongly compressed thrust sheet complex with internal fractures and imbricate thrust tectonics. At least three facies zones can be recognized which comprise the variations of a sedimentary basin. The Figure 1-12 summarizes the stratigraphy and facies relationship of various rocks of the Carnic Alps. With minor modification this scheme is also valid for the Karawanken Alps [*Schönlaub*, 1992]. The oldest fossiliferous rocks are Caradocian in age and comprise thick acid volcanics and volcaniclastics of the Fleons Formation which laterally and vertically grade into the Ugwa shale and Himmelberg Sandstone which contain detrital muscovites of Cadomian age [*Dallmeyer and Neubauer*, 1994]. They are succeeded by bioclastic limestones. The global regression during Late Ashgillian is documented by arenaceous limestones of the Plöcken Formation with erosion and local non deposition [*Schönlaub*, 1988] [*Schönlaub*, 1992]. Thus the Late Ordovician generally is overlain disconformably by Silurian Strata.

The Silurian transgression began at the base of Llandovery in the graptolite zone of the Bischofsalm facies. Silurian lithofacies is split up into four major facies reflecting different depths of deposition and hydraulic conditions [*Schönlaub and Heinisch*, 1993]. The typical section of the Cellon profile (shallow marine Plöcken facies) contain detailed conodont zonation established by [*Walliser*, 1964]. The shallower environment of the Wolayer facies is characterized by fossiliferous limestones with nautiloids, trilobites, brachiopods and crinoids. The Finding facies comprises interbedded black graptolite shales, marls and limestone beds. Finally the Bischofsalm facies represents graptolitic black siliceous shales, lydites and clayish shales [*Jaeger et al.*, 1975]. The Silurian sedimentation suggests a steadily subsiding basin and a transgressional regime. This tendency decreased during the Pridoli to form balanced conditions with uniform limestones. At the base of the Devonian in the Bischofsalm facies the deep-water graptolite environment was restored by the end of Lochkovian Stage. The succeeding strata of the Zollner Formation also represent a deep-water off-shore setting that lasted until the end of the Devonian/Early Carboniferous.

Again in the Devonian subsidence and mobility of the sea bottom significantly increased. This is documented in a Lower Devonian transgressional sequence including the Rauchkofel limestone which is up to 180 m thick [*Schönlaub and Heinisch*, 1993]. During Pragian and Emsian Stages the differences further increased. Within short distances a strongly varying facies pattern developed indicating a progressive but non-uniform deepening of the basin [*Kreutzer*, 1992]. It was filled with thick reef and near-reef organodetritic limestones including different intertidal lagoonal deposits of more than 1000 m thickness in the Carnic Alps. In the Carnic and Karawanken Alps reef growth started in the lower Emsian [*Schönlaub*, 1992]. In both regions the reef development ended in the Frasnian when the former shallow sea subsided and the reefs drowned and were partly eroded [*Kreutzer*, 1990]. Subsequently uniform pelagic goniatite and clymeniid limestones were deposited lasting from the Frasnian boundary to late Tournaisian Stage [*Schönlaub*, 1992]. They were named the Pal and Kronhof limestone. These wackestones contain cephalopods, trilobites, radiolariens, foraminifera, ostracodes, conodonts and fish teet [*Schönlaub and Heinisch*, 1993]. Rise of the sea-level and collapse of the carbonate basin promoted deposition of the transgressive Hochwipfel Formation, which started in the Lower Visean [*Schönlaub and Histon*, 2000]. Based on its characteristic lithology and sedimentology the Hochwipfel Formation was interpreted as a Variscan flysch sequence [*Spalletta and Venturini*, 1988]. In the Southern Alps the Variscan orogeny reached the climax between the Late Namurian and Late Westphalian stages as a result of collision and consolidation with the Central Alps in the North and the Laurussian margin [*Schönlaub and Histon*, 2000]. The Variscan rocks are overlain unconformably by Late Carboniferous sediments. According to [*Kahler*, 1983] the Late Palaeozoic cover comprises clastic and calcareous shallow marine sediments of the Latest Carboniferous Auernig Formation followed by Lower Permian shelf and shelf edge deposits.

These fossiliferous sections of the Carnic Alps and the Western Greywacke Zone yield abundant palaeobiogeographic indicators which are combined with the palaeomagnetic studies of this thesis (see chapters 4.1 and 4.2).

### 1.4.3 MALOPOLSKA MASSIF (HOLY CROSS MOUNTAINS – POLAND)

The Trans-European Suture Zone is a broad and complex NE/SW trending zone which extends from the North Sea through to the Black Sea over Europe and is the most prominent geological boundary separating the ancient Precambrian lithosphere of the East European Craton from the Variscan and Alpine mobile belts of Western Europe [*Berthelsen*, 1992; *Pharaoh et al.*, 1997]. In this rather complex tectonic zone, build up of various crustal blocks at the SW margin of the Baltic shield [*Berthelsen*, 1992] [*Guterch et al.*, 1986], a number of different Palaeozoic terranes have been identified. These terranes are defined by their contrasting lithological, stratigraphic and tectonic characteristics [*Pozaryski*, 1990] [*Pozaryski et al.*, 1992] [*Berthelsen*, 1992]. The Palaeozoic geodynamic development of the these terranes, however, and the process of accretion to the present day southern margin of Baltica remains unequivocal.

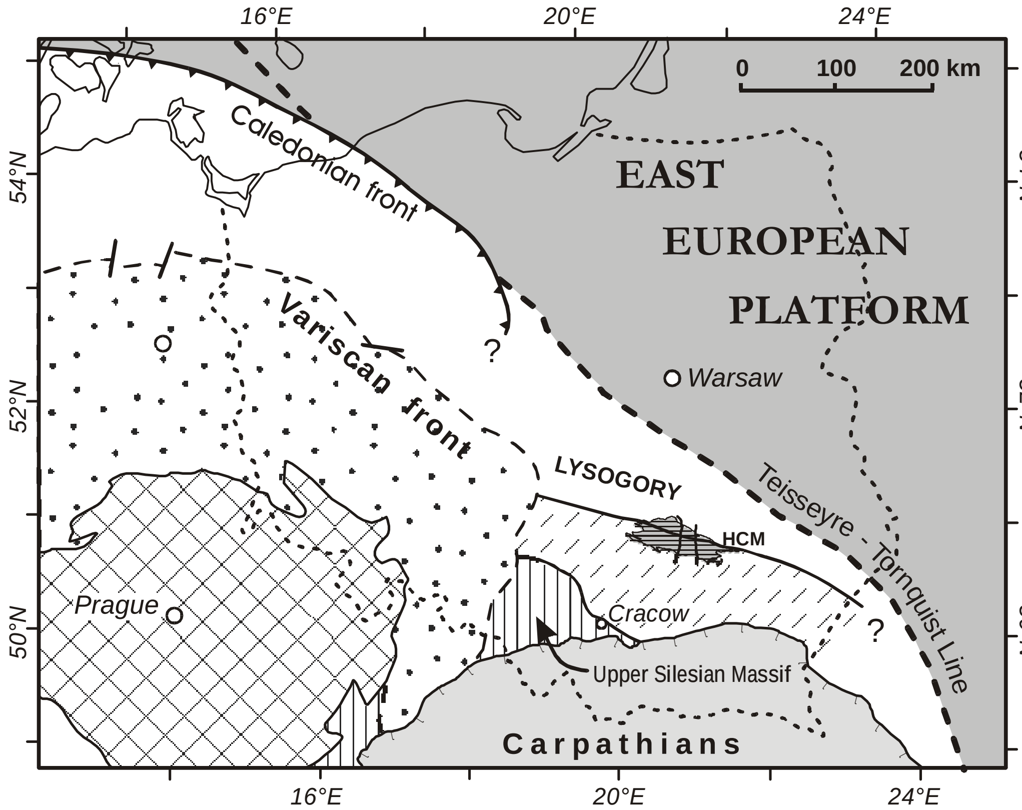



Figure 1-13: Simplified structural map of central Europe showing the crustal units of the southeastern margin of the East European Platform. Dotted line indicates the Polish border.

In Southern Poland, three tectonostratigraphic fault bounded crustal units, the Lysogory, Malopolska and Upper Silesian Massif are juxtaposed, and differentiated by their contrasting lithological and tectonic characteristics (Fig. 1-13), and are now incorporated into the Variscan and Alpine orogenic belts. Different opinions exist concerning the tectonic evolution and affinities of these terranes which now border the Baltic Shield [*Pozaryski et al.*, 1992] [*Belka et al.*, 2000].

The Holy Cross Mountains (HCM) expose Palaeozoic rocks at their core (Fig. 1-14). These sequences can be divided into two distinct tectonostratigraphic units, the Lysogory Unit and Malopolska Massif, which are separated by the NW-SE trending Holy Cross Fault (HCF). Both of these units are largely covered by Upper Proterozoic to Lower Carboniferous sedimentary rocks. The two regions differ in facies, thickness of stratigraphic successions and continuity of their stratigraphic record. Both terranes are separated from other Palaezoic blocks of central

Europe by Permian and Mesozoic systems thus lateral extensions and affinities are difficult to establish.

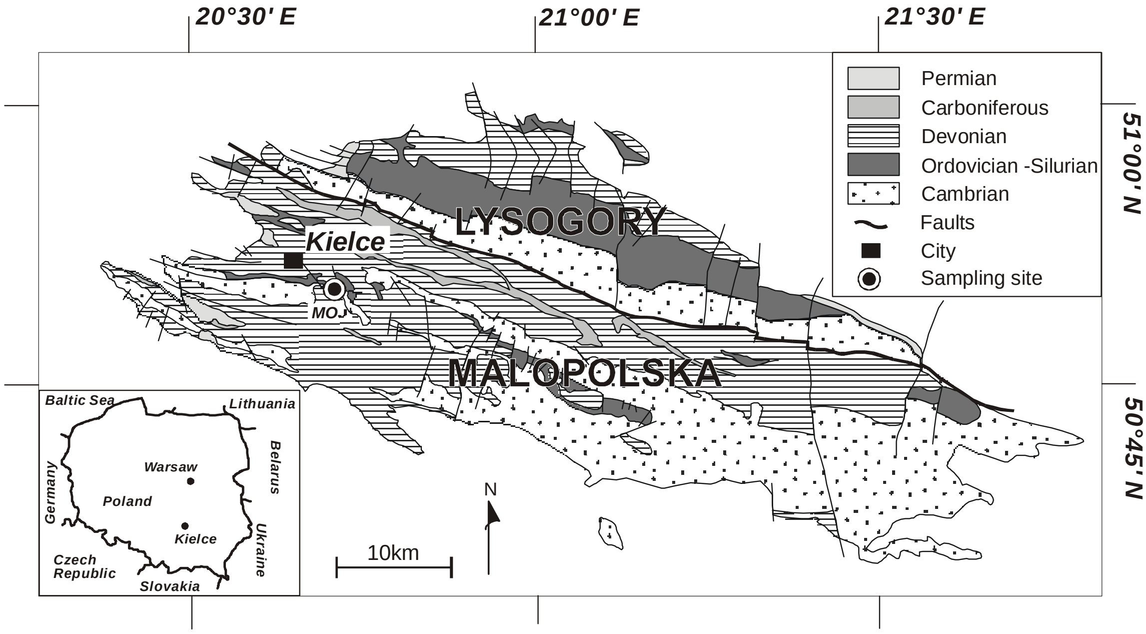



Figure 1-14: Geological sketch map of the Holy Cross Mountains, Poland. The Holy Cross Dislocation separates the Malopolska Massif in the South from the Lysogory tectonostratigraphic unit in the North. The sampled section is situated near the village of Mójcza, ca. 10 km southeast of Kielce.

Both terranes comprise clastic and carbonate sediments which were deposited from Vendian and Cambrian times [*Dadlez*, 1995] [*Bergström*, 1984] [*Vidal and Moczydlowska*, 1995] through to the Carboniferous. Those of the Lysogory unit represent a continuous uninterrupted sequence ([*Belka et al.*, 2000] and references therein), whereas those of the Malopolska unit show a distinct discontinuity between the Middle Cambrian and the Early Ordovician sequences. The Lower Ordovician rocks of the Malopolska Massif discordantly overlie the folded Precambrian to Middle Cambrian successions, and comprises a transgressive sequence, changing from offshore facies sandstones at the base to marine carbonate rocks with marls and shales on the top of the succession of Late Ordovician age. A second disconformity is seen between the Late Silurian and the Lower Devonian sediments [*Szulczewski*, 1995]. In this study, samples from the condensed mid-Late Ordovician sediments of the Malopolska Massif were collected to determine the palaeogeographic ralations based on palaeomagnetic data.

Three different formations build up the Ordovician succession in the Malopolska Massif. These are the Bukówka Sandstone Formation at the base, a light fossiliferous sandstone of Tremadocian to lower Arenigian in age, followed by the organodetrital Mójcza Limestone, latest Arenigian to latest Caradocian in age and the marly limestones and shales of the Zalesie Formation at the top, Ashgillian in age. The aim of sampling was the Mid Ordovician Mojcza limestone whose type locality is exposed in an old quarry at Mójcza near to the city Kielce (Fig. 1-14). In general the succession indicates a transgressive and shallow-marine facies [*Szulczewski*, 1995] [*Dzik*, 1978]. Sedimentation occurred under stable conditions with a rather shallow depth of reworking as shown by the sudden appearance and disappearance of species in the section [*Dzik and Pisera*, 1994].

In this study palaeomagnetic investigations were undertaken to produce data for the palaeopole position for the HCM and to solve geotectonic problems in the TESZ in Mid Ordovician times and the palaegeographic relation of the HCM particulary the Maloposlka Massif (see Chapter 5).

# Chapter 2 Sampling and Methods

## Chapter 2.1 Sampling and Laboratory procedure

For the palaeomagnetic investigations of this thesis usually samples were taken by a portable gasoline-powered drill with a diamond drilling bit. A pump was used to force cooling water through the drill bit. In total 264 sites were collected, corresponding to approximatly 2000 single core samples (Table 2-1). All samples were oriented in the field. An orientation device was placed over the in situ core, recording the geographic coordinates, the inclination and azimuth of the z-axis of the core sample with a standard magnetic compass.

Samples were collected only from sections which are well dated, usually biostratigraphically or radiometrically and for which there are good structural control. The structural parameters were readily measured at each site. Also during sampling the grade of weathering was taken into consideration to reduce the possibility of a modern chemical remagnetization. Samples were only collected from sequences which are known to be non-methamorphic or at most very low grade from Illite crystallinity, vitrinite reflectivity, CAI and others.

The field-drilled cores were cut into standard 2.2 x 2.5 cm specimens. The natural remanence magnetisation (NRM) of the standard samples was measured using a 2G-Enterprises cryogenic magnetometer in a magnetically shielded room at the Laboratory for Palaeo- and Rockmagnetism of the University of Munich in Niederlippach. Samples were demagnetized using both stepwise thermal and alternating field techniques also in the magnetically shielded room. After each heating step during the thermal demagnetisation experiments, the magnetic susceptibility was measured, using a KLY2 Kappabridge, to monitor thermochemical alteration.

Demagnetisation results were analysed using orthogonal vector plots [*Zijderveld*, 1967], and stereographic projections. Directions of linear demagnetisation trajectories, defined by at least three successive data points, were selected by eye and calculated using principal component analysis [*Kirschvink*, 1980]. Where overlapping coercivity or blocking temperature ($T_b$) spectra prevented complete separation of the single components, great circle analysis was performed using the method of [*McFadden and McElhinny*, 1988] on the planar demagnetisation trajectories to identifiy the underlying direction. This method uses an iterative procedure whereby the maximum likelihood estimate of the endpoint direction within the acceptable sector for each great circle can be determined. This is then combined with the stable endpoint directions, resulting in determination of the mean directions, which are calculated using [*Fisher*, 1953] statistics in all cases.

Table 2-1: Collected samples in segments of the Variscan fold belt of Central Europe

| Tectonostrat.-Unit | Segment | | Lithology | Period | sites |
|---|---|---|---|---|---|
| Saxothuringian Basin | Thuringian Synclinorium | | Limestone<br>Porphyroid | Late Silurian<br>Ordovician | 6<br>4 |
| | Saxonian Synclinorium | | volcanics<br>limstone<br>Shale | Late Devonian<br>Late Silurian<br>Ordovician | 4<br>10<br>5 |
| | Franconian Forest | | Volcanics<br>Silt-,sandstones<br>Ignimbrit | Devonian<br>Late Ordovician<br>Early Ordovician | 7<br>4<br>15 |
| Alpine realm | Upper Austroalpine | Greywacke Zone<br><br><br>Graz nappe<br><br>Gurktal nappe | Volcanics<br>Limestone<br>Silt-,sandstone<br>Limestones<br>Volcanics, Limest.<br>Volcanics | Mid Devonian<br>Late Silurian<br>Palaeozoic<br>Mid Devonian<br>Silurian/Devonian<br>Ordovican | 15<br>6<br>7<br>11<br>7<br>7 |
| | South Alpine | Carnic Alps | Limestone<br>Limestone<br>Limestone<br>Limestone<br>Silt-,sandstone | Carboniferous<br>Mid Devonian<br>Early Devonian<br>Silurian<br>Ordovician | 8<br>14<br>24<br>23<br>18 |
| West-Sudetes | Kłodzko Unit | | Konglomerat<br>Limestone<br>Shale, lydit | Carboniferous<br>Late Devonian<br>Silurian | 2<br>4<br>3 |
| Malopolska Massif | | | Limestones<br>Sand-,limestones<br>Volcanics, clastics<br>Sand-,limestones<br>Silt-,sandstone | Carboniferous<br>Devonian<br>Silurian<br>Ordovician<br>Cambrian | 5<br>7<br>6<br>15<br>10 |
| Lysogory | | | Silt-,sandstone<br>Silt-,sandstone<br>Silt-,sandstone | Silurian<br>Ordovician<br>Cambrian | 6<br>3<br>8 |

Various rock magnetic experiments were also carried out to characterize the carriers of magnetizations. These include isothermal remanent magnetisation (IRM) measurements and thermal demagnetisation of the saturation IRM (SIRM), determination of hysteresis properties and also the anisotropy of magnetic susceptibility (AMS).

To establish the possible influence of structural deformation on the magnetization directions, the anisotropy of magnetic susceptibility (AMS) of mainly volcanic rocks was measured. The susceptibilities of the sedimentary rocks are, in general, too low to yield accurate AMS measurements. The magnetic fabric is determined from the principal axes (K1 = maximum axis, K2 = intermediate, K3 = minimum) and the shape of AMS ellipsoids was used as a proxy for the petrofabric which allows recognition of any penetrative deformation or patterns of strain on or within the rocks [*Tarling and Hrouda*, 1993]. The magnetic fabric was characterized by applying the conventional parameters, $P_J$ (corrected anisotropy degree) and the shape parameter T after [*Jelinek*, 1981]. Comparing the in the samples identified preferred shape of the anisotropy ellipsoid with the average degree of anisotropy of recent basaltic flows which behave up to 10% [*Tarling and Hrouda*, 1993], it is possible to detect the degree of strain which could affect the direction of the magnetisation. If the average degree of anisotropy in samples is above 10 % and with a dominate oblate shape parameter it is indicative for a intensive teconic deformation and these samples normally yield no primary magnetisations. Plotting the directional data of the anisotropy $k_{max}$, $k_{int}$ and $k_{min}$ on a stereographic projection it is possible to detect the spherical orientation and the shape of the magnetic susceptibility ellipsoid. This kind of projection has proved useful in analyses of flow plane of volcanic rocks or for reconstruct the bedding plane or favoured sedimentation directions. The magnetic foliation of undeformed rocks is approximately parallel to the flow or sedimentation plane measured in the field [*Hrouda*, 1982] and it is possible to confirm the bedding plane determined in the field.

The identification of ferromagnetic minerals in a rock can help to guide the design of partial demagnetization experiments and the interpretation of results. The aim is to associate a particular component of NRM with a particular ferromagnetic mineral. This information can often help to determine whether a characteristic NRM is primary or secondary. Therefore isothermal remanent magnetisation (IRM) and thermal demagnetisation of the saturation IRM (SIRM) experiments were also carried out to determine the magneto-mineralogy of the samples by analysing their coercivity spectrum. The coercivity spectrum analysis uses the contrast in coercive force between hard (e.g. hematite or goethite) and soft (e.g. magnetite) magnetic minerals. It is common to combine the IRM acquisition experiment with a later thermal demagnetisation. SIRM decreases during thermal demagnetisation as unblocking temperatures are reached. Major decreases in IRM during thermal demagnetisation allow the estimation of Curie temperatures and with it the corresponding magnetic mineral, because maximum blocking temperatures are always slightly lower than the Curie temperature. Different behaviours of IRM acquisition curves and thermal demagnetisation curves of the SIRM are illustrated in the following chapters of palaeomagnetic investigations.

Thermal demagnetisation up to 680°C of a composite IRM, containing three orthogonal axis of IRM’s (1,5 T, 0,5 T, 0,2 T, method as described by [*Lowrie*, 1990]) demonstrate the characteristics in more detail. Examples for samples with different magnetic behaviour are also given in the chapters 3 – 5.

The grain sizes of the remanence carriers were also investigated for many samples by measuring hysteresis properties e.g. [*Day et al.*, 1977]. These measurements were performed using the alternating gradient force magnetometer (AGFM, Princeton Measurements Corporation Micromag 2900) at the University of Tübingen. The hysteresis parameters $M_{rs}/M_s$ and $H_{cr}/H_c$ obtained, plot predominantly in the PSD field of the [*Day et al.*, 1977] diagram, or tend to higher $H_{cr}/H_c$ ratios. This behaviour is typical for specimens with mixed magneto-mineralogy, as also demonstrated by the other rockmagnetic experiments. In some cases, the behaviour of the hysteresis loop is characterized by slightly ‘wasp-waisted’ curve which could indicate a bimodal coercivity distribution. According to [*Jackson*, 1990], [*Channell and McCabe*, 1994] and [*McCabe and Channell*, 1994] this shape is interpretated as indication for remagnetisation and typically for such samples it was not possible to identify a primary magnetisation.

Widespread remagnetisation of rocks has been known since the early days of palaeomagnetism. This was evident from the bahaviour of the rocks during demagnetisation and from a comparsion of isolated remanence components with expected “younger” remanence directions for the sampling localities. In palaeomagnetic studies it often seems that remagnetisation is the rule rather than the exeption, thus the complex phenomenon of remagnetisation has been the subject of several different studies. For suggested reading and reinforcement the variety of processes e.g. thermal, chemical and tectonic influence leading to a remagnetisation is refered to following studies by [*Creer*, 1968], [*Henry*, 1973], [*Pullaiah et al.*, 1975], [*Kligfield et al.*, 1983], [*Oliver*, 1986], [*Bachtadse et al.*, 1987], [*Jackson et al.*, 1988], [*McCabe and Elmore*, 1989], [*Suk et al.*, 1990], [*Suk et al.*, 1991], [*Suk et al.*, 1993] and [*Van der Voo*, 1993] and references therein.

In summary, besides statistics of palaeomagnetic data rock magnetic studies are a necessary tool to assess the quality of obtained palaeomagnetic directions and the primary character of the isolated remanence components. Studies of the magneto-mineralogy can suggest that ferromagnetic grains carrying a characteristic remanent magnetisation (ChRM) are capable of retaining a primary NRM. However, laboratory tests alone cannot prove that the determined ChRM is primary. Field tests of palaeomagnetic stability can provide crucial information about the timing of ChRM acquisition. Common field tests of palaeomagnetic stability, applied in this thesis, are the fold test [*McElhinny*, 1964; *McFadden and Jones*, 1981], inclination only test [*Enkin and Watson*, 1996], conglomerate test [*Graham*, 1949] and reversal test [*McFadden and McElhinny*, 1990]. For further discussion the basic techniques of field tests to constrain the age of magnetisation the reader is refered to following textbooks by [*Soffel*, 1991], [*Butler*, 1992], [*Van der Voo*, 1993] and [*Opdyke and Channell*, 1996].

# Appendix to Chapter 2

The following photograps show a selection of several sampling locations of central and southern Europe, which studied in this thesis.

Photo 1: Outcrop of Upper Ordovician to Devonian sediments at the eastern flank of the Cellon near the Plöckenpass, Carnic Alps (results see chapter 4.2).

Photo 2: Outcrop of Early Ordovician volcanic sequences (KU) in the upper section of the Schicker quarry near Kupferberg (results see chapter 3.1).

Photo 3: Photo showing an example of a sampling location (DOL) of the Late Silurian Ockerkalk near Plauen (results see chapter 3.2).

Photo 4: Drilling of samples by using a petrol-driven rock drill in the Klingler-Kar, Kitzbüheler Alps, after a premature influx of the winter (results see chapter 4.1).

Photo 5: Outcrop of Middle Devonian volcanics at the Gaisstein (GS) in the Kitzbüheler Alps (results see chapter 4.1).

Photo 6: Taking samples from the Early Devonian Rauchkofel limestone in the Hochwipfel nappe (OBB), Carnic Alps (results see chapter 4.2).

Photo 7: Sampling Siluro/Devonian limestones near the top of the Rauchkofel, Carnic Alps.

Photo 8: Outcrop of Ordovician limestones (MOJ) in an old quarry near Mójzca, Holy-Cross-Mountains (results see chapter 5.1)

# Chapter 3 Palaeozoic Palaeomagnetic results from the Saxo-Thuringian

## Chapter 3.1 The Early Palaeozoic break-up of northern Gondwana, new palaeomagnetic data from the Saxothuringian Basin, Franconian Forest, Germany

INTRODUCTION

This investigation examines a subaqueous pyroclastic succession of Early Palaeozoic age from the Randschiefer Series and Vogtendorf Beds, exposed near Kupferberg in Bavaria (Figs 1-9 and 3-1). These rocks, which belong to the Saxothuringian Basin of the Saxothuringian Zone within the Variscan fold belt, are thought to be bi-modal in character. The age of these rocks is determined as 478±1.8 Ma [*Schätz et al.*, 2002], whereas the geodynamic and palaeogeographic setting under which they were deposited can only be loosely inferred. Few palaeogeographically diagnostic faunas have as yet been identified in Saxothuringia, but those that have (trilobites) show certain similarities to Barrandia, Baltica and Gondwana in the Early Ordovician [*Sdzuy*, 1955] [*Sdzuy*, 1971]. On geological grounds, it is clear that the Saxothuringian terrane was part of the Armorican Terrane Assemblage which encompasses Palaeozoic units of the European Variscan fold belt located south of the Rhenohercynian Zone (Avalonia) and the presence of Cadomian basement in most units of the Armorican Terrane Assemblage and recent provenance analyses demonstrate they were adjacent to the North African Gondwana margin in the Early Cambrian [*Kröner et al.*, 1995]. No palaeomagnetic data have yet been published for the Saxothuringian in early Palaeozoic times, however for the Early Ordovician few data have been published for the Armorican Terrane Assemblage as a whole (see chapter 1.2) which indicate high palaelatitudes for this time and set in rifting in Mid Ordovician times [*Tait et al.*, 1997]. Early Ordovician sedimentary and volcanic rocks in the Saxothuringian Terrane, however, are suggestive of rifting from northern Gondwana [*Linnemann et al.*, 2000], similar to the scenario indicated by palaeomagnetic evidence from Barrandia by [*Tait et al.*, 1997].

For Late Ordovician times the picture is even less clear (see chapter 1.2). No reliable palaeomagnetic data are as yet available for the Late Ordovician of Saxothuringia, thus the affinities and movement of these blocks can only be constrained by faunal and lithological indicators. Late Ordovician sediments of Saxothuringia are characterised by Ashgillian glacio-marine deposits, similar to those found elsewhere in the Armorican Terrane Assemblage and are related to the Saharan glaciation centred in northern Africa. Studies indicate, however, that these diamictites were deposited from floating or seasonal ice [*Brenchley et al.*, 1991] and not from the main ice sheet itself, and the effects of the Ashgillian cooling may have extended to more intermediate palaeolatitudes [*Owen et al.*, 1991]. The presence of glacio-marine deposits, therefore, does not necessarily restrict deposition in a peri-polar environment.

To help to constrain the palaeogeography setting of Saxothuringia in Ordovician times, detailed palaeomagnetic studies of the well dated formation of the volcanic rocks [*Schätz et al.*, 2002], and the biostratigraphic dated overlying Ashgillian sediments [*Sdzuy*, 1971] [*Sannemann*, 1955] were carried out.

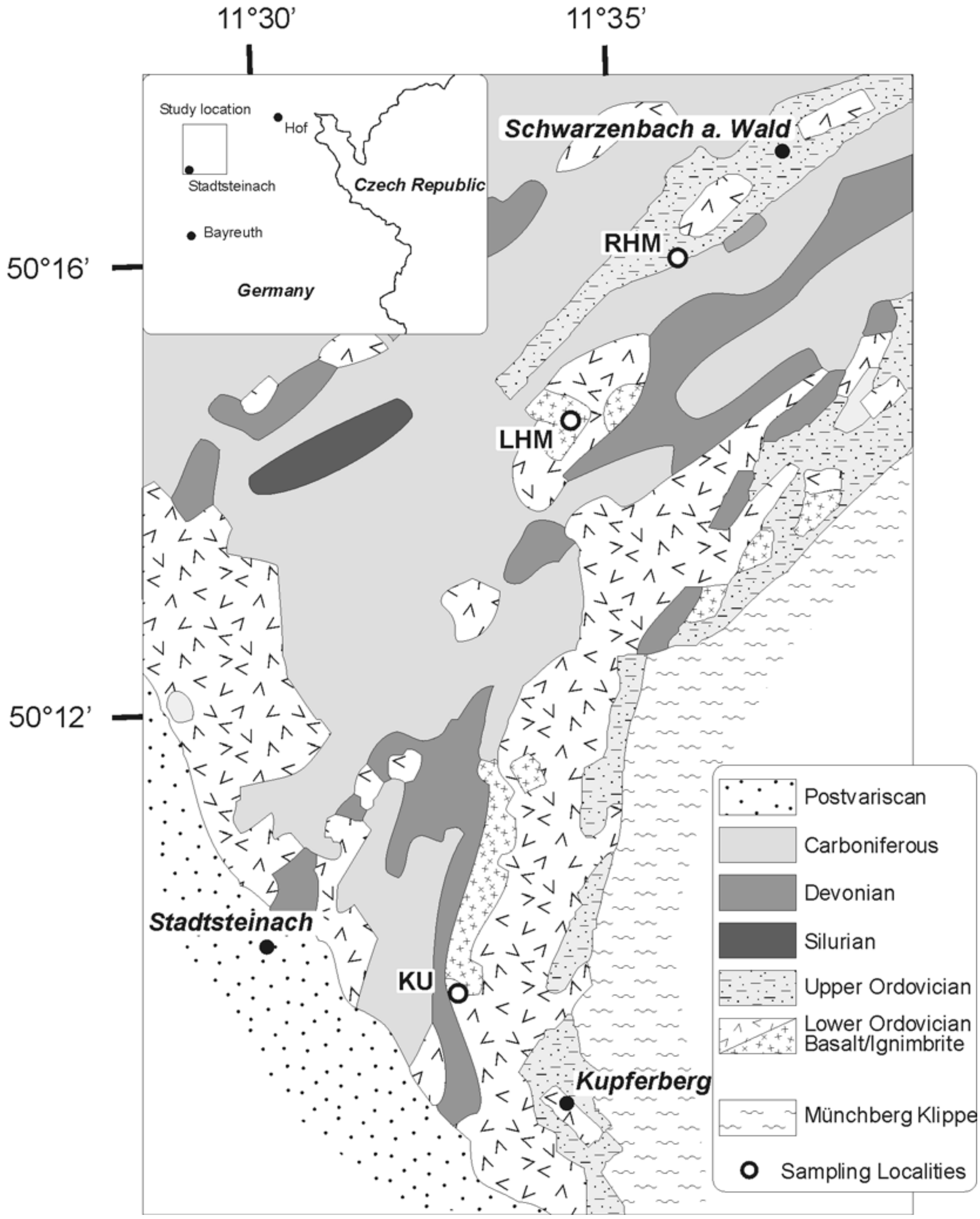



Figure 3-1: Simplified geological sketch map of the Franconian Forest, NE Bavaria. Open circles represent sampling localities discussed in the text.

## SAMPLING AND MEASUREMENTS

Samples for Early Ordovician palaeomagnetic analysis were collected from two localities (Fig. 3-1), comprizing twelve sites from the Early Ordovician volcanic sequence exposed at the Schicker quarry near Kupferberg (sites KU1 - KU12), and three sites from coeval volcanic rocks which crop out close to Schwarzenbach am Wald (LHM1 - LHM3), thus allowing for a fold test on a regional scale. Petrological and geochemical studies show that these sequences have undergone no, or at most only extremely low grades of metamorphism. The deformation in the region as a whole is thought to have peaked in Carboniferous times, associated with main thermal event identified in this part of the Variscan fold belt at about 340 Ma. Late Ordovician samples (four sites, 21 samples) were also collected from the Döbra sandstone (RHM1 – RHM4), which is exposed near Schwarzenbach am Wald (Fig. 3-1). These fine grained sediments are well constrained biostratigraphically (conodonts) as being Ashgillian in age [*Sannemann*, 1955] and were sampled for palaeomagnetic analysis to determine the palaeogeography of Saxothuringia in the Late Ordovician.

Table 3-1: Palaeomagnetic site mean directions and overall mean directions obtained for the Early Ordovician volcanic, and Late Ordovician sedimentary rocks of the Saxothuringian.

| Site | Age | Rocktype | Bedding | N/n | Dec/Inc (IS) | k | $\alpha_{95}$° | Dec/Inc (TC) | k | $\alpha_{95}$° |
|---|---|---|---|---|---|---|---|---|---|---|
| KU 7 | 478.2±1.8# | ignimbrites | 009/34 | 11 / 8 | 203/50 | 9.4 | 19.0 | 249/80 | 9.4 | 19.0 |
| KU 8 | 478.2±1.8# | ignimbrites | 354/32 | 9 / 4 | 181/33 | 92.0 | 9.6 | 187/64 | 92.0 | 9.6 |
| KU 9 | 478.2±1.8# | ignimbrites | 354/32 | 6 / 4 | 163/52 | 28.7 | 17.4 | 112/79 | 31.7 | 16.6 |
| KU 10 | 478.2±1.8# | ignimbrites | 006/34 | 5 / 3 | 179/36 | 36.7 | 20.6 | 171/70 | 36.7 | 20.6 |
| LHM 1 | Early Ord | ignimbrites | 124/20 | 11 / 6 | 263/64 | 7.4 | 26.4 | 215/73 | 7.4 | 26.4 |
| site mean direction | | | | 42 / 25 | 190/51 | 10.7 | 24.5 | 189/76 | 44.7 | 11.6 |

| Site | Age | Rocktype | Bedding | N/n | Dec/Inc (IS) | k | $\alpha_{95}$° | Dec/Inc (TC) | k | $\alpha_{95}$° |
|---|---|---|---|---|---|---|---|---|---|---|
| RHM 1 | Ashgill. | clastics | 027/15* | 5 / 4 | 226/40 | 11.4 | 28.4 | 003/-54 | 11.4 | 28.4 |
| RHM 2 | Ashgill. | clastics | 027/15* | 5 / 4 | 206/58 | 11.0 | 29.0 | 029/-73 | 11.0 | 29.0 |
| RHM 3 | Ashgill. | clastics | 027/15* | 6 / 4 | 199/35 | 16.2 | 23.6 | 037/-50 | 16.2 | 23.6 |
| RHM 4 | Ashgill. | clastics | 027/15* | 5 / 3 | 191/34 | 7.6 | 48.6 | 047/-48 | 7.6 | 48.6 |
| site mean direction | | | | 21 / 15 | 205/42 | 25.7 | 18.5 | 030/-58 | 25.7 | 18.5 |

N, number of samples measured; n, number of samples used in calculation of site mean; Dec/Inc, declination/inclination of site means in situ (IS) and after tilt correction (TC) in degrees; k, precision parameter and $\alpha_{95}$°, half angle of the cone of 95% confidence [*Fisher*, 1953]. Bedding correction is given in terms of direction of dip/dip in degrees, # $^{207}Pb/^{206}Pb$ age on single zircons, *beds overturned

The structural parameters (e.g. strike and dip) were readily measured at each site. The bedding attitude, lithologies, and palaeomagnetic results of those yielding interpretable results are listed in Table 3-1. In addition, the anisotropy of magnetic susceptibility was measured for a number of volcanic samples and the results show that the rocks are essentially isotropic, with maximum anisotropy values for individual samples of 4%, with the majority having 1-2% anisotropy.

PALAEOMAGENTIC RESULTS

**Early Ordovician volcanic rocks**

During thermal demagnetisation of samples from the fifteen sites (KU1-12, LHM1-3) collected from the Early Ordovician volcanic rocks, up to 2 palaeomagnetic directions, termed B and C, can be identified, often after removal of a low unblocking temperature (Tub) A component which generally parallels the local present day Earth's magnetic field (Fig. 3-2), and is thought to be of recent origin.

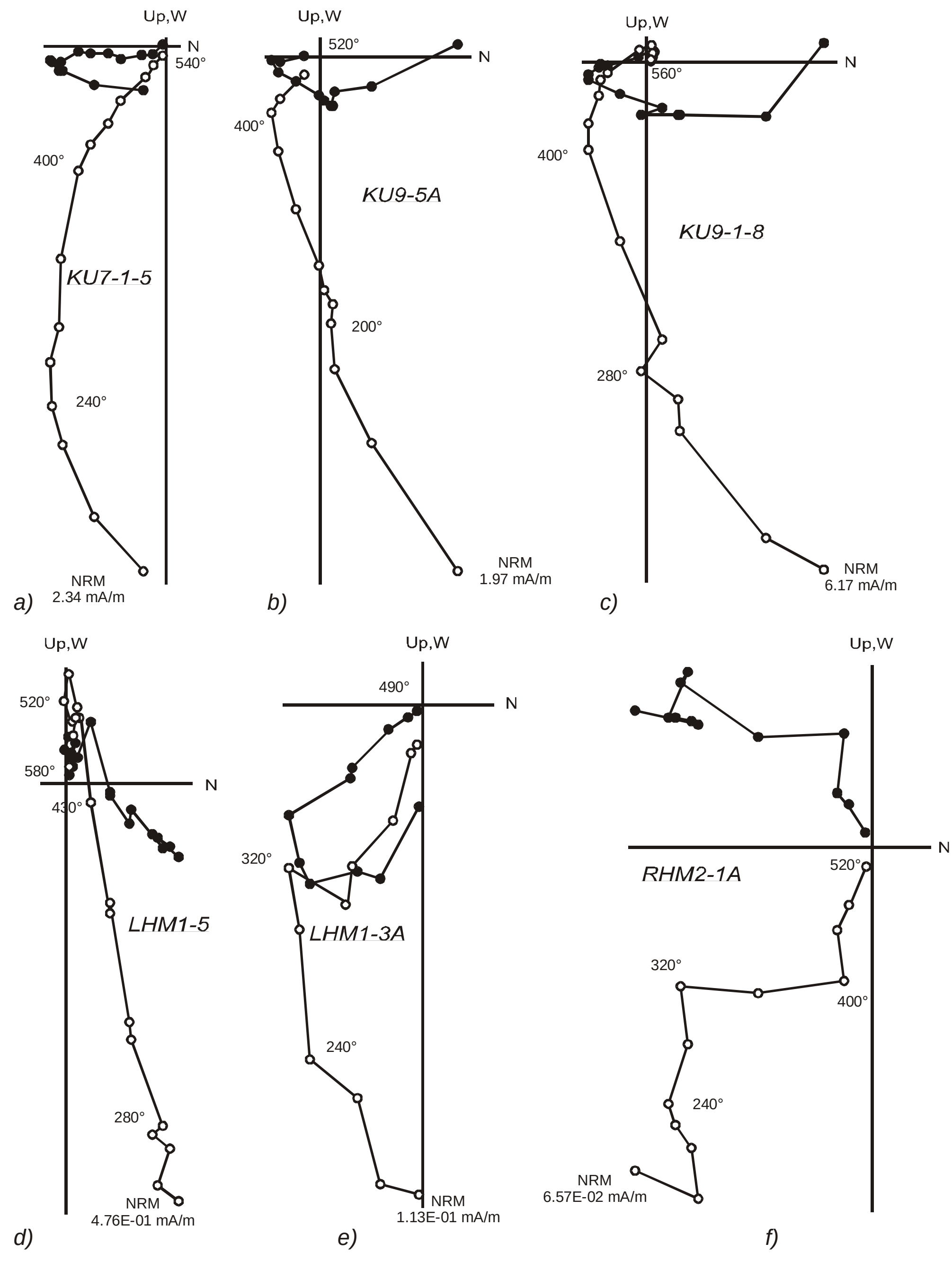



Figure 3-2: Orthogonal projection of thermal demagnetisation data for samples of the Early Ordovician volcanic samples (a-e) and the Late Ordovician clastic samples (f). Directions are plotted in situ, solid (open) symbols represent the horizontal (vertical) component, respectively.

Many samples then show removal of an intermediate-high temperature B magnetisation. Though only poorly defined, this component is shallow and southerly directed in situ, and is of single polarity. It often persists until total demagnetisation, and demagnetisation trajectories often follow great circle paths towards a higher temperature direction. No stable end point directions for this higher temperature component could be identified, however, thus further analysis of the great circle paths is not possible. This was the only palaeomagnetic component identified in sites KU1-KU4 which are located in the basal part of the quarry. For these sites, rock magnetic experiments suggest coarse grained magnetite as the main carrier and micro and macroscopic observations suggest relatively high degrees of alteration. Component B, therefore, is considered to be secondary in origin. Due to the uncertainties involved, no further interpretation is given for this component and it will not be discussed further.

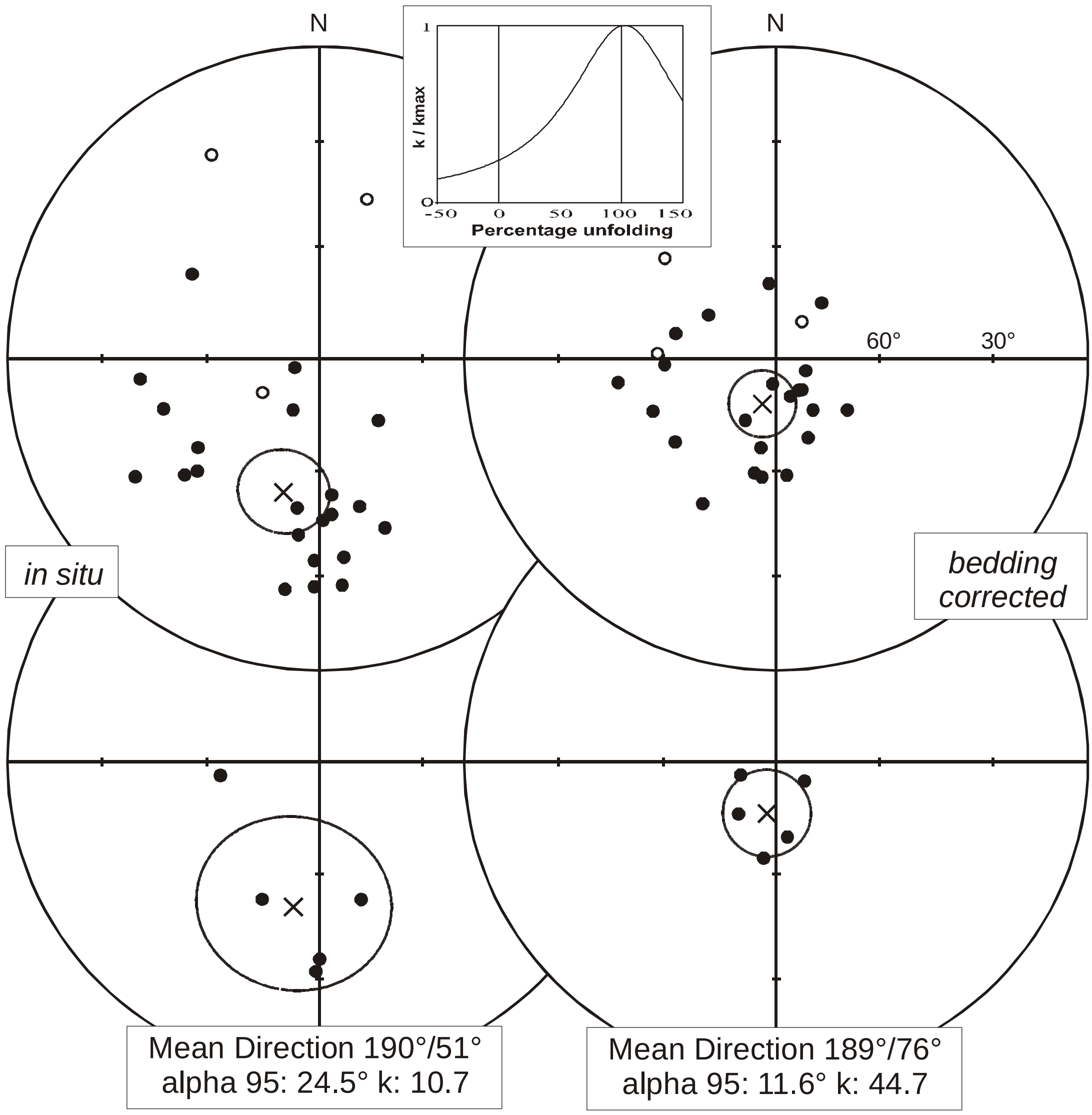



Figure 3-3: Equal area stereographic projections of the Early Ordovician sample (top) and site mean (bottom) C magnetisation before and after structural correction, and the k/kmax rations versus percentage unfolding. Solid (open) circles represent lower (upper) hemisphere projections; k is the precision parameter [*Fisher*, 1953], and alpha 95 is the cone of 95% confidence.

Component C can be identified in 25 samples from five sites (KU7-10, LHM1, Table 3-1) at high unblocking temperatures ($T_{ub}$), generally after removal of direction A and a poorly defined B direction at lower temperatures (Fig. 3-2a-e). Component C is identified as the stable end point direction, is of dual polarity (Fig. 3-2), and is generally well defined with discrete $T_{ub}$ and maximum $T_{ub}$ of 580°C, suggesting magnetite as the principal magnetic carrier. Direction C is fairly scattered in situ with southerly to westerly declinations and intermediate to steep inclination values (Fig. 3-3). There is significant improvement in the statistical distribution of results after bedding correction, however, yielding a steep southerly mean direction of magnetisation (Fig. 3-3, Table 3-2).

Isothermal remanent magnetisation acquisition curves of these samples indicate saturation at about 300mT to 500mT (Fig. 3-4a). A few samples also contain a small amount of a higher coercivity phase (Fig. 3-4a). Thermal demagnetisation of the SIRM yields high unblocking temperatures of 550°-580°C, thus supporting the interpretation that magnetite is the principal carrier of the magnetisation in these rocks.

Among the samples from the remaining six sites collected in the Early Ordovician sequences (sites KU5-6, KU11-12, LHM2-3), the sample magnetisations yielded no stable direction of magnetisation. The remanences were dominated by direction A, or there was no within site consistency of results. The data obtained from these six sites are not listed in Table 3-1 and are not discussed further.

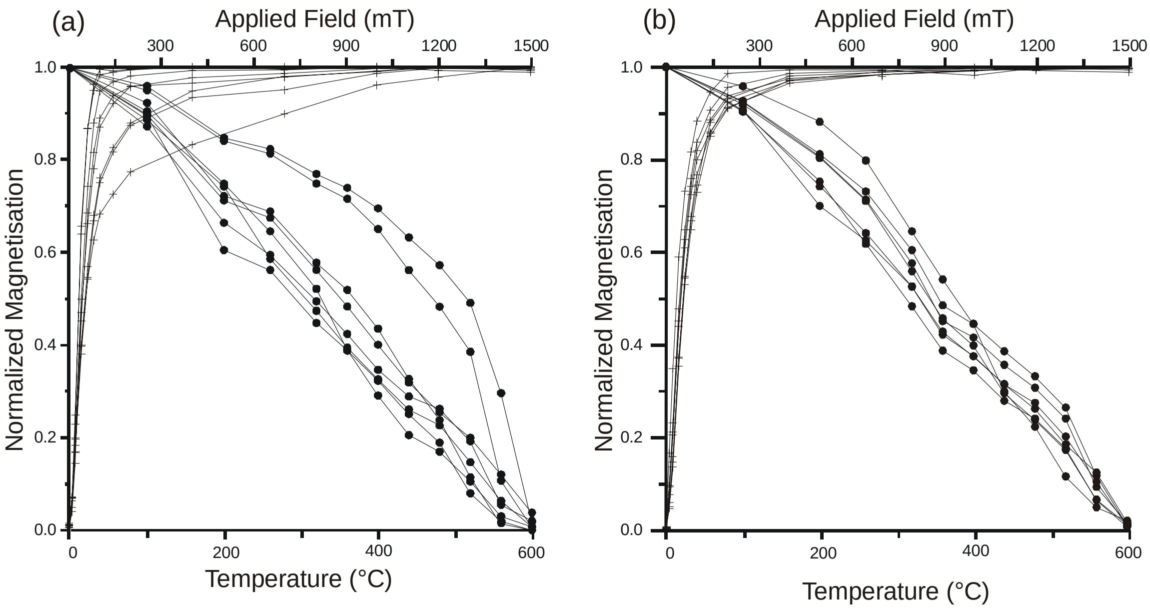



Figure 3-4: Normalised isothermal remanent (IRM) acquisition curves (crosses) and intensity decay plots of the subsequent thermal demagnetisation of the IRM (circles) for the Early Ordovician volcanic samples (a) and the Late Ordovician clastic sediments (b).

**Upper Ordovician Döbra sandstone**

Samples from the Late Ordovician Döbra sandstone (sites RHM1-4) are generally only very weakly magnetised. Nevertheless it was possible to identify a high temperature palaeomagnetic direction termed component D (Fig. 3-2f). Demagnetisation up to 300°C results in removal of a low temperature magnetic component, similar to direction A. At intermediate temperatures, some samples show removal of a poorly defined component of magnetisation which is similar in direction to component B (Fig. 3-2f). Up to the maximum $T_{ub}$, component D can be identified in a total of fifteen samples (4 sites), with southerly declinations and intermediate inclination values in situ (Fig. 3-5). Rock magnetic studies (IRM and thermal demagnetisation of SIRM) again demonstrate the predominance of magnetite as the carrier of the magnetisation in these samples (Fig. 3-4b).

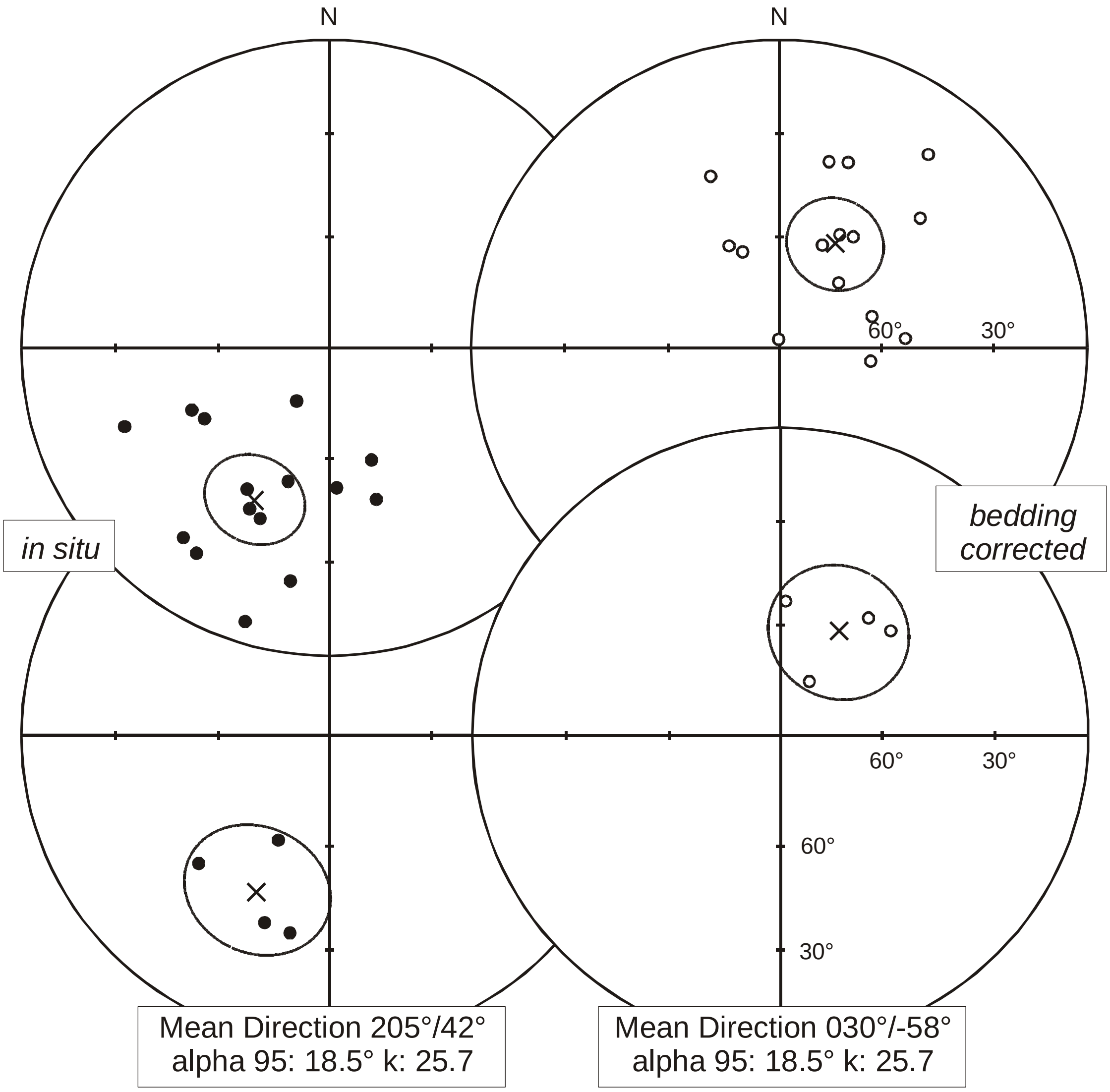



Figure 3-5: Equal area projection of the Late Ordovician sample (top) and site mean (bottom) D magnetisation before and after structural correction. Notation as for figure 3-3.

## INTERPRETATION AND PALAEOGEOGRAPHIC IMPLICATIONS

Direction C, identified in 25 samples (five sites) of Early Ordovician age, is of dual polarity and yields an overall mean direction of 190°/51°, 95=24.5°, k=10.7 in situ, and 189°/76°, 95=11.6°, k=44.7 after bedding correction (Fig. 3-3, Table 3-1). This improvement in grouping is statistically significant and passes the fold test of [*McElhinny*, 1964] with 99% confidence at 105% unfolding. Component C, therefore, is interpreted as being primary in origin and closely linked to the Early Arenig rock age. The resulting palaeopole position of 24°N/007°E translates into palaeolatitudes of approximately 63°S (+10.4°, -9.3°) for the Saxothuringian Terrane in Early Ordovician (Arenig) times. This result is similar to results obtained from Barrandia [*Tait et al.*, 1994] and the Armorican Massif [*Cogne and Perroud*, 1988] [*Cogne et al.*, 1991] [*Perroud et al.*, 1986], thus demonstrating that, as inferred from geological evidence, Saxothuringia was part of the Armorican Terrane Assemblage (at least in early Ordovician times) and was at high latitudes, close to the northern margin of Gondwana.

Component D is identified only in the Late Ordovician age samples, and D yields an overall mean direction of 205°/42°, a95=18.5°, k=25.7 in situ, and 030°/-58°, a95=18.5°, k=25.7 after bedding correction (Fig. 3-5, Table 3-1). This component can only be identified in 15 samples and from beds of uniform dip, thus it is not possible to conduct a fold test to help to constrain the relative age of magnetisation. As no palaeomagnetic data from this region have as yet been published, the direction to be expected for Late Ordovician to Late Carboniferous times in this region is not clearly defined. For post late Carboniferous times geological evidence is clear that Saxothuringia was part of Laurasia, thus expected directions can be estimated from the APW path for Stable Europe. The in situ D direction, however, is not similar to any direction which may be expected for post-folding times (i.e. post Late Carboniferous times). Furthermore, a component interpreted as being a late Carboniferous overprint direction (component B) is removed at intermediate temperatures, suggesting an older age for component D. Other possible interpretations are either that D is a pre-folding remagnetisation, or it is primary in origin. Given the fact that the bedding corrected direction is significantly different to that obtained from the Early Ordovician sequences (thus ruling out a widespread pre-folding remagnetisation in the region), and that there is no apparent geological evidence for a local pre-Carboniferous thermal event which may cause a remagnetisation, the D direction is considered to be primary in origin and Late Ordovician in age. Given this interpretation, the inclination values indicate palaeolatitudes of 38°S(+11.4°,-8,9°) for the Saxothuringian Terrane in the Late Ordovician. This is coincident with palaeolatitudes obtained for Ashgillian age rocks from the Tepla-Barrandian of the Armorican Terrane Assemblage [*Tait et al.*, 1995]. However, we are aware that given the small number of samples in which this component is identified in this study, and the lack of field tests to accurately constrain the relative age of magnetisation, these results for Saxothuringia must remain preliminary until such time as more data become available.

# Chapter 3.2 Silurian palaeogeography of the Variscan fold belt: Palaeomagnetic constraints for the Armorican terrane Assemblage from the Saxothuringian Basin

## INTRODUCTION

Within the Variscan foldbelt of Europe a number of Gondwana-derived palaeozoic terranes and blocks, i.e. Iberian, Armorican and Bohemian Massifs and the Saxothuringian basin are revealed (see chapter 1.2) However palaeomagnetic data have brought into question the hypothesis of a coherent Armorican microplate and the term Armorican Terrane Assemblage (ATA) is suggested for the Palaeozoic tectono-stratigraphic units now situated within the European Variscan fold belt [*Tait*, 1999]. Palaeomagnetic, faunal and lithological data demonstrate fairly clear that in Late Ordovician times Barrandia and Saxothuringia were moving northward and away from northern Gondwana. The onset of rifting of Saxothuringia from the north African margin of Gondwana and the start of the relative northward migration of the Saxothuringian Terrane were shown in chapter 3.1. By the Late Ordovician palaeomagnetic data for the Tepla-Barrandian terrane of the Bohemian Massif [*Tait et al.*, 1995] and the new data for Saxothuringian reported in chapter 3.1 both terranes have palaeolatitudes of approximatly 40°S and have an almost common drift history.

By the Latest Silurian-Early Devonian the Tepla-Barrandian terrane, the Catalunian and Eastern Pyrenean block of NE Spain and the Armorican Massif were situated at some 30°S [*Tait*, 1999; *Tait et al.*, 2000b], the Tepla-Barrandian terrane, however, was inverted with respect to its present day orientation. Rotations have not identified in the Armorician Massif, thus leading to the hypothesis of the Armorican terrane assemblage.

Within this model, the rotation of the Bohemian massif is constrained to have occurred prior to the late Devonian.. The role of the Saxothuringian zone within this system, however, remains unresolved. Faunal and lithological indicators suggest affinities with Armorica and Bohemia, but no palaeomagnetic data are available to constrain its palaeogeography and tectonic relationships in the mid Palaeozoic. The paleogeographic position of the Saxothuringian zone, therefore, is the key area within the Variscan fold belt to help to resolve questions concerning the the Armorican terrane assemblage and the tectonic evolution of central Europe.

## SAMPLING

To resolve questions concerning the palaeogeographic position of the STB and its tectonic relationship to other elements of the ATA in Mid Palaeozoic times, samples have been collected from Upper Silurian age Ockerkalk sequences of the Thuringian facies in the area SE of the Berga Anticline in Saxonia in the Vogländisches Synclynorium (Figs. 1-9 and 3-6). Ordovician shales and sandstones have also been sampled but no primary directions were achieved by palaeomagnetic studies.. In total, 61 palaeomagnetic cores from of 9 sites (DOL1-DOL3, JOK1-

JOK4, NED1 and NED2) of the Ockerkalk were collected at three different localities near Plauen (Fig. 3-6) with differences in the structural attitude allowing for a fold test on the regional scale.

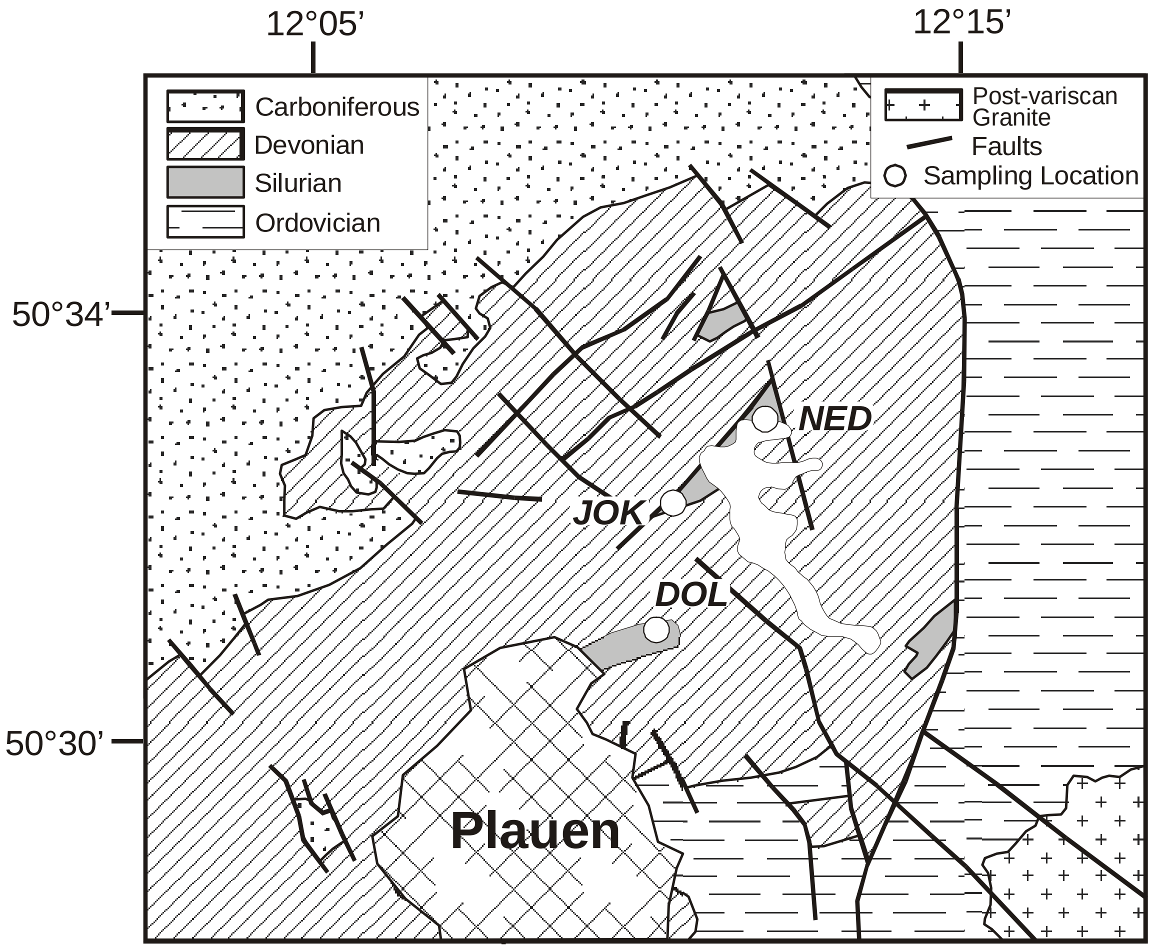



Figure 3-6: Geological sketch map of Vogtländisches Synclynorium, SE Saxonia. Open circles represent the sampling locations discussed in the text.

ROCK MAGNETISM

During IRM experiments saturation of most samples is reached by 400-500 mT, demonstrating the predominance of low and intermediate coercivity minerals as remanence carriers (Fig. 3-7). Occasionally, there is a slight gradual increase in magnetisation at higher fields, up to the maximum applied field of 1.5 T, indicating the presence of high coercivity minerals.

Demagnetisation of multi component IRM's show that magnetite is the predominant carrier, with low coercivity (<0.2T) and maximum unblocking temperatures of approx. 580°C. The higher coercivity minerals (0.2-0.5T and 0.5-1.5T) are thought to be pyrrhotite (Fig. 3-8a) with demagnetization by approximatly 400°C, and varying amounts of goethite (Fig. 3-8a), magnetite (Fig 3-8b), and occasionally haematite (Fig. 3-8c). Hysteresis measurements show that the grain sizes of magnetite vary in between the pseudo-single domain and the multidomain range [*Day et* *al.*, 1977]. This bimodal coercivity distribution is also indicated by wasp-waisted hystersis loops in some cases [*Nagata and Carleton*, 1987], typical for rocks with mixed magneto-mineralogy and mixed grain sizes.

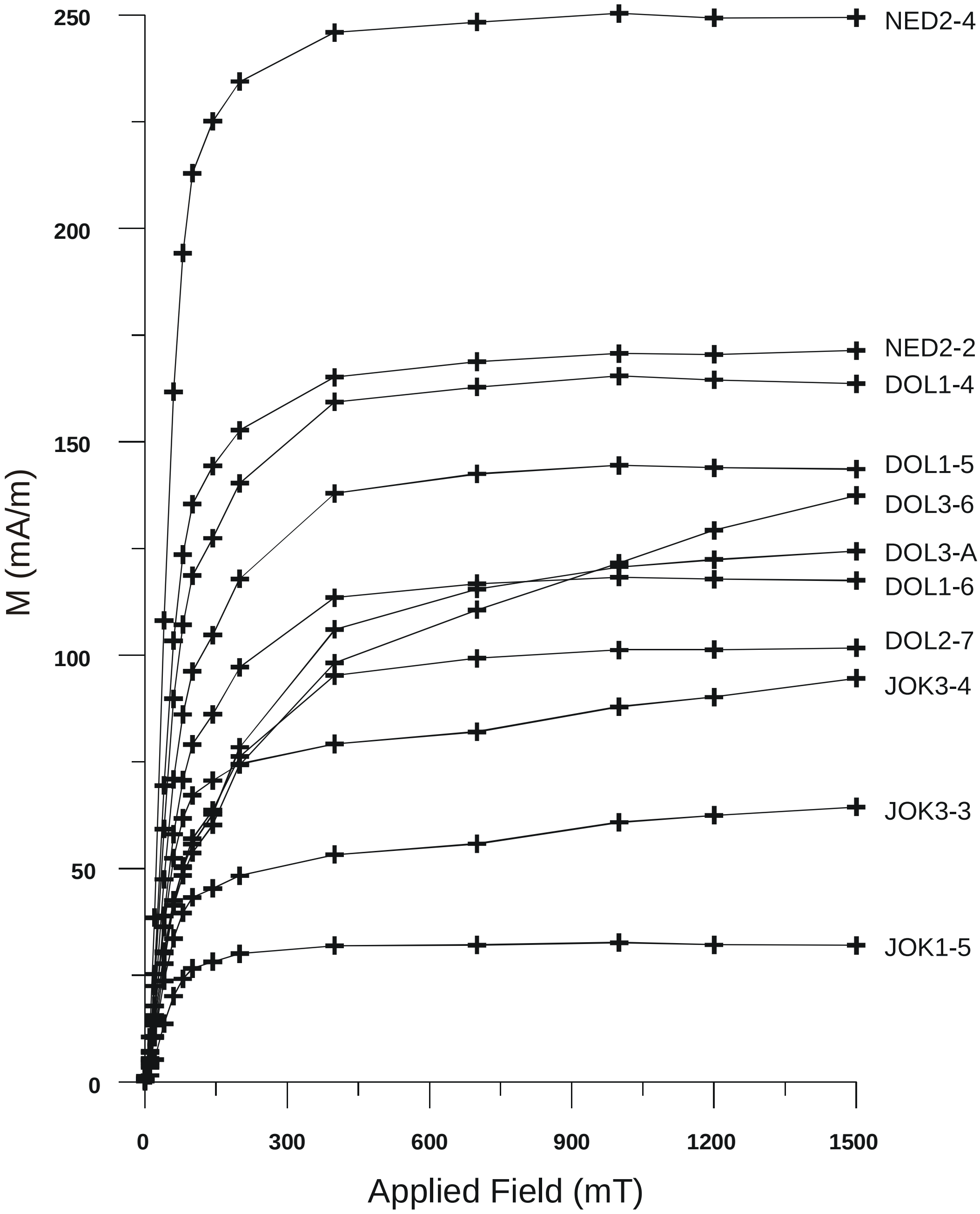
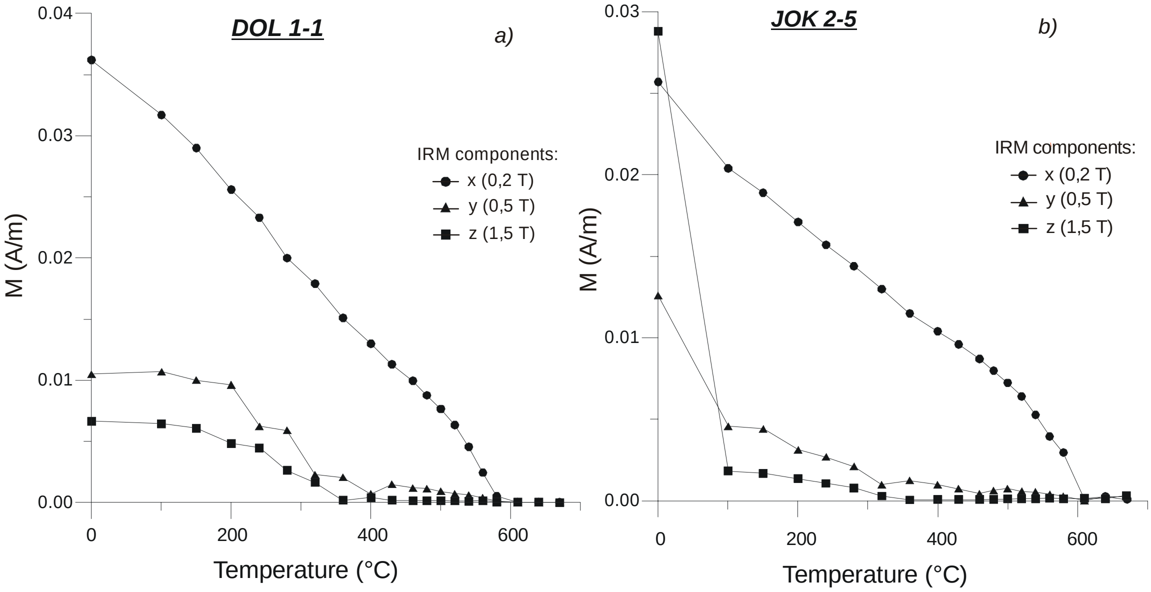



Figure 3-7: IRM acquisition curves of monodirectional isothermal remanent magnetisation



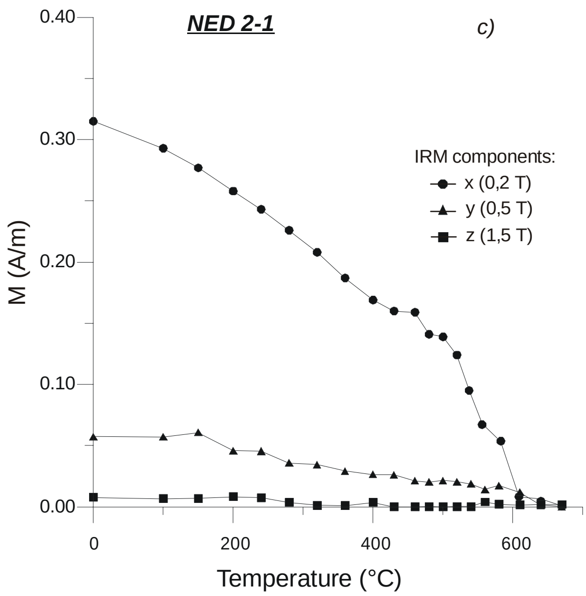




Figure 3-8: Thermal demagnetisation of triaxial differntial IRM (method described in [*Lowrie*, 1990]) in (a) of speciment DOL1-1, in (b) of speciment JOK2-5 and in (c) of speciment NED2-1.

## PALAEOMAGNETIC RESULTS

The initial natural remanent magnetisation (NRM) intensities are low, rangingfrom 0.2 – 0.9 mA/m. Most of the samples show multivectorial behaviour, sometimes with overlapping unblocking temperature (Tub) spectra, leading to great circle behaviour in the demagnetisation trajectories. (Fig. 3-9).

In order to allow full identification of the remanence carriers, given the mixed mineralogy indicated by the rock magnetic experiments, some samples were heating to 200°C (to remove effects of goethite), prior to subsequent alternating field (AF) demagnetization. Most of the samples, however, have been subjected simply to detailed stepwise thermal demagnetization. Between room temperature and 200°C almost all samples reveal a northerly steep magnetic direction termed component A (Table 3-2, Fig. 3-9). It is generally parallel to the local present-day Earth's magnetic field and is an overprint direction of recent origin. The results of subsequent AF and thermal demagnetisation, however, show differing characteristics.

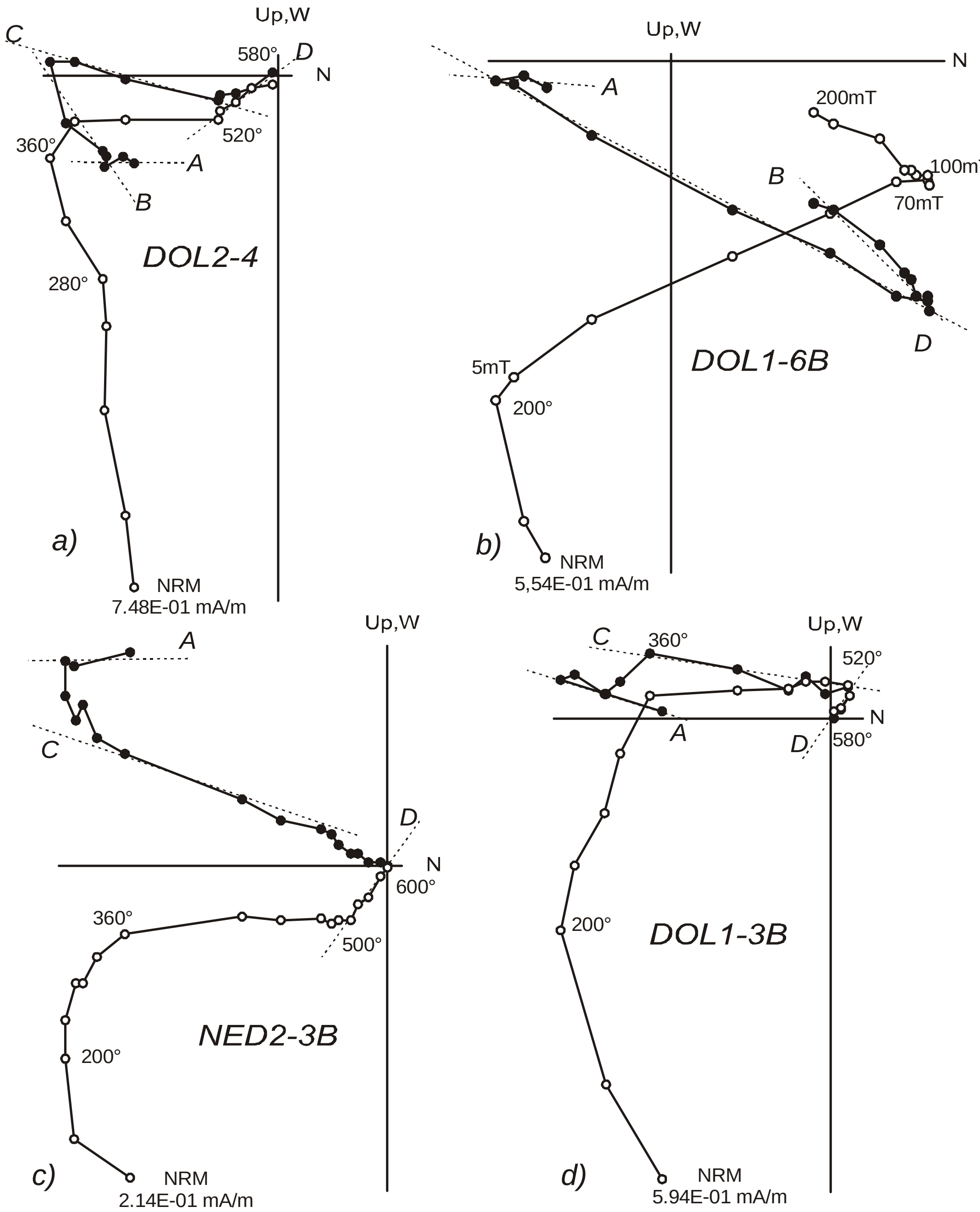



Figure 3-9: Orthogonal projection of thermal and AF-demagnetisation behaviour of samples of the Silurian carbonates. All directions are plotted in geographic coordinates and solid (open) symbols represent the horizontal (vertical) component respectively. Specific demagnetisation temperatures and applied field are given in degree Celsius and mT.

**Alternating field demagnetisation**

During stepwise AF-demagnetisation in the 5 mT to 70 mT interval of samples already heated to 200°C, a southerly intermediate direction D can be identified (Fig. 3-9b). At higher fields a

Table 3-2: Palaeomagnetic site mean directions and overall mean directions for components A, B and C obtained for the Upper Silurian Ockerkalk

| | | | Component A In Situ | | | | Component B In Situ | | | | Component C In Situ | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| site | age | Lithology | N/n | Dec/Inc | k | $\alpha_{95}$ | N/n | Dec/Inc | k | $\alpha_{95}$ | N/n | Dec/Inc | k | $\alpha_{95}$ |
| DOL 1 | U. Sil | limestone | 7 / 7 | 009 / 66 | 66.7 | 7.4 | 7 / 7 | 039 / 30 | 36.0 | 10.2 | 7 / 5 | 195 / -2 | 93.9 | 7.9 |
| DOL 2 | U. Sil | limestone | 9 / 8 | 009 / 66 | 96.2 | 5.7 | 9 / 8 | 043 / 31 | 25.6 | 11.2 | 9 / 4 | 196 / 5 | 76.3 | 10.6 |
| DOL 3 | U. Sil | limestone | 6 / 5 | 033 / 72 | 26.7 | 15.1 | 6 / 4 | 054 / 24 | 33.6 | 16.1 | 6 / 0 | - | - | - |
| JOK 1 | U. Sil | limestone | 6 / 6 | 005 / 60 | 15.3 | 17.7 | 6 / 0 | - | - | - | 6 / 3 | 195 / -2 | 33.7 | 21.6 |
| JOK 2# | U. Sil | limestone | 7 / 6 | 004 / 64 | 54.4 | 9.2 | 7 / 0 | - | - | - | 7 / 4 | 194 / -2 | 22.9 | 19.6 |
| JOK 3 | U. Sil | limestone | 5 / 5 | 351 / 64 | 45.4 | 11.5 | 5 / 0 | - | - | - | 5 / 3 | 196 / 15 | 132.2 | 10.8 |
| JOK 4# | U. Sil | limestone | 9 / 9 | 320 / 72 | 53.7 | 7.1 | 9 / 0 | - | - | - | 9 / 7 | 193 / -6 | 102.9 | 6.0 |
| NED 1 | U. Sil | limestone | 6 / 5 | 062 / 80 | 22.0 | 16.7 | 6 / 6 | 021 / 57 | 15.4 | 17.6 | 6 / 0 | - | - | - |
| NED 2 | U. Sil | limestone | 6 / 5 | 348 / 75 | 31.1 | 13.9 | 6 / 2 | 046 / 42 | - | - | 6 / 4 | 214 / 6 | 13.0 | 26.5 |
| overall mean direction | | | 9 sites | 002 / 69 | 26.6 | 3.8 | 5 sites | 040 / 35 | 16.1 | 6.4 | 7 sites | 197 / 1 | 28.2 | 4.8 |

N, number of samples measured; n, number of samples used in calculation of site mean; Dec/Inc, declination/inclination of site means in degree; k, precision parameter [*Fisher*, 1953]; $\alpha_{95}$ , half angle of the cone of 95% confidence [*Fisher*, 1953]. # sites not used in calculation of the respective overall means of component B.

second northeasterly directed magnetization with intermediate positive incination values can be seen. This is termed component B.

**Thermal demagnetisation**

Between the temperature steps of 200°C and 360°C the demagnetization trajectories generally move along great circles from the present day direction to a southerly shallow C direction of magnetization which has maximum unblocking temperatures of 520°C(Table 1, Fig. 3-9a,b,c and d). Occassionally, a northeasterly steep direction B can be identified in the 200°-360°C unblocking temperature range (Fig. 3-9a). At unblocking temperatures of 520°-600°C , the stable endpoint direction D can be identified (Fig. 3-9a,c and d). This direction is identified with dual polarity and is directed towards the origin in orthogonal projection(Table 3-3, Fig. 3-9a,c and d). In some samples, however, no stable endpoint direction could be determined due to overlapping demagnetisation spectra and demagnetisation trajectories move along great circle paths from northerly positive A directions to south-westerly intermediate D directions.

Table 3-3: Palaeomagnetic site mean direction and overall mean directions for components D of the Upper Silurian Ockerkalk

| | | | | Component D | | | | | |
|---|---|---|---|---|---|---|---|---|---|
| | | | | In Situ | | | 90% Bedding Corrected | | |
| site | age | Lithology | N/n | Dec/Inc | k | $\alpha_{95}$ | Dec/Inc | k | $\alpha_{95}$ |
| DOL 1 | U. Sil | limestone | 7 / 5 | 172 / 22 | 9.2 | 26.6 | 225 / 51 | 9.2 | 26.6 |
| DOL 2 | U. Sil | limestone | 9 / 6 | 182 / 15 | 30.5 | 12.3 | 218 / 40 | 30.5 | 12.3 |
| DOL 3 | U. Sil | limestone | 6 / 3 | 181 / 11 | 70.0 | 14.9 | 225 / 34 | 196.5 | 8.8 |
| JOK 1[#] | U. Sil | limestone | 6 / 0 | - / - | - | - | - / - | - | - |
| JOK 2 | U. Sil | limestone | 7 / 6 | 190 / -2 | 53.2 | 9.3 | 219 / 33 | 53.2 | 9.3 |
| JOK 3 | U. Sil | limestone | 5 / 3 | 238 / -28 | 73.2 | 14.5 | 234 / 32 | 73.2 | 14.5 |
| JOK 4 | U. Sil | limestone | 9 / 4 | 212 / -7 | 24.5 | 25.7 | 200 / 47 | 24.5 | 25.7 |
| NED 1 | U. Sil | limestone | 6 / 3 | 208 / 33 | 60.8 | 16.0 | 246 / 37 | 60.8 | 16.0 |
| NED 2 | U. Sil | limestone | 6 / 4 | 200 / 35 | 115.8 | 8.6 | 234 / 27 | 115.8 | 8.6 |
| overall mean direction | | | 8 sites | 197 / 8 | 9.1 | 19.5 | 226 / 38 | 38.2 | 9.1 |

N, number of samples measured; n, number of samples used in calculation of site mean; Dec/Inc, declination/inclination of site means in degree; k, precision parameter [*Fisher*, 1953]; $\alpha_{95}$ , half angle of the cone of 95% confidence [*Fisher*, 1953]. [#] sites not used in calculation of the respective overall means of component D.

## INTERPRETATION

The palaeomagnetic and rockmagnetic analysis reveals a complex and multivectorial magnetic behaviour within the sampled sections with identification of up to four directions of magnetisation (Table 3-2 and 3-3). The component A was determined in almost all specimens from the nine sites (Table 3-2, Fig. 3-10a) and is removed below 200°C. Combining the site mean data listed in Table 3-2 yields an overall in situ mean direction of 002°/69°, k=26.6 and α95=3.8 which is similar to the local present-day Earth`s magnetic field and is interpreted as being of recent origin. Magnetisation B is identified in a limited number of samples in with a maximum Tub of 360° and intermediate to high coercivity range (Table 3-2). It yields an in situ site mean direction of 040°/35°, k=16.1 and α95=6.4 and fails the fold test (Table 3-2, Fig. 3-10b). Folding of these sequences occurred in the Carboniferious during the Variscan orogeny, by which time it is clear from geological and palaeomagnetic evidence [*Van der Voo*, 1993] [*Tait et al.*, 2000a] [*Franke*, 2000] that the STB and the ATA were part of the Old Red Continent. The resulting paleopole of 312°E, 46°S calculated for component B corresponds to the Triassic segment of the apparent polar wander path (APWP) of Baltica [*Smethurst et al.*, 1998] [*Van der Voo*, 1993]. Therefore, Component B is considered to be of secondary origin, and Triassic in age. The southerly shallow C direction of magnetization is identified in a total of seven sites (30 samples) yields an overall in situ mean direction of 197°/01°, k=28.2 and α95=4.8 (Table 3-2, Fig. 3-10c). It also fails the fold test and corresponds to a paleopole position of 351°E, 37°S. This clearly coincides with the Permo-Carboniferous sector of the European APWP and, therefore, represents a remagnetisation event of this age, an event which is ubiquitous throughout Variscan Europe.

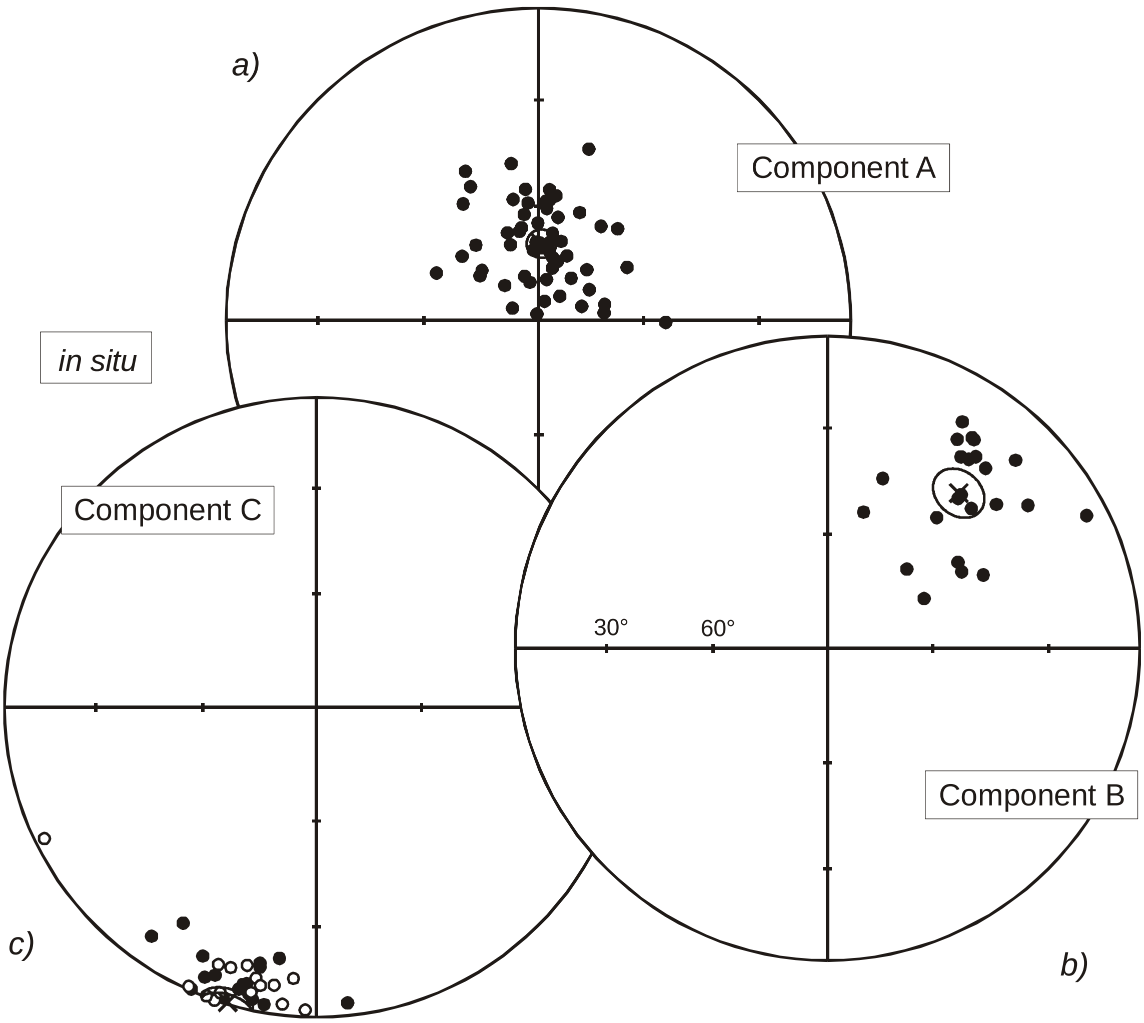



Figure 3-10: Equal area stereographic projection of thermal and alternating field demagnetisation results of samples of the Silurian Ockerkalk in geographic coordinates, in which straight line segments are identified. In (a) the component A, in (b) the component B and in (c) component C. Also shown is the cone of 95% confidence and the overall mean direction.

After removal the componets A, B and C a stable endpoint direction of dual polarity and maximum unblocking temperatures of 600°C can be isolated. The combined stable endpoint and great circle analysis of [*McFadden and McElhinny*, 1988] yield to a in situ site mean direction of 197°/08°, k=9.1 and α95=19.5 (Table 3-3, Fig. 3-11). Using the fold test of [*McElhinny*, 1964], the improvement in site mean grouping reaches a maximum after 90% unfolding which is significant at the 99% level of confidence (Fig. 3-11), resulting in an overall mean direction of 226°/38°, k=38.2 and α95=9.1 after bedding correction. These observations support the interpretation that this magnetisation is prefolding in age and closely linked to the age of deposition. The resulting palaeopole of 329°E, 8°S corresponds to a palaeolatitude of 21°S (+3.3°, -3,0°).

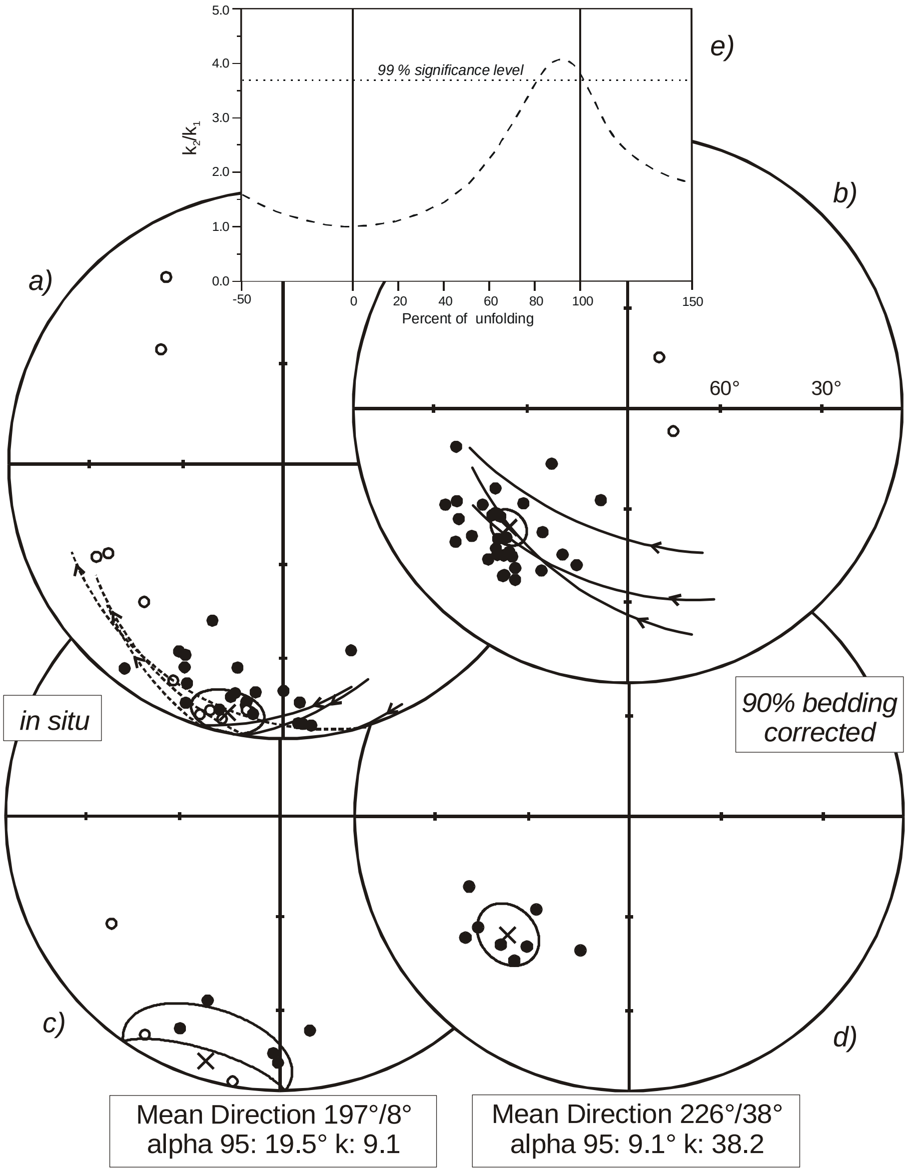



Figure 3-11: Equal area stereographic projection of thermal and alternating field demagnetisation results of single sample directions of the Silurian Ockerkalk of component D in geographic (a) and stratigraphic (b) coordinates (data from linear segments and stable endpoints), where solid (open) symbols represent lower (upper) hemisphere projection. In (c) and (d) the site mean directions are represented. Also shown is the cone of 95% confidence and the overall mean direction both before and after bedding correction. (e) The calculated k2/k1 ratio vs. percantage of unfolding.

# Chapter 4 Palaeozoic Palaeomagnetic results of the Alpine realm

## Chapter 4.1 Palaeozoic geography of the Alpine realm, new palaeomagnetic data from the Northern Greywacke Zone, Eastern Alps

INTRODUCTION

For early Palaeozoic times, the drift histories of the Iapetus-bordering continents, Baltica and Laurentia, and some Gondwana derived microcontinents such as Avalonia and the Armorica Terrane Assemblage (ATA), are now fairly well constrained (see chapters 1.2 and 3). The Palaeozoic units of Alpine Europe are generally accepted as having been part of the Northern margin of Gondwana throughout the Cambro-Ordovician (see chapter 1.4.2). By Permo-Carboniferous times Northern Gondwana had collided with Laurasia, forming Pangaea, but the palaeogeography affinities of the alpine units in intervening times, and whether they remained an integral part of Gondwana or formed part of the ATA, remains unclear. The tectonic relationship between the Proto-Alps and the ATA, however, is more enigmatic. In the lack of unequivocal evidence to the contrary, the Alps are traditionally considered to have remained adjacent to Gondwana. Recent palaeobiogeographic evidence, however, have brought this scenario into question and the hypothesis of an independent Proto-alpine terrane has been proposed [*Ziegler*, 1990] [*Schönlaub and Heinisch*, 1993] [*Neubauer and Frisch*, 1988] [*Neubauer*, 1988] [*Läufer et al.*, 2001]. The Cambro-Ordovician marks a phase of crustal extension and rifting along the northern margin of Gondwana and is documented in widespread sedimentary and volcanic sequences within the European Variscan fold belt [*Franke*, 1992]. Such basin-development can also be recognized in the Alpine region where these rocks form part of the basement units of the Alps [*Neubauer and Raumer*, 1993] [*v. Raumer*, 1998]. From late Ordovician to Late Carboniferous times continuous sedimentation in a passive margin-type environment took place [*Schönlaub*, 1992] [*Stampfli*, 1996] [*Frey et al.*, 1999] [*Neubauer et al.*, 1999]. Palaeozoic sequences in the Alpine realm (Noric-Bosnian terrane [*Neubauer and Raumer*, 1993]) document Ordovician crustal thinning and rifting, deposition in a back-arc basin environment, followed by development of a passive margin with shallow-water marine sedimentation, which persisted until Late Carboniferous [*Neubauer et al.*, 1999].

No reliable palaeomagnetic data are as yet available for the Palaeozic, however from lithological and faunal indicators [*Schönlaub*, 1992] a constant movement from steep southern latitudes in Late Ordovician to the equatoial belt during the Permian is supposed [*Schönlaub and Histon*, 2000]. In order to resolve this problem, a palaeomagnetic study of Late Silurian and Mid Devonian rocks from the Kitzbühel Alps of the Northern Greywacke Zone, Eastern Alps (part of the Norian-Bosnian terrane, or Proto-Alpine terrane), has been carried out.

## SAMPLING AND TECTONIC ENVIRONMENT

The Kitzbühel Alps belong to the Western part of the Northern Greywacke Zone which forms the Palaeozoic basement of the unconformably overlying Mesozoic sediments of the Northern Calcareous Alps. Both the Northern Greywacke Zone and its transgressive cover form part of the tectonic complex allochthonous Upper Austroalpine-Nappe complex [*Tollmann*, 1963] in the Eastern Alps (Fig. 4-1).

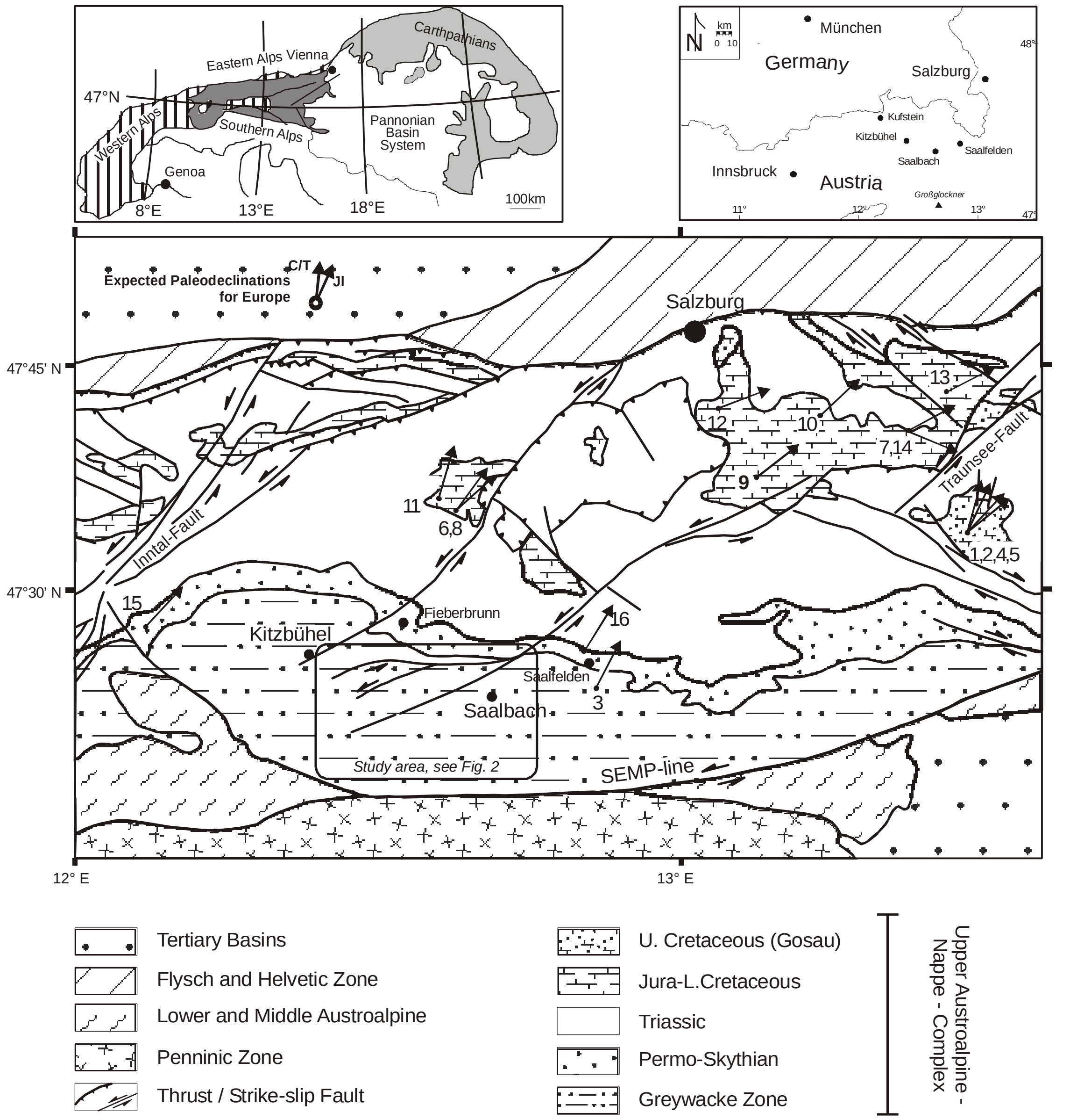



Figure 4-1: Simplified geological and structural map of the northern part of the central Eastern Alps, tectonic structure within the Northern Calcareous Alps after [*Linzer et al.*, 1995]. Numbered arrows indicate palaeomagnetic declinations according to the studies listed in Table 4-1.

Oblique convergence, due to the shortening of the Alpine units during the Cretaceous (eo-Alpine phase), was accompanied by partitioning of deformation by NW-W-vergent thrusting and folding with combined WNW trending strike-slip faults. As a result, the Austroalpine stack as a whole has been translated toward the West with clockwise rotation in the central Eastern Alps. This deformation is documented by numerous shear criteria in the Northern Calcareous Alps [*Linzer et al.*, 1995] and in the Eastern Greywacke Zone [*Ratschbacher*, 1986] [*Ratschbacher*, 1987] [*Ring et al.*, 1989] [*Ratschbacher and Neubauer*, 1989] [*Platt et al.*, 1989]. In the sampling area structures related to this transpressive deformation are exposed as subhorizontal thrusts with high angle transfer faults in the E-W striking Hochhörndler imbricate shear zone, and west vergent small scale folds with N-S trending fold-axes. These structures indicate west-vergent movement with the post-Variscan-transgressive series acting as a décollement zone [*Heinisch*, 1986] [*Wunderlich*, 1990] [*Meißner*, 1995]. Palaeomagnetic data for Mesozoic and Cenozoic sediments from the central Eastern Alps (Table 4-1, Fig. 4-1) show clockwise rotations of 40° to 60°, with respect to stable Europe [*Besse and Courtillot*, 1991]. The amount of rotation observed at the various localities conforms to the rotational pattern predicted by the tectonic model as outlined above. This is also valid for the rotational pattern identified in Permo-Triassic sandstones (Nos. 15 and 16 in Table 4-1) and remagnetised Devonian Dolomites (No. 3) from the Greywacke Zone. This suggests therefore that there has been little or no decoupling between the Palaeozoic basement and the cover rocks in the Upper Austroalpine Nappe Complex.

Table 4-1: Palaeomagnetic Directions for the Upper Austroalpine Nappe Complex

| No. | Rock Unit | Age | Dec (°) | Inc (°) | N | k | α (°) | Reference |
|---|---|---|---|---|---|---|---|---|
| 1 | Gosau-group, Elendgraben | Palaeocene | 51,3 | 33,6 | 31 | 7,5 | 10,1 | [*Mauritsch and Becke*, 1987] |
| 2 | Gosau-group, Elendgraben | Palaeocene | 191,4 | -51,9 | 20 | 3,7 | 19,8 | [*Mauritsch and Becke*, 1987] |
| 3 | Magnesite, Entachen | C/T boundary# | 22 | 75 | 32 | 431 | 5,9 | [*Mauritsch*, 1980] |
| 4 | Gosausediments, Elendgraben | U. Cretaceous | 222,2 | -43,9 | 31 | 7,6 | 10,1 | [*Mauritsch and Becke*, 1987] |
| 5 | Gosausediments, Gosau | U. Cretaceous | 23,2 | 51,2 | 4 | 37,2 | 8 | [*Becke and Mauritsch*, 1983] |
| 6 | Radiolarite, Lofer | Dogger/Malm | 36,7 | 47,8 | | 100 | 5,5 | [*Hargraves and Fischer*, 1959] |
| 7 | Limestone, Radiolarite | Dogger/Malm | 62 | 45 | 8 | 29,1 | 11,3 | [*Mauritsch and Frisch*, 1978] |
| 8 | Adnet Limestones, Lofer | Liassic | 47,9 | 50,6 | | 70,7 | 6,5 | [*Hargraves and Fischer*, 1959] |
| 9 | Adnet Limestones, Golling | Liassic | 52,6 | 27,2 | 8 | 102,9 | 4,5 | [*Heer*, 1982] |
| 10 | Adnet Limestones, Hintersee | Liassic | 47,1 | 25,3 | 4 | 92 | 9,5 | [*Heer*, 1982] |
| 11 | Adnet Limestone, Lofer Area | Liassic | 15,1 | 55 | | 71,2 | 6,6 | [*Channell et al.*, 1990] |
| 12 | Adnet Limestone, Adnet Area | Liassic | 80,8 | 57,1 | | 63,9 | 15,5 | [*Channell et al.*, 1990] |
| 13 | Adnet Limestones, Wolfgangsee | Liassic | 61,5 | 61,3 | 105 | 84,8 | 6 | [*Channell et al.*, 1992a] |
| 14 | Rhaetian Kössen-Fm., Osterhorn | Rhaetian | 115 | 55,1 | 2 | | | [*Mauritsch and Frisch*, 1978] |
| 15 | Red Sandstones, Woergl | Permo-Scythian | 44,3 | 59,2 | 30 | 102,7 | 6,9 | [*Soffel*, 1979] |
| 16 | Red Sandstones, Saalfelden | Permo-Scythian | 31,5 | 19,6 | 24 | 414,8 | 4 | [*Soffel*, 1979] |

Dec/Inc, declination and inclination in degrees; N, number of samples used in calculation of palaeomagentic direction; k, precision parameter after [*Fisher*, 1953]; $\alpha_{95}$ , semi-cone of 95% confidence; # magnetization age

Recent mapping of the Kitzbühel-Saalbach area (Fig. 4-2) demonstrate the presence of two distinct facies associations, the Wildseeloder Unit and the Glemmtal Unit [*Heinisch*, 1986]

[*Heinisch*, 1988] [*Heinisch et al.*, 1987] [*Schlaegel*, 1988]. These units are separated by the complex Hochhörndler imbricate shear zone (Fig. 4-2) [*Heinisch*, 1988].

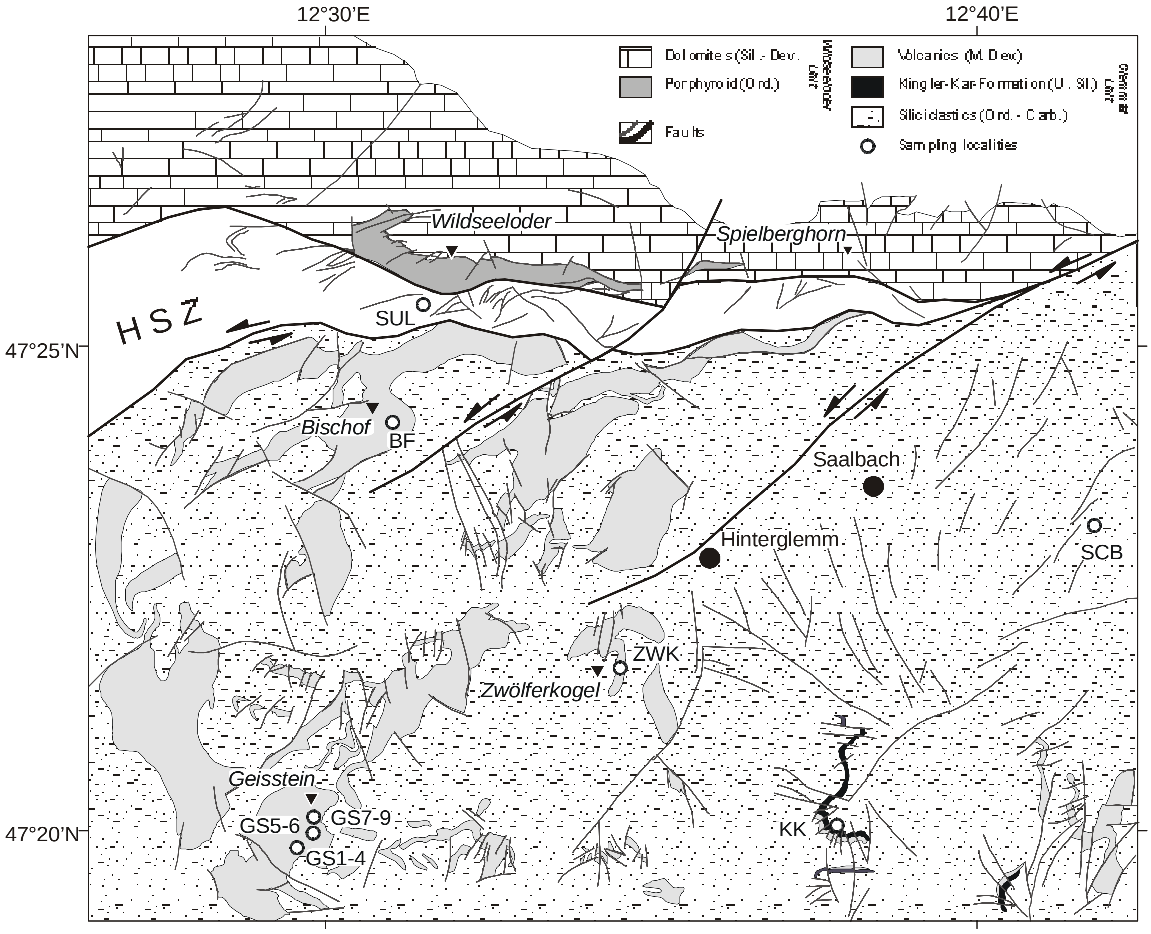



Figure 4-2: Geological sketch map of the sampling area. HSZ – Hochhörndler imbricate zone; BF, GS, ZWK, SUL, SCB, KK - sampling localities.

In this present study a total of 27 sites were collected from volcanic, intrusive and sedimentary rocks, which have not undergone penetrative deformation and at most very low grade metamorphism (Fig. 4-2). The Klingler-Kar Formation was sampled at only one locality, where 6 sites of gently tilted marls and limestones (sites KK1 – KK6) of Ludlovian and Pridolian in age were collected (Table 4-3). The sites are situated in several blocks which are bounded by vertical faults, and variations in strike of up to 40° was observed, which may be indicative for differential vertical axis rotations of these rocks. Mid Devonian basaltic sequences were sampled at three localities (Fig. 4-2 and Table4-4). Pillow lavas were sampled near the top of Bischof (sites BF1 – BF4); pillows lavas, flows and tuffs were collected from the upper slopes of Geisstein (sites GS1 – GS9), and samples of gabbroic rocks were collected from Zwölferkogel (sites ZWK1 – ZWK2). Due to the almost undeformed structure of the pillows it was possible to identify way-up structures and several interbedded and overlying tuff layers allow for accurate structural control of the sampling sites.

Post-Palaeozoic rocks are missing in this region. Therefore, to control younger (Alpine) rotation and deformation, samples were also collected from more deformed rocks where a strong Alpine (magnetic) overprint would be expected. Samples were collected in two different localities, from fine-grained sediments in the eastern part of the sampling area (sites SCB1 – SCB2), and from the Hochhörndler imbricate zone (sites SUL1 – SUL4, Table 4-2). The latter represent a para-conglomerate, where samples from both the silty matrix and a number of boulder clasts, composed of sandstone, were collected.

## PALAEOMAGNETIC RESULTS

The natural remanent magnetization (NRM) intensities are generally low and range from 0.5 – 5 mA/m for both igneous and sedimentary rocks. Most samples show multivectorial behaviour during stepwise thermal demagnetisation experiments, often with overlapping unblocking temperature ($T_{ub}$) spectra at low to intermediate temperatures. At high $T_{ub}$ stable endpoint directions can be identified in orthogonal projection of the results and a total of three different high temperature directions of magnetization, termed B, C and D can be identified in the samples, often after removal of a low temperature component A which is similar to the local present-day Earth's magnetic field and is interpreted to be an overprint of recent origin. The characteristic of the three higher temperature magnetization directions are described below.

### Deformed Clastic Sedimentary Rocks

Palaeozoic clastic sediments, which were deformed during the Alpine orogeny, were collected from the vertically dipping Hochhörndler imbricate zone (conglomerates, SUL1-4) and the Schrambachgraben (fine grained sandstones, SCB1-2) in the east of the study area (Fig. 4-2, Table 4-2).

Table 4-2: Site mean directions for conglomerate and clastics - component B

| | | | | | Component B-In Situ | | |
|---|---|---|---|---|---|---|---|
| Site | Type | Lithology | Bedding | N/n | Dec/Inc (°) | k | $\alpha_{95}$ (°) |
| | conglomerate | | | | | | |
| SUL 1 | clast 1 | sandstone | 178/87 | 6 / 4 | 052/51 | 35.1 | 10.3 |
| SUL 1 | clast 2 | sandstone | 178/87 | 4 / 2 | 050/55 | - | - |
| SUL 1 | clast 3 | sandstone | 178/87 | 5 / 4 | 039/49 | 31.7 | 46.0 |
| SUL 2 | matrix | siltstone | 178/87 | 8 / 4 | 054/48 | 12.6 | 37.7 |
| SUL 3 | clast 4 | sandstone | 178/87 | 4 / 2 | 068/52 | - | - |
| SUL 3 | clast 5 | sandstone | 178/87 | 5 / 2 | 075/15 | - | - |
| SUL 3 | clast 6 | sandstone | 178/87 | 6 / 4 | 083/20 | 13.2 | 26.3 |
| SUL 4 | matrix | siltstone | 178/87 | 6 / 3 | 085/63 | 15.4 | 15.9 |
| | | | mean | | 065/46 | 11.5 | 17.1 |
| | clastic sediments | | | | | | |
| SCB 1 | clastics. | silt-, sandstone | 275/10 | 8 / 4 | 060/55 | 15.3 | 22.1 |
| SCB 2 | clastics. | silt-, sandstone | 275/10 | 7 / 3 | 051/34 | 11.4 | 28.3 |
| | | | mean | | 054/47 | - | - |

N, number of samples measured; n, number of samples used in calculation of site mean; Dec/Inc, declination/inclination of site means in degree; k, precision parameter [*Fisher*, 1953]; $\alpha_{95}$ , half angle of the cone of 95% confidence [*Fisher*, 1953]; U. Sil. – Upper Silurian. Bedding correction is given in terms of direction of dip/dip in degrees. $^{\#}$ sites not used in calculation of the respective overall means.

The sample matrix contains a high proportion of phyllosilicates, which are probably related to the penetrative deformation that the sediments have undergone. These minerals also tend to undergo mineralogical alteration during thermal treatment, but due to their paramagnetic characteristics alternating field methods were unsuccessful. Nevertheless in some samples stable endpoint directions could be determined with unblocking temperatures of up to 550°C. This component B can be identified in seven samples from sites SCB1 and SCB2. Grouping of the directions is rather poor, but all have northeasterly declinations and positive inclination values (Fig. 4-3a; Table 4-2).

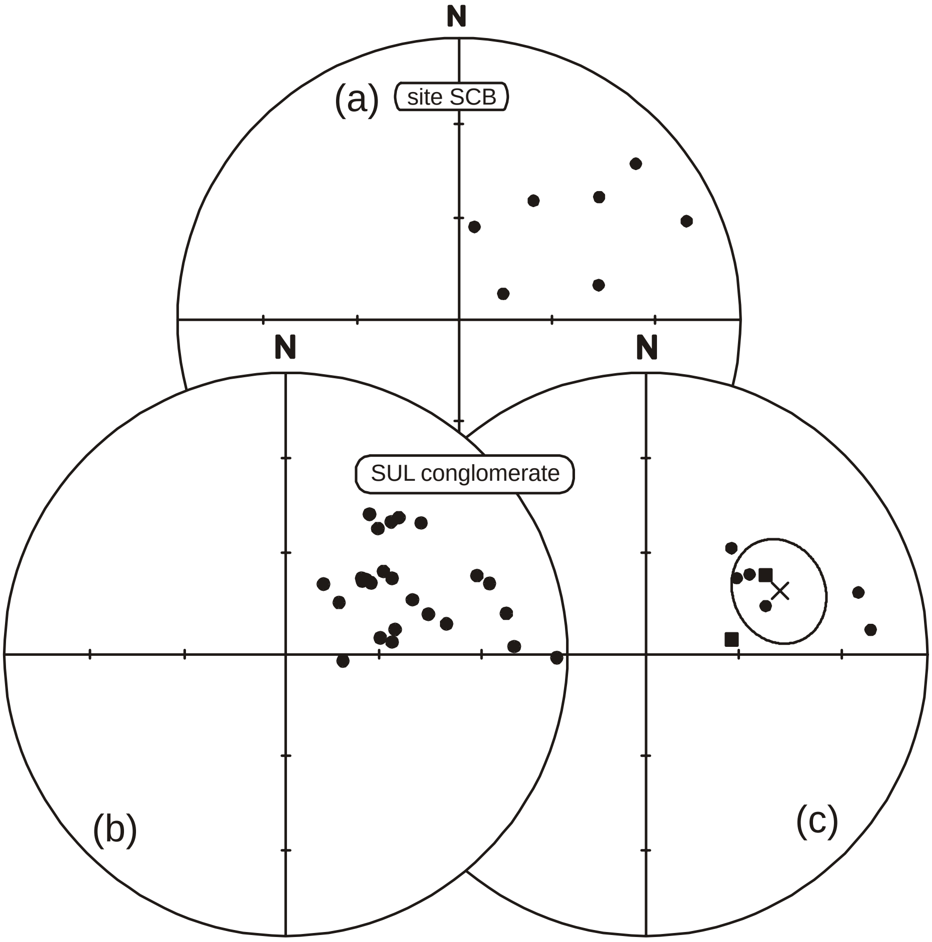



Figure 4-3: Equal-area lower-hemisphere stereographic projection of in situ (a) sample directions from locality SCB, and (b) SUL-conglomerate. In (c) the mean directions of 6 clasts (solid circles) and 2 site mean directions from the matrix (solid squares) of the conglomerate are shown in situ, with the cone of 95% confidence and the overall mean direction for component B.

Samples from the clasts and from the matrix of the conglomerate (sites SUL1-4) show similar behaviour during demagnetization, revealing northeasterly positive directions of magnetisation (Figs 4-3b and 4-3c). Directions identified in six clasts and two sites of the matrix are similar in situ, and, therefore, the samples fail the conglomerate test (Fig. 4-3b, Table, 4-2). Combining the data from localities SCB and SUL yields an overall mean direction of 065°/46° (Dec/Inc), k = 15.5 , $\alpha_{95}$ = 14.5° in situ (Table 4-2). Although not particularly well-defined, the component B identified in these samples is clearly of secondary in origin, while the in situ direction is close to post-Variscan directions identified elsewhere in this region (see Table 4-1, Fig. 4-1).

**Upper Silurian Carbonate Rocks, Klingler Kar Formation**

Late Silurian samples were collected from six sites in the Klingler-Kar Formation (sites KK1-KK6). Natural remanent magnetisations are characteristically low in intensity, and during thermal demagnetization experiments samples from KK1 show inconsistent and unstable directions of magnetisation. The results from this site will not be discussed further. For sites KK2–KK6, demagnetization up to 240°C results in removal of a low temperature component (labeled component A) followed by the removal of a second component (B) at intermediate temperatures. In some samples (Fig. 4-4d) component B can be clearly identified as a north-easterly intermediate direction with unblocking temperatures up to 450°C. In most samples, however, the separation of components A and B is difficult due to overlapping unblocking temperature spectra (Fig. 4-4). At higher temperatures and up to the maximum $T_{UB}$ in these rocks (between 500°C and 550°C) a third stable endpoint direction is determined.

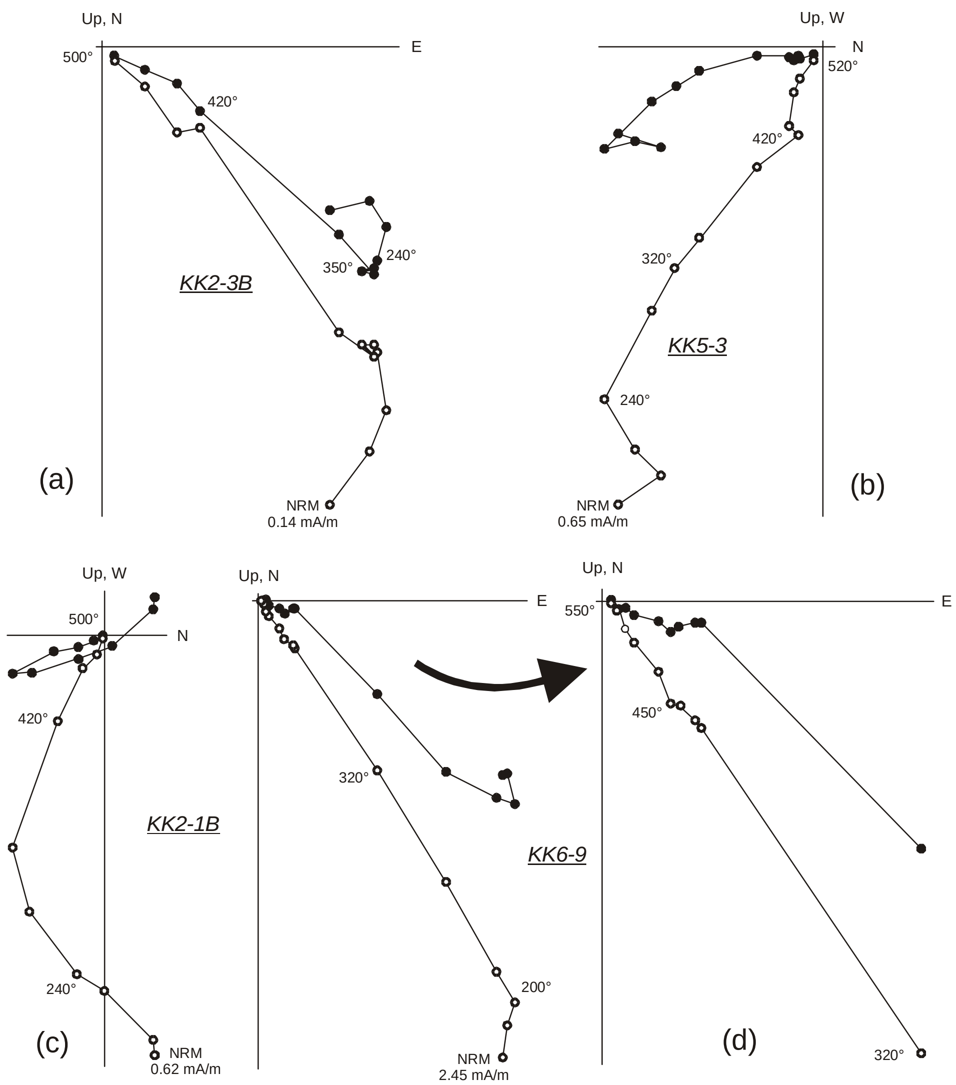



Figure 4-4: Orthogonal projection of typical thermal demagnetization results for the Upper Silurian sediments. All directions are plotted in situ, and solid (open) symbols represent the horizontal (vertical) component respectively. Temperatures are given in degrees Celsius. All samples show a removal of A up to 240°C, removal of B up to 450° is obvious in sample (d). At temperatures higher than 420°-450°, C is identified as intermediate south-southeastward direction of magnetization.

The overall mean direction obtained for this magnetization (component C) is 134°/59°, k = 110.4, $\alpha_{95}$ = 7.3° in situ (5 sites, 24 samples). Application of the bedding correction, does not improve the grouping of directions. However, the bedding corrected directions form a small circle distribution with steep inclination values (Fig. 4-5a,b and Table 4-3). This distribution is considered to reflect small scale block rotations of the sampling locations.

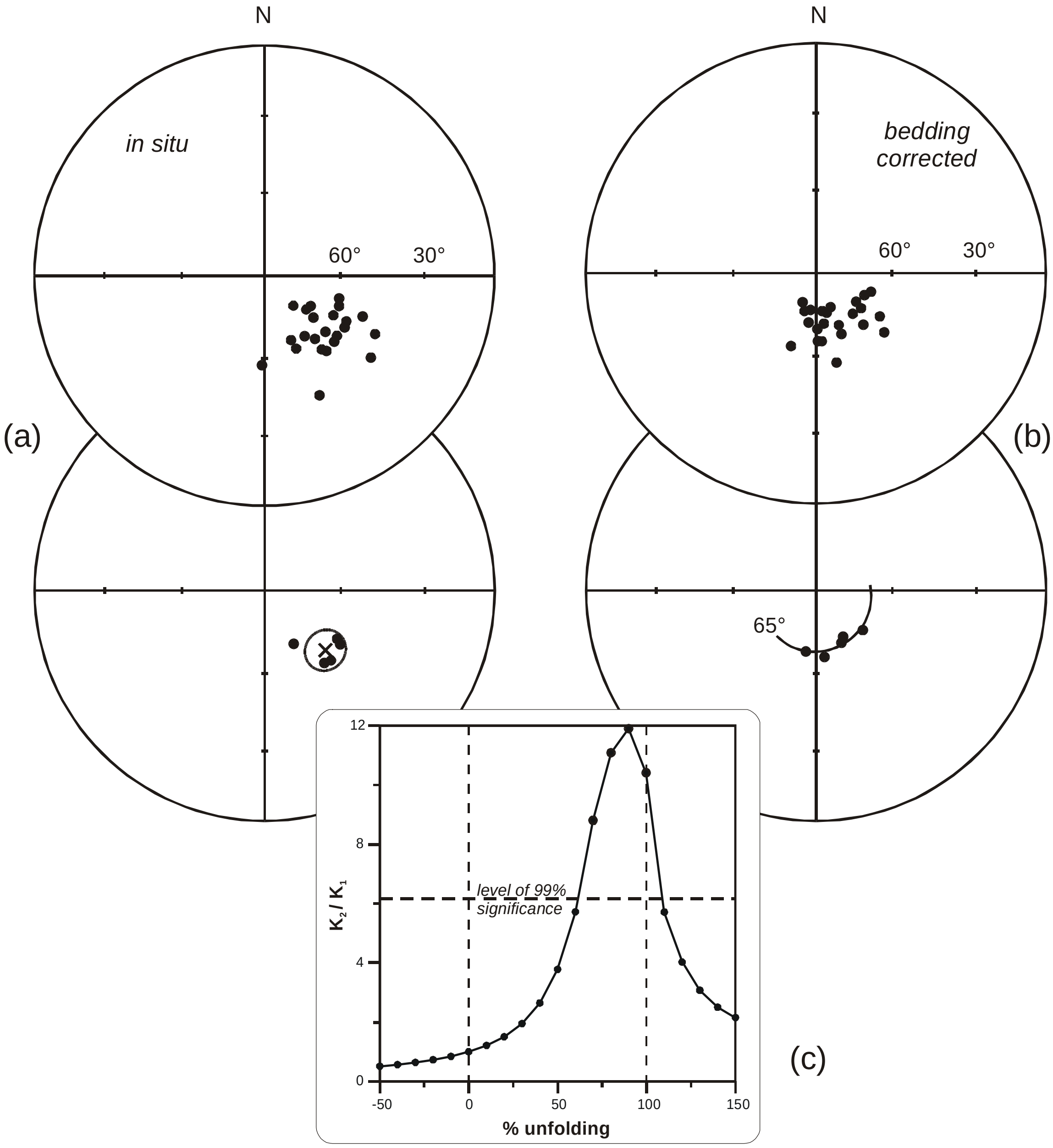



Figure 4-5: Equal area projection of the (a) in situ sample directions (above) and site mean directions (below), and (b) bedding corrected sample directions (above) and site mean directions (below) for component C along a small circle segment. (c) The calculated k2/k1 ratio vs percentage of unfolding applying the inclination-only test. Notation as in Fig. 3.

Applying an inclination only fold test [*Enkin and Watson*, 1996] results in a maximum k (the precision parameter of [*Fisher*, 1953]) at 90% unfolding (Fig. 4-5c). Since there is no significant difference in inclination at 90% unfolding and 100%, we assume the primary origin of this magnetization and ascribe the fact that maximum k is reached at 90% to small internal variations in bedding attitude. Thus the fold test for the sites of the Klingler-Kar Formation is positive at the 99% confidence level [*McElhinny*, 1964] and amounts to an overall inclination value of +65° with $\alpha_{95}$ = 1.7°.

Table 4-3: Site mean directions for the Upper Silurian Carbonate rocks, Klingler Kar Formation, component C

| | | | | | Component C | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|
| | | | | | In Situ | | | 90% Bedding Corrected | | |
| Site | Age | Lithology | Bedding | N/n | Dec/Inc (°) | k | $\alpha_{95}$ (°) | Dec/Inc (°) | k | $\alpha_{95}$ (°) |
| KK 1# | U. Silur | limestone, tuffite | 265/07 | 7 / 0 | -/- | - | - | -/- | - | - |
| KK 2 | U. Silur | black limestone | 288/10 | 6 / 5 | 124/59 | 40.7 | 12.1 | 130/65 | 40.7 | 12.1 |
| KK 3 | U. Silur | limestone, marl | 275/20 | 5 / 5 | 140/56 | 26.5 | 15.2 | 173/64 | 26.5 | 15.2 |
| KK 4 | U. Silur | limestone, marl | 259/15 | 6 / 4 | 151/68 | 40.5 | 14.6 | 190/65 | 40.5 | 14.6 |
| KK 5 | U. Silur | limestone, marl | 300/17 | 7 / 5 | 136/55 | 52.5 | 10.7 | 150/68 | 52.5 | 10.7 |
| KK 6 | U. Silur | limestone, marl | 275/20 | 9 / 5 | 125/56 | 110.6 | 7.3 | 152/67 | 104.6 | 7.5 |
| | Mean inclination | | | | 59° | 79.4 | 6.1 | 65° | 876.2 | 1.7 |

Notation as in Table 4-2.

**Middle Devonian Volcanic Rocks, Geisstein, Zwölferkogel and Bisschof**

Most samples collected from Mid Devonian basaltic magmatic rocks (localities BF, ZWK and GS) show similar behaviour during stepwise thermal demagnetization. All trajectories of the NRM trend to have northerly intermediate directions (component A), similar to the present day field direction and which occasionally persist up to 275°C (Fig. 4-6). Some samples from localities GS and ZWK then show removal of a northeasterly component, component B, up to temperatures of 400°C (Fig. 4-6b,c). The third high temperature direction of magnetization (component D) identified in these samples is isolated at temperatures above 400°C and is generally directed towards the origin in orthogonal projection (Fig. 4-6a - d). Samples from locality BF show great circle behaviour during removal of component B, but at higher temperatures, stable endpoint directions can be identified which persisting until complete demagnetization (Fig. 4-6e – g).

Site mean directions for this high temperature D component from all three localities are poorly grouped in situ (Table 4-4, Fig. 4-7a). After bedding correction, however, the directions for localities GS and BF show a marked improvement in grouping and the directions for each locality pass the fold test of [*McFadden*, 1990]. Samples collected from ZWK are from beds of uniform dip so no fold test is possible. The combined declination values for all sites are widely scattered and plot on a small circle, i.e., the inclination values converge towards a common mean value (Fig. 4-7b).

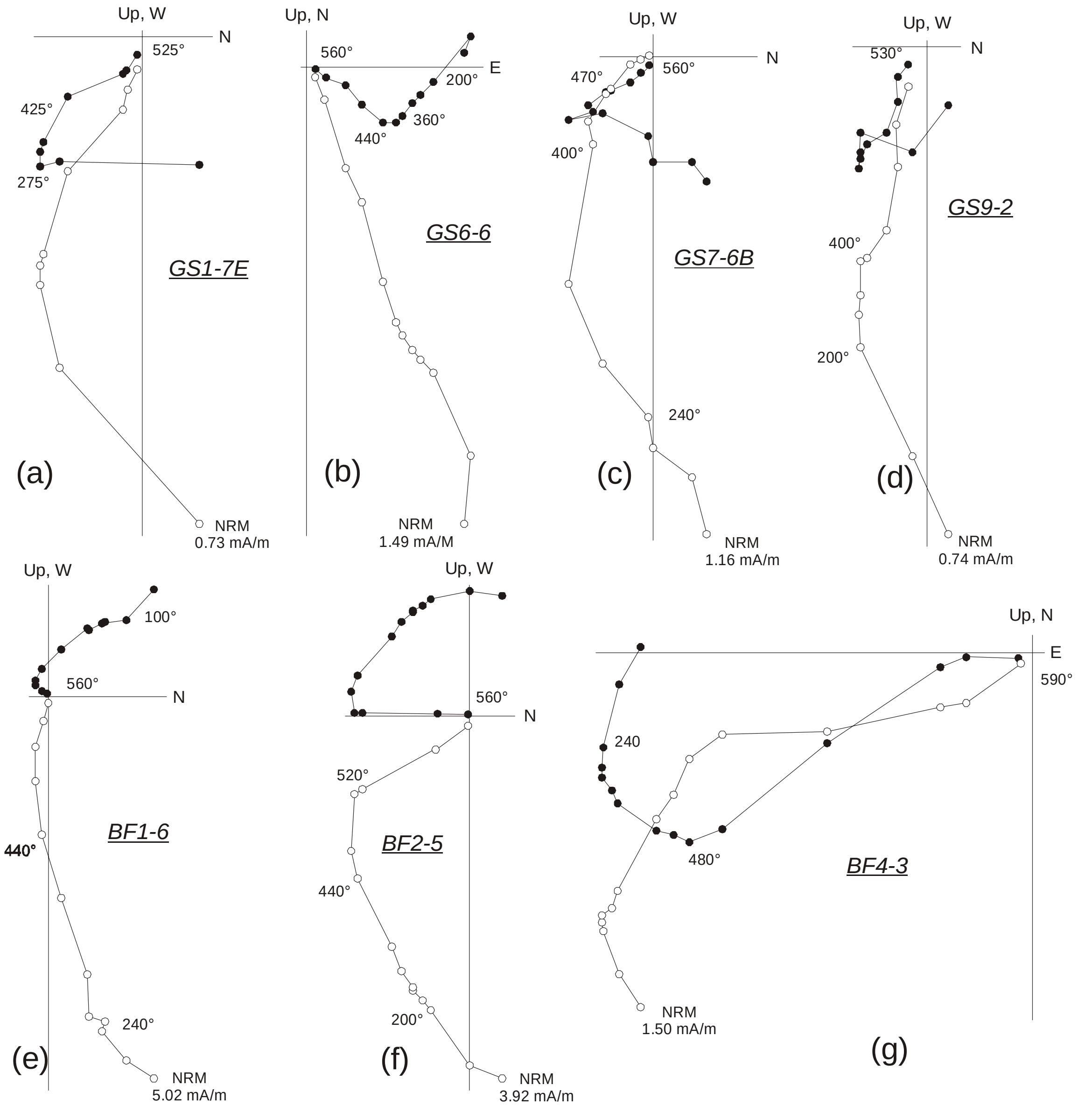



Figure 4-6: Orthogonal projection of thermal demagnetization data for samples from the Mid Devonian volcanics. All directions are plotted in in situ coordinates. Notation as in Fig. 4-4. Samples (a), (d), (e), (f), (g) show a removal of component A up to 200°-275°C. At intermediate temperatures up to 400°-440°C component B is removed in (b) and (c). At temperatures higher than 400°-480°C, the samples reveal the component D.

Using the inclination-only fold test [*Enkin and Watson*, 1996] results in an increase in k from 24.71 (in situ) to 53.67 (100% bedding corrected), yielding a positive inclination-only fold test at the 95% confidence level (Fig. 4-7c) [*McElhinny*, 1964]. The resulting overall mean inclination value is 42°, $\alpha_{95}$ = 4.6°. The positive fold test within localities GS and BF, the small circle distribution of combined localities is thought to represent differential vertical axis rotations between the sampling areas and component D is considered to be primary in origin.

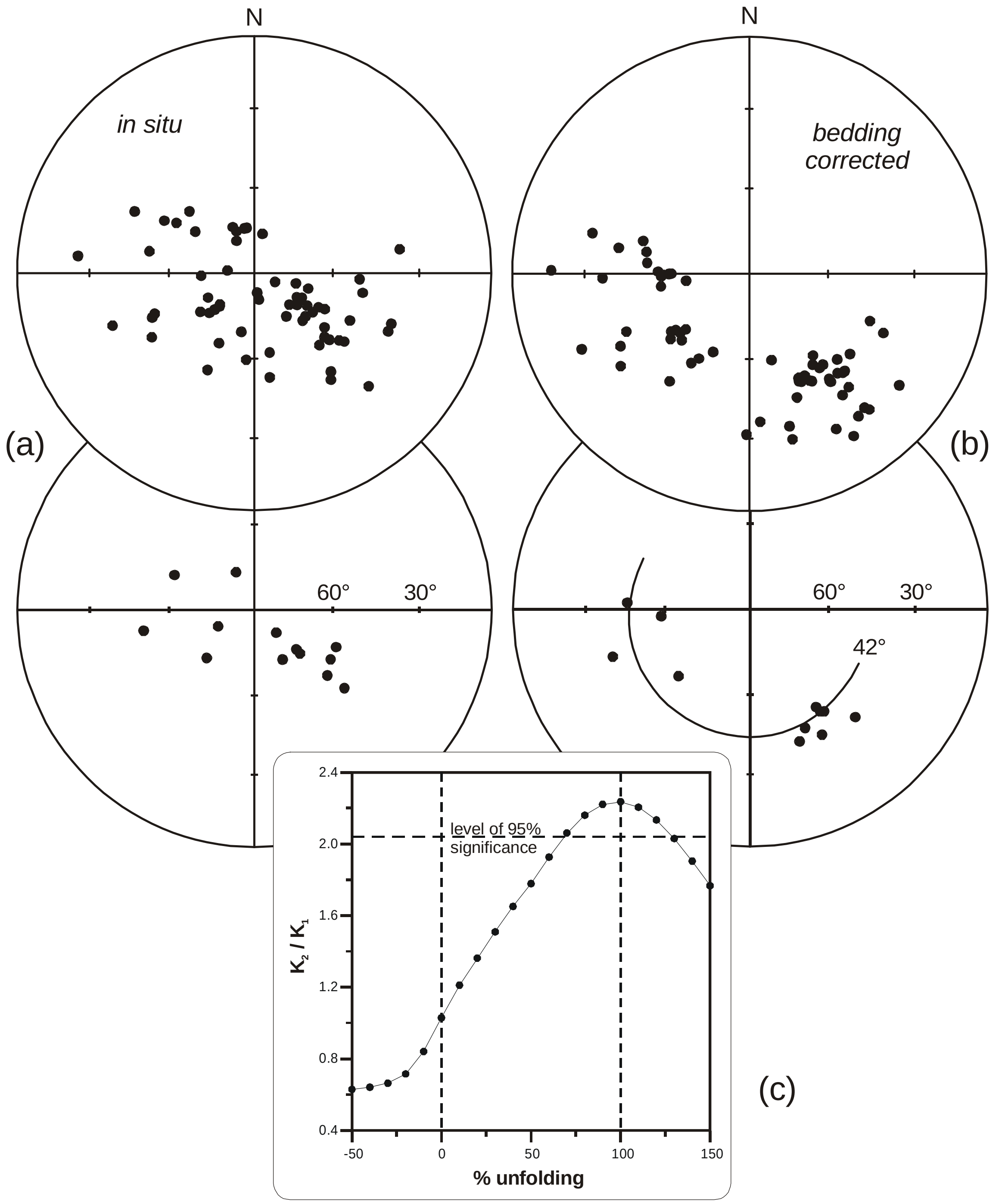



Figure 4-7: Equal area projection of the (a) in situ sample directions (above) and site mean directions (below), and (b) bedding corrected sample directions (above) and site mean directions (below) for component D along a small circle segment. (c) The calculated k2/k1 ratio vs percentage of unfolding applying the inclination-only test. Notation as in Fig. 4-3.

Table 4-4: Site mean directions for Mid Devonian volcanic rocks, component D

| | | | | | Component D | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|
| | | | | | In Situ | | | Bedding Corrected | | |
| Site | Age | Lithology | Bedding | N/n | Dec/Inc (°) | k | $\alpha_{95}$ (°) | Dec/Inc (°) | k | $\alpha_{95}$ (°) |
| GS 1 | M. Dev. | pillow lava | 150/32 | 8 / 6 | 151/70 | 21.2 | 14.9 | 150/38 | 21.2 | 14.9 |
| GS 2 | M. Dev. | pillow lava | 150/32 | 6 / 5 | 136/79 | 36.6 | 12.8 | 146/48 | 32.7 | 13.6 |
| GS 3# | M. Dev. | Lapilli tuff | 150/32 | 3 / 2 | 132/56 | - | - | 139/24 | - | - |
| GS 4# | M. Dev. | Tuff | 150/32 | 2 / 0 | -/- | - | - | -/- | - | - |
| GS 5# | M. Dev. | pillow lava | 177/30 | 9 / 3 | 246/76 | - | - | 199/53 | - | - |
| GS 6 | M. Dev. | sheet flow | 190/26 | 7 / 5 | 115/59 | 12.3 | 22.7 | 144/45 | 12.3 | 22.7 |
| GS 7 | M. Dev. | pillow lava | 220/30 | 6 / 6 | 131/48 | 23.2 | 14.2 | 160/39 | 23.2 | 14.2 |
| GS 8# | M. Dev. | pillow lava | 175/28 | 8 / 0 | -/- | - | - | -/- | - | - |
| GS 9 | M. Dev. | pillow lava | 175/28 | 7 / 7 | 134/68 | 15.4 | 15.9 | 155/43 | 15.4 | 15.9 |
| Mean | | | five sites | | 131/65 | 38.1 | 12.6 | 151/43 | 169.3 | 5.9 |
| ZWK 1 | M. Dev. | Gabbro | 157/25 | 6 / 4 | 123/58 | 22.2 | 19.9 | 136/36 | 22.2 | 19.9 |
| ZWK 2 | M. Dev. | Gabbro | 157/25 | 4 / 3 | 133/70 | 390.6 | 6.2 | 146/46 | 390.6 | 6.2 |
| Mean | | | two sites | | 127/64 | - | - | 140/41 | - | - |
| BF 1 | M. Dev. | pillow lava | 233/20 | 6 / 6 | 294/59 | 23.8 | 14.0 | 273/46 | 23.8 | 14.0 |
| BF 2 | M. Dev. | pillow lava | 232/11 | 7 / 6 | 225/66 | 53.7 | 9.2 | 227/55 | 53.7 | 9.2 |
| BF 3 | M. Dev. | pillow lava | 237/39 | 8 / 8 | 334/75 | 37.1 | 9.2 | 266/59 | 37.1 | 9.2 |
| BF 4 | M. Dev. | pillow lava | 220/15 | 9 / 6 | 259/50 | 13.1 | 19.2 | 251/38 | 13.1 | 19.2 |
| Mean | | | four sites | | 271/67 | 16.2 | 23.6 | 255/51 | 26.7 | 18.1 |
| | Mean inclination | | | | 65° | 24.7 | 15.5 | 42° | 53.7 | 4.6 |

Notation as in Table 4-2. M. Dev. – Middle Devonian

## ROCK MAGNETIC STUDIES

To characterize the carriers of magnetization several rock magnetic experiments were carried out on the Silurian sedimentary and the Devonian igneous rocks. Isothermal remanent magnetization (IRM) acquisition studies of the Silurian (Fig. 4-8a) indicate saturation typically at about 500 mT. Thermal demagnetization of samples which have been given three orthogonal IRMs at fields of 1.5 T, 0.5 T, and 0.25 T show the presence of low and intermediate coercivity minerals. The low coercivity phase has an unblocking temperature range of 550°-580°C and is interpreted as being magnetite. It coexists with varying amounts of iron sulphides as shown by the intermediate coercivity mineral which has unblocking temperatures of 300 - 350°C.

In the basaltic samples magnetic saturation is reached at approximately 200 mT. A few samples also contain a higher coercivity phase which is not saturated by the maximum applied field of 1.5 T (Fig.4-8b). During thermal demagnetization these samples show an abrupt drop in intensity at about 100°C, thus indicating the presence of minor amounts of goethite. The intensity (of the soft coercivity fraction) then gradually decreases until final demagnetization at about 580°C, demonstrating the predominance of magnetite as the remanence carrier. The intensity decay curves of the intermediate coercivity fraction of some samples also show a clear inflection at 300° – 350°C, indicating a subordinate content of pyrrhotite and, or maghemite.

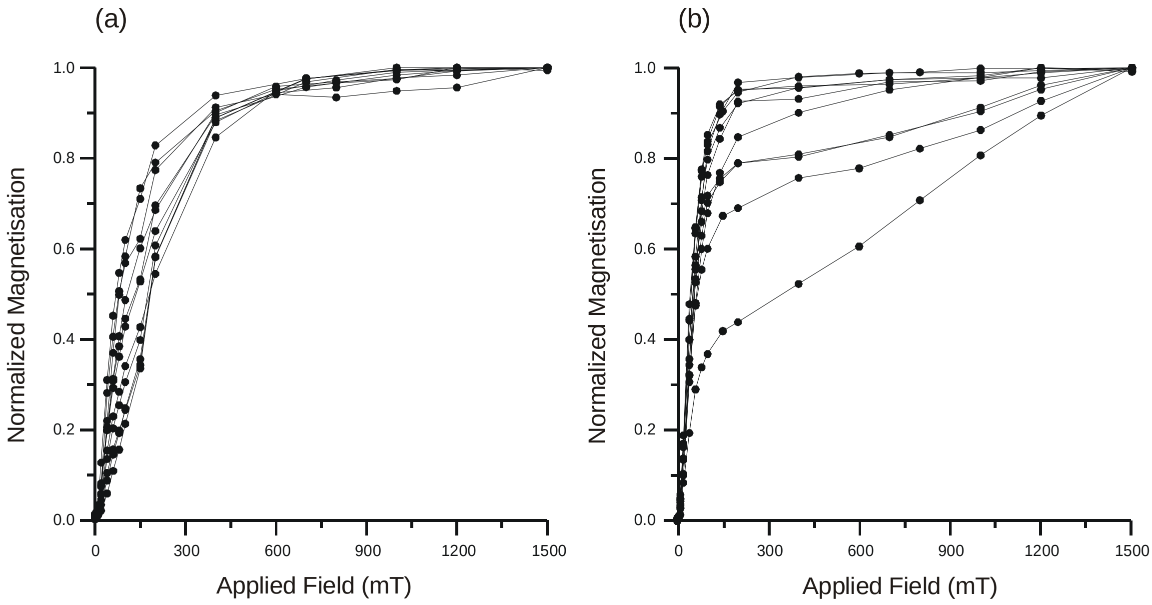



Figure 4-8: Normalized acquisition curves of a monodirectional isothermal remanent magnetization (IRM) in (a) sediments of the Klingler-Kar-Formation and (b) volcanics of the localities of Geisstein (GS) and Bischof (BF).

The hysteresis parameters $M_{rs}/M_s$ and $H_{cr}/H_c$ obtained plot predominantly in the PSD field of the [*Day et al.*, 1977] diagram, or tend to higher $H_{cr}/H_c$ ratios. This behaviour is typical for specimens with mixed magneto-mineralogy and again demonstrates the influence of the softer coercivity fraction as the magnetic carrier. In summary, these rock magnetic characteristics are in agreement with the NRM demagnetisation experiments, showing the dominance of fine-grained magnetite as the main remanence carrier and carrying the high temperature C and D magnetisation directions. Iron sulphides and/or occasionally goethite which has low unblocking temperatures, are also present and, at least in part, are the carriers of the secondary A and B components.

To establish the possible influence of structural deformation on the magnetization directions, the anisotropy of magnetic susceptibility (AMS) of the volcanic rocks was measured (Fig. 4-9). The susceptibilities of the sedimentary rocks are, in general, too low to yield accurate AMS measurements. The magnetic fabric is determined from the principal axes (K1 = maximum axis, K2 = intermediate, K3 = minimum) and the shape of AMS ellipsoids used as a proxy for the petrofabric which allows recognition of any penetrative deformation within the rocks [*Tarling and Hrouda*, 1993]. The magnetic fabric was characterized by applying the conventional parameters, $P_J$ (corrected anisotropy degree) and the shape parameter T after [*Jelinek*, 1981]. The average anisotropy degree is about 4.5%. In comparison to the average degree of anisotropy in recent basaltic flows of up to 10% [*Tarling and Hrouda*, 1993], it is clear that in these samples there is no identifiable preferred shape of the anisotropy ellipsoid which could be related to stress and thus affect the direction of the magnetisation. As would be expected for undeformed rocks, the magnetic foliation is approximately parallel to the flow plane measured in the field [*Hrouda*, 1982].

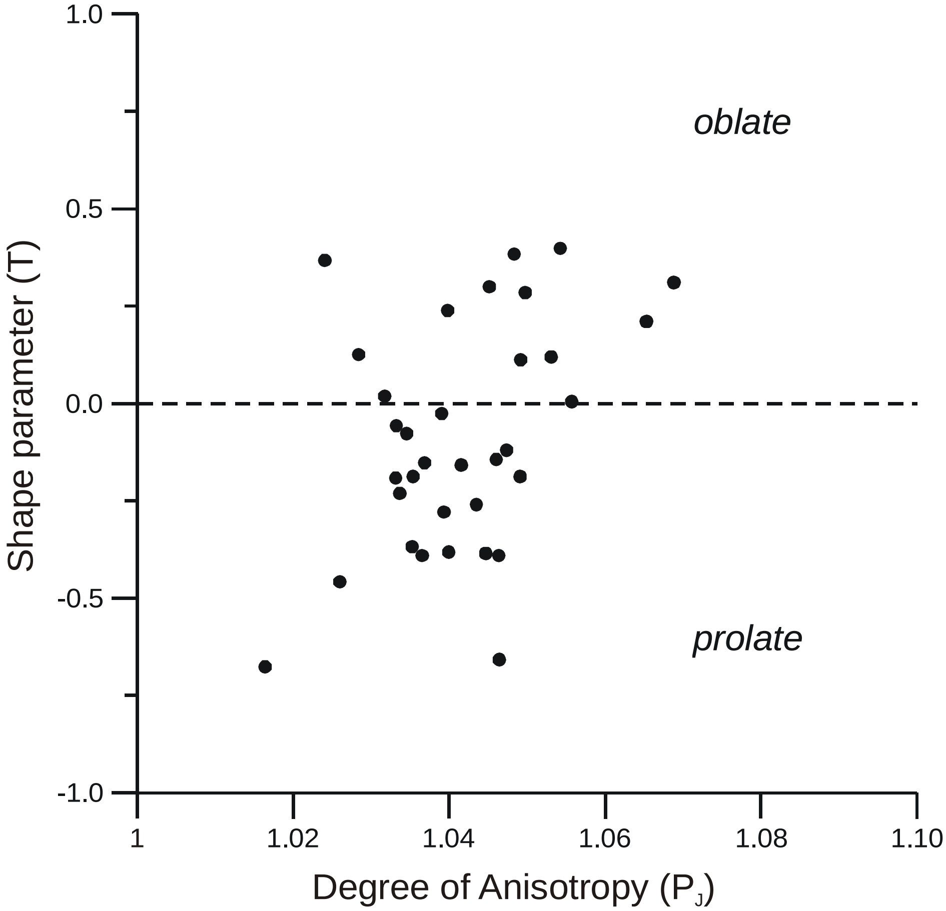



Figure 4-9: AMS fabric diagram for volcanics from GS and BF. The dashed line (T = 0) represent an isotropic shape of the anisotropy ellipsoid.

INTERPRETATION

Of the three palaeomagnetic directions identified, component B is present predominantly in the penetratively deformed rock sequences (SUL, SCB), and at low-intermediate unblocking temperature intervals in some samples from the localities KK and GS. This secondary magnetization is similar to magnetization directions previously identified in Mesozoic-Cenozoic and remagnetised Palaeozoic rocks in this region (Table 4-1) and is thus considered to be related to an alpine overprint of Cretaceous to Tertiary age. Consequently, these results support the model that Alpine rotations have also affected the Palaeozoic basement rocks, and the rotations identified must be taken into account when considering the older, primary components C and D. Surprisingly, no indication for any remagnetisation event of Permo-Carboniferous age was observed in any of the measured samples.

Components C, identified in the Silurian sediments (Table 4-3), and D, in the Middle Devonian basalts (Table 4-4), are identified in a high temperature range. Both C and D pass the fold test within locality, and the inclination-only fold test when directions from the different localities for each age group are compared. The C magnetisation can be reliably identified in five sites (24 samples), it yields an overall mean inclination of +65° after bedding correction. Component D, identified in eleven sites (62 samples), yields an overall mean inclination of +42° after bedding correction. Both C and D are interpreted to be primary in origin and to be Silurian (C) and Devonian (D) in age. Opting for the reversed polarity option, these inclination values translate into palaeolatitudes for the sampling area of 47° South (+2.3° and –2.1°, upper and lower 95% confidence level) in Upper Silurian times and 25° South (+3.7° and –3.4°) in the Mid Devonian.

## Chapter 4.2 Palaeomagnetic results from lower Devonian sediments of the southern Alps and their palaeogeographic implications

INTRODUCTION

As already discussed, central and southern Variscan Europe represent a collage formed by several suspect terranes and microplates which show a different drift history during the Palaeozoic. According to palaeomagnetic data from the Northern Greywacke Zone discussed in this thesis, the Proto-Alps underwent a drift history independent of the ATA and the margin of Northern Gondwana in the Mid Palaeozoic. In order to further constrain the Proto-Alps studies were extended to the Carnic Alps.

The Alpine fold belt of Southern Europe contains a number of Palaeozoic sequences, which form a mosaic-like pattern of isolated units incorporated into the nappe system of the Eastern Alps. Palaeozoic basement units in the Eastern Alps are largely exposed in the Austroalpine, Penninic and Southalpine units with complex internal tectonic structures, which orginated during Cretaceous and Cenozoic tectonic processes [*Neubauer and Handler*, 2000]. The Alpine metamorphic events range from very low grade to eclogite-grade conditions [*Frey et al.*, 1999] [*Hoinkes et al.*, 1999]. Within the Eastern Alps five units with different Late Devonian/Carboniferous tectonic evolutions have been distinguished by [*Neubauer*, 1988] and [*Frisch and Neubauer*, 1989]. They support models which explain the Variscan history of Alpine basement units as a result of continent-continent collision between Gondwana-derived continental elements and elements of the Central European Variscides in the North. In this scenario finally the Noric-Bosnian Terrane (NBT, see chapter 1.4.2) was accreted during late Early Carboniferious by subsequent continental plate collision [*Neubauer and Handler*, 2000]. The NBT –the Austroalpine and Southalpine units- comprise well-studied early Palaeozoic to Late Carboniferous sedimentary sequences in a passive continental margin which show no indications for a pre-Carboniferous orogenic event [*Neubauer et al.*, 1999]. This precludes any major collision with Gondwana to the south before Late Carboniferous times [*Stampfli*, 1996].

In the South of the Periadriatic Line (Gailtal Fault) Southalpine palaeozoic sequences are exposed in the Carnic Alps, forming the basement of the Southern Alps (Fig. 1-11). The palaeomagnetic data set of the Alpine region for pre-Variscan times is still rather sparse and of low quality. Thus, no coherent model for the geodynamic evolution of the Alpine elements during the variscan orogeny has been brought forward yet. To confirm the Silurian and Devonian palaeolatitude for the NBT, determined in Palaeozic sediments of the Austroalpine unit (chapter 4.1), palaeomagnetic studies were taken out in Lower Devonian limestones of the Carnic Alps which belong to the common Palaeozoic sedimentary basin of the NBT [*Frisch and Neubauer*, 1989] [*Schönlaub*, 1992] [*Neubauer and Raumer*, 1993].

SAMPLING AND TECTONIC STRUCTURES

The Carnic Alps at the border of Southern Austria and Northern Italy (Fig. 4-10) comprise an almost complete and biostratigraphically well dated [*Schönlaub*, 1992] succession ranging from the Ordovician up to the Carboniferous (Fig. 1-12). Varying stratigraphic sequences range from shelf deposits, pelagic and reef carbonates and flysch sediments and reflect several changes in the palaeoenvironment during the Palaeozoic. According to *Vai*, [1979], the horizontal shortening of the sedimentary basin of the Carnic Alps during the Variscan deformation is estimated to range between 75-80%. Deformation began in Visean/Namurian times, the early stage of the Variscan collision of the Noric Bosnian Terrane with the northern continents [*Läufer et al.*, 2001]. Intense crustal shortening creates a S-verging fold and thrust belt indicative for the collision of the northern margin of Gondwana with the complex collage of composite terranes of the Variscan realm. During Late Carboniferous the collision has been completed. The deformed basement is covered by post-Variscan successions with a distinct angular unconformity [*Schönlaub and Heinisch*, 1993]. These post-Variscan cover-rocks are only slightly deformed, thus the main tectonic structures and prominent deformations of the Palaeozoic rocks are of Variscan age [*Läufer et al.*, 2001]. According to [*Läufer et al.*, 2001] in the whole Carnic Alps two deformation events ($D_1$ and $D_2$) of truly Variscan age can be recognised. The ductile deformation $D_1$, however, affected only the western Carnic Alps. The second deformation $D_2$ affected the whole Carnic Alps particularly the central Carnic Alps, the sampling area (Fig. 4-10). The structures of $D_2$ corresponds to more or less SW-verging thrusts and folds with approximately ESE/WNW-trending axes and an axial planar cleavage, whereas quartz is not recrystallized. Therefore, peak temperatures in the central Carnic Alps must have been somewhat lower than 350°C [*Läufer et al.*, 2001]. This is also shown by Illite crystallinity data, which reveal a low-grade metamorphism for the central Carnic Alps [*Hubich et al.*, 1999] [*Läufer*, 1996], as well as $^{40}Ar/^{39}Ar$-data of detrital muscovites of Ordovician sediments, which show Cadomian ages [*Dallmeyer and Neubauer*, 1994], demonstrate the low intense of Variscan and Alpine thermal overprint in the sampling area.

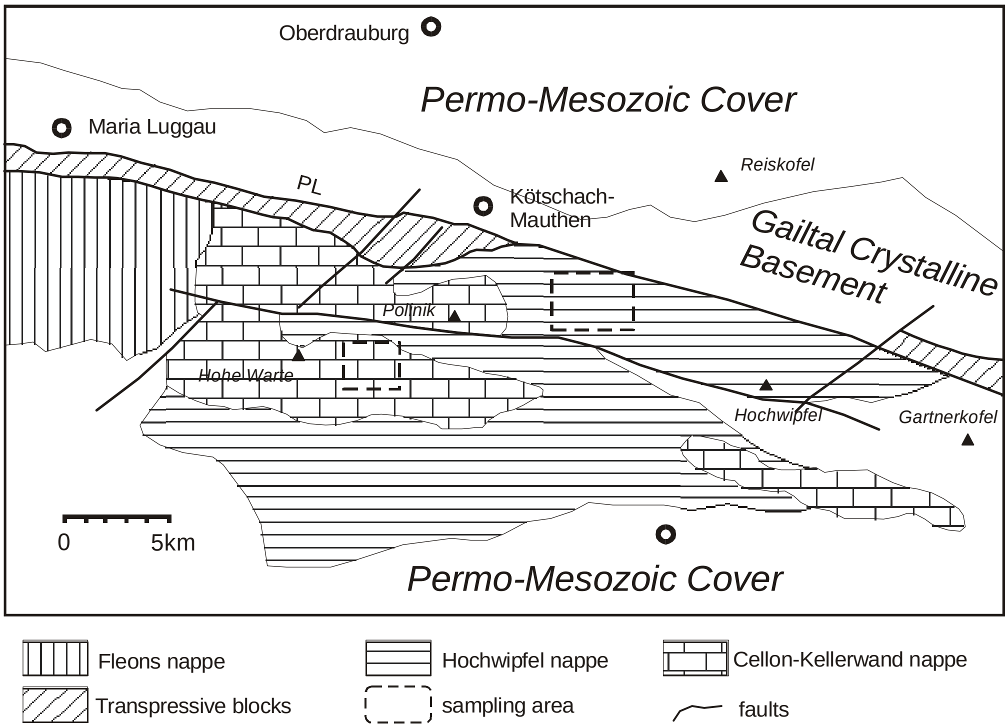



Figure 4-10: Tectonic map of the Carnic Alps with their four tectonic units (Fleons nappe, Cellon-Kellerwand nappe, Hochwipfel nappe, transpressive blocks along the Periadriatic Lineament) after [*Läufer et al.*, 2001]. PL: Periadriatic Lineament. Open circles represent the sampling localities.

Palaeomagnetic data of Mesozoic and Cenozoic sediments from the post-Variscan cover in the surrounding of the sampling area (Venetian Alps) show NW-declinations with intermediate inclinations. This is an indication for counter clockwise rotations of 20° to 40°, with respect to stable Europe [*Besse and Courtillot*, 1991] since Early Permian time. According to [*Channell et al.*, 1992b] the observed rotations are not significant for thrust tectonics, rather they demonstrate a uniform rotational history for the Venetian Alps. This is due to the subsequent convergence and indentation of the Adriatic Plate with a counter clockwise rotation during the Alpine orogeny. It is assumed, that magnetic directions with similar behaviour determined in palaeozoic sequences are of secondary origin acquired during low-grade alpine metamorphic events in the Southalpine unit.

There are several opinions to distinguish the nappe structures in the sampling area of the Carnic Alps regarding to their stratigraphic, sedimentological, deformational and metamorphic environment. According to [*Schönlaub*, 1992] six subnappes are differed by their different sedimentological evolution and bordered by thrust faults (Fig. 4-11). On the other hand [*Läufer et al.*, 2001] describs four major tectonic units (Fig. 4-10): the Late-Alpine transpressive blocks along the Periadriatic Line, the high-grade metamorphic Fleons nappe in the western Carnic Alps and two low-grade metamorphic nappes, the Cellon-Kellerwand nappe in the central Carnic Alps and the Hochwipfel nappe in the central and eastern Carnic Alps. All nappe units contain imbricate thrusts, duplex structures and internal folding. During the Alpine Orogeny the standing

Variscan tectonic structures, especially the thrust faults, were reactivated by transpressiv tectonics due to the Alpine indenter and lead to internal rotations in between the thrusted nappe units.

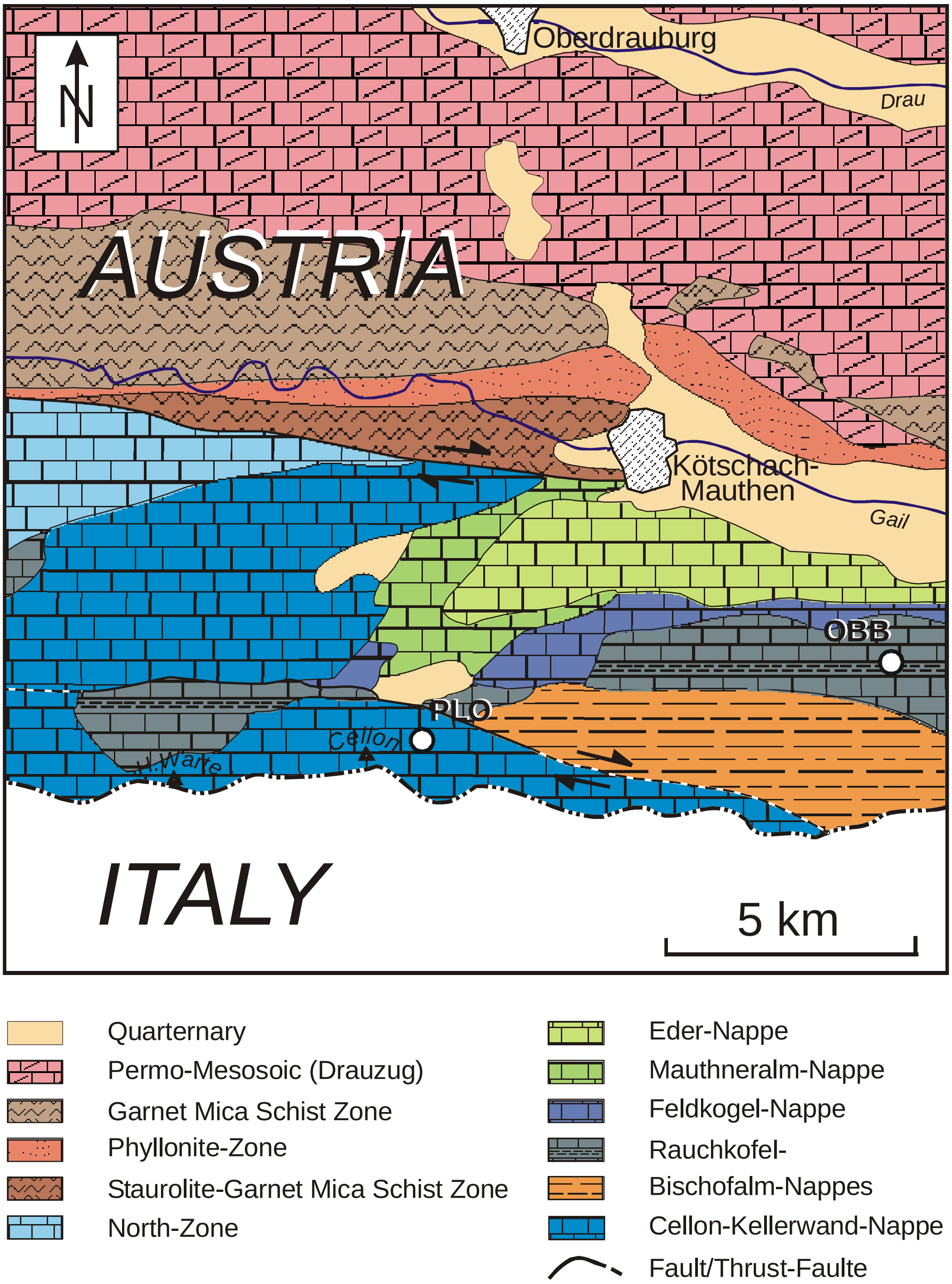



Figure 4-11: Structural map of the Carnic Alps according to [*Schönlaub*, 1992]. Open circles represent the sampling localities.

Lower Devonian samples have been collected from the Cellon-Kellerwand nappe (PLO) and from the Hochwipfel nappe (OBB, OBI) (Figs 4-10 and 4-11). Both tectonic units comprise a continuous Late Orodovician to Lower Carboniferous stratigraphic sequence. During the Silurian a significant increase in basin subsidence can be identified which peaked during the Devonian. In the Devonian, the lithological environment comprises shallow-water limestones, neritic-pelagic limestones, and other pelagic deposits with lateral facies changes. This is documented in a Lower Devonian transgressional sequence including the Rauchkofel Limestone and the Finding (Tentaculiten) limestone. During the Pragian and Emsian stages a pronounced differentiation in the sedimentary basin is observed. A strongly varying facies pattern develops, indicating a progressive but not uniform deepening of the basin [*Kreutzer*, 1992].

Gently folded sequences of the Cellon-Kellerwand nappe unit near the Plöckenpass were sampled. A total of 26 samples of six different fold sections were collected (Table 4-4) (PLO), from the Lower Devonian Rauchkofel limestone. In the Hochwipfel nappe unit coeval Rauchkofel and Finding limestones were also collected (10 sites, Table 4-4) from distinct imbricate thrust sheets exposed near Oberbuchbach (OBB and OBI).

PALAEOMAGNETIC RESULTS

For palaeomagnetic investigations 97 oriented core samples were taken at 18 sites from two distinct nappes -the Cellon-Kellerwand-nappe and the Hochwipfel-nappe. The samples cover a wide spectrum of various lithologies ranging from the black, bituminous Plattenkalk (Lochkovian), red clay flaser-limestones and red-grey Goniatiten-Limestones (Pragian).

During thermal and alternating field demagnetization experiments up to three components of magnetization can be identified in most samples. Demagnetization at low temperatures (150°-200°C) or low AF (up to 25mT) show removal of component A pointing to the North with steep positive inclinations. This component resembles the direction of the present day geomagnetic field in the region studied and is interpreted as being of secondary, recent, origin. A second component (B) is removed at blocking temperatures of 360°C to 400°C or at peak alternating fields of 70 mT. The magnetic direction B displays north-westerly and intermediately inclined directions of magnetization, which are similar to Mesozoic and Cenozoic directions reported from the post-Variscan cover of the Southern Alps, and are interpreted to reflect Alpine overprints. In the same blocking temperature range some samples yield a north-easterly and shallow inclined direction, yet with a wide scatter of declination directions. These occasional recognised directions could reflect affects of the above mentioned deformation event $D_2$ described by [*Läufer et al.*, 2001]. Some samples show curved demagnetization trajectories up to 380 °C due to overlapping demagnetisation spectras of component A and B.

Above 380°C or 70 mT a third component can be identified and interpreted as the characteristic remanent magnetization (ChRM). Both sampled nappes show different directions of the ChRM which are labeled C1 and C2 and will be discussed in the following sections.

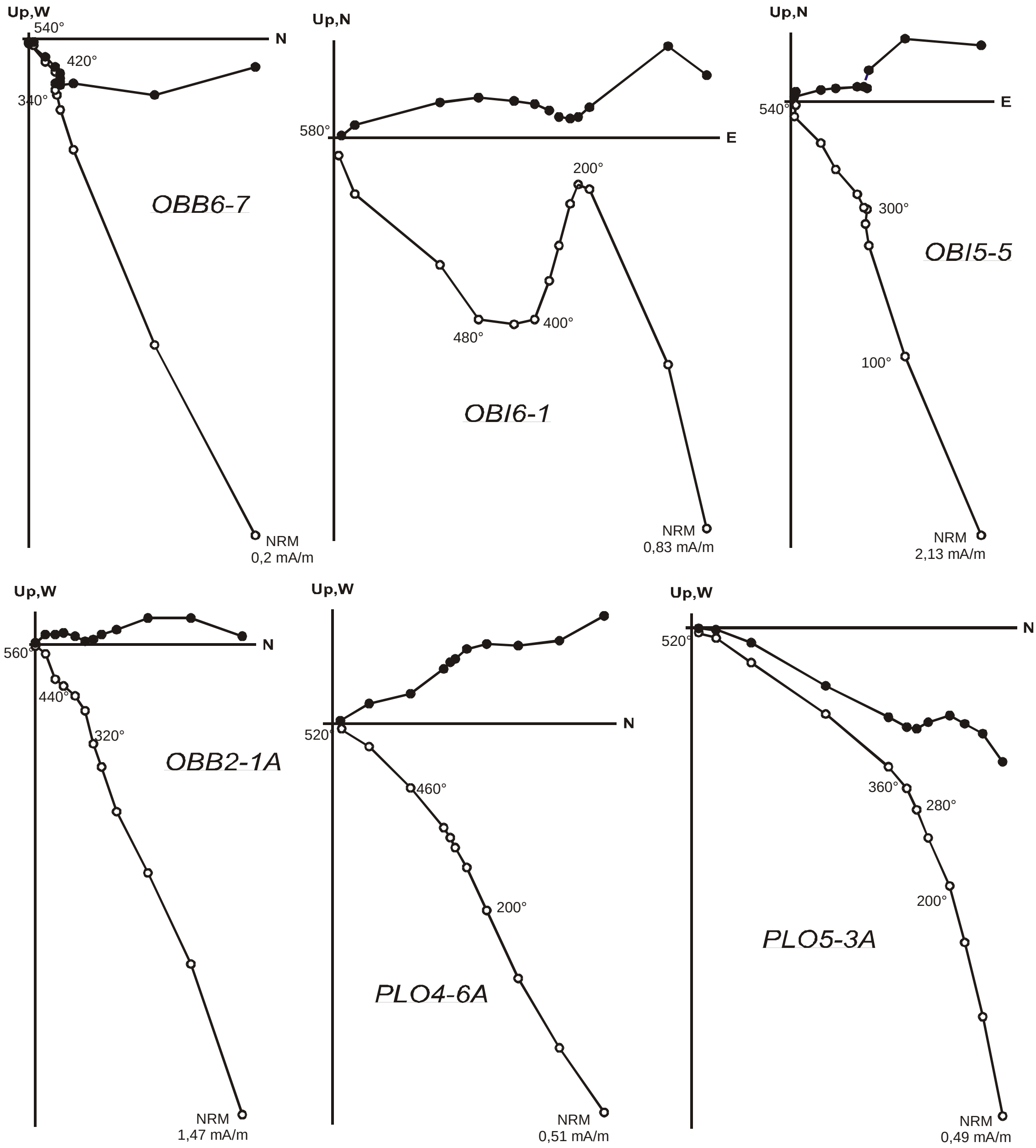



Figure 4-12: Orthogonal projection of thermal demagnetization data for samples from the Lower Devonian carbonates. All directions are plotted in in situ coordinates. Notation as in Fig. 4-4. Almost all Samples show a removal of component A up to 200°-240°C. At intermediate temperatures up to 360°-400°C component B is removed. At temperatures higher than 400°-440°C, the samples reveal the component C.

## Cellon-Kellerwand-Nappe

The stable endpoint direction identified in this tectonic unit has maximal $T_{ub}$ of 580°C and is termed component C1. The six fold sections yield an overall site direction of Dec=354°, Inc=44° ($\alpha_{95}$=14.5°, k=22.4) in situ, which translates into Dec=257°, Inc=48° ($\alpha_{95}$=3.9°, k=84.9, N=6 fold sections) upon bedding correction (Fig. 4-13, Table 4-5). The statistical distribution of the sample mean direction was tested during stepwise unfolding of the fold sections using the procedure of [*McFadden*, 1990].

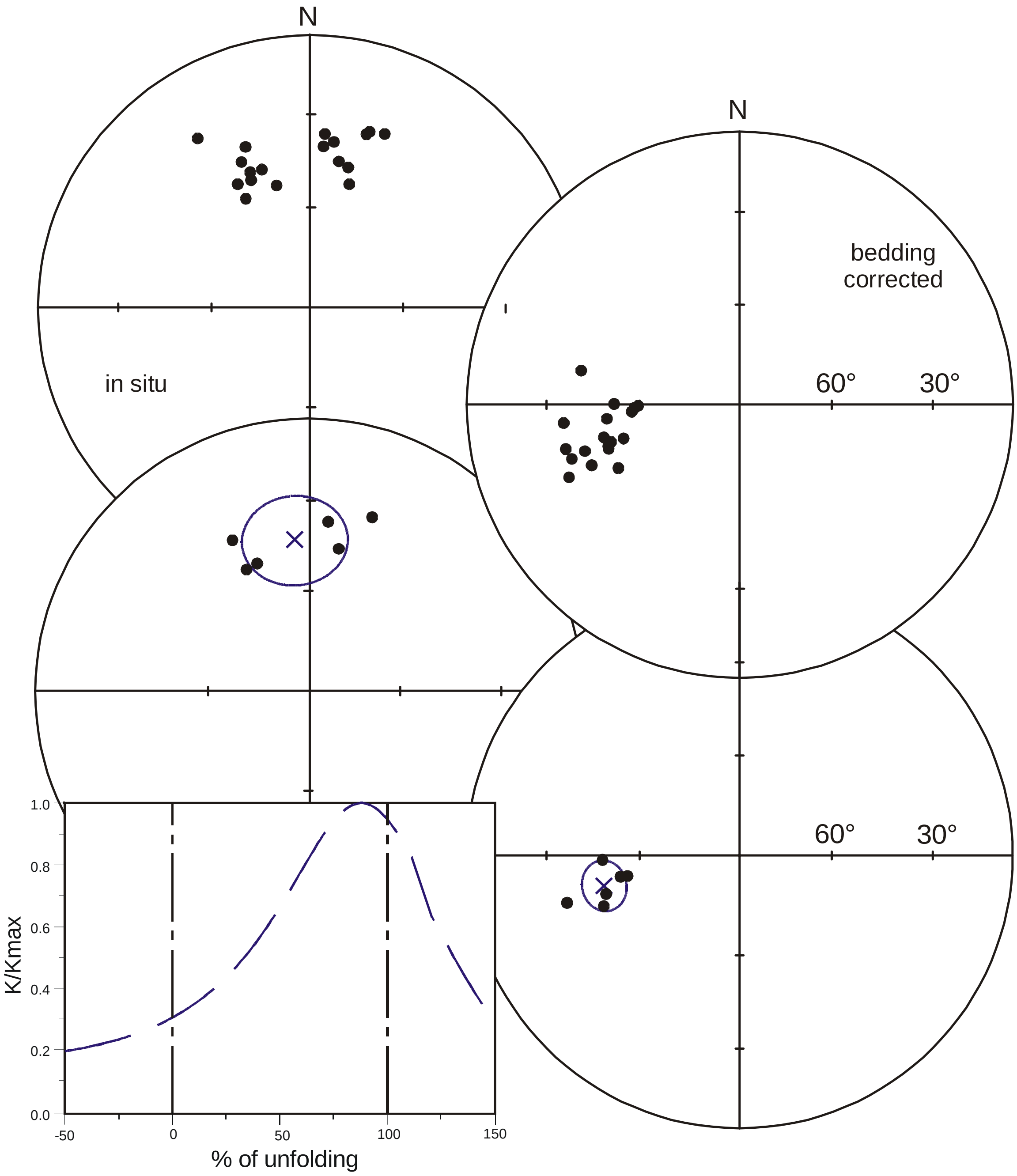



Figure 4-13: Equal area stereographic projections of the Early Devonian sample (top) and fold section mean (bottom) C1 magnetisation before and after structural correction, and the k/kmax rations versus percentage of unfolding. Solid (open) circles represent lower (upper) hemisphere projections; k is the precision parameter [*Fisher*, 1953], and alpha 95 is the cone of 95% confidence.

The improvement in grouping of the mean directions results in a maximum k (the precision parameter of [*Fisher*, 1953]) of 110.2 at 90 % of untilting. There is no significant difference in the mean direction at 90 % unfolding and 100 % and thus the fold test is positive at the 99 % level of confidence using the fold test of [*McElhinny*, 1964], indicating a pre-folding age of magnetization.

Table 4-5: Mean directions for Lower Devonian sedimentary rocks of the central Carnic Alps, component C

| | | | | | Component C | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|
| | | | | | In Situ | | | Bedding Corrected | | |
| Site | Age | Lithology | Bedding | N/n | Dec/Inc (°) | k | $\alpha_{95}$ (°) | Dec/Inc (°) | k | $\alpha_{95}$ (°) |
| | | | | **Cellon-Kellerwand nappe** | | | | | | |
| | Fold sections | | | | | | | | | |
| A | L. Dev. | limestones | 205/66 | 5 / 3 | 338/48 | 172,8 | 9,4 | 249/46 | 172,8 | 9,4 |
| B | L. Dev. | limestones | 200/50 | 4 / 2 | 333/49 | 199,0 | 17,8 | 260/56 | 199,0 | 17,8 |
| C | L. Dev. | limestones | 203/60 | 4 / 4 | 332/38 | 71,1 | 11,0 | 268/48 | 71,1 | 11,0 |
| D | L. Dev. | limestones | 217/84 | 3 / 2 | 001/36 | 197,6 | 7,5 | 254/47 | 197,6 | 7,5 |
| E | L. Dev. | limestones | 224/90 | 4 / 3 | 020/33 | 74,9 | 11,8 | 260/54 | 74,9 | 11,8 |
| F | L. Dev. | limestones | 226/85 | 6 / 4 | 012/46 | 177,1 | 6,9 | 255/35 | 177,1 | 6,9 |
| C1 | Mean direction | | | 26/18 | 354/44 | 22,4 | 14,5 | 257/48 | 90,9 | 7,1 |
| | | | | **Hochwipfel nappe** | | | | | | |
| OBB 1 | L. Dev. | limestones | 195/66 | 6 / 5 | 015/62 | 94.9 | 7.9 | 195/52 | 94.9 | 7.9 |
| OBB 2 | L. Dev. | limestones | 200/73 | 6 / 5 | 006/50 | 32.6 | 13.6 | 217/56 | 32.6 | 13.6 |
| OBB 6 | L. Dev. | limestones | 208/70 | 10 / 5 | 062/48 | 42.0 | 11.9 | 171/51 | 42.0 | 11.9 |
| OBB 7 | L. Dev. | limestones | 208/70 | 6 / 3 | 048/59 | 27.7 | 23.9 | 192/48 | 27.7 | 23.9 |
| OBB 8 | L. Dev. | limestones | 189/80 | 6 / 2 | 019/44 | 44.3 | 32.4 | 179/51 | 44.3 | 32.4 |
| OBB 10 | L. Dev. | limestones | 192/82 | 6 / 5 | 022/46 | 98.3 | 7.8 | 181/53 | 98.3 | 7.8 |
| OBB 11 | L. Dev. | limestones | 183/83 | 4 / 4 | 017/47 | 35.6 | 15.6 | 168/48 | 35.6 | 15.6 |
| OBB 12 | L. Dev. | limestones | 188/86 | 13/11 | 025/44 | 44.0 | 6.6 | 170/48 | 44.0 | 6.6 |
| OBI 5 | L. Dev. | limestones | 188/46 | 9 / 6 | 067/54 | 141.6 | 5.6 | 135/51 | 141.6 | 5.6 |
| OBI 6 | L. Dev. | limestones | 187/48 | 5 / 3 | 063/56 | 112.3 | 11.7 | 139/52 | 112.3 | 11.7 |
| C2 | Mean inclination | | | 71/49 | 45° | 70.0 | 6.1 | 51° | 439.5 | 1.7 |
| | Overall Mean inclination | | | 97/67 | 42° | 33.2 | 5.2 | 50° | 345.0 | 1.6 |

Notation as in Table 4-2. L. Dev. – Lower Devonian

### Hochwipfel-Nappe

The stable endpoint directions obtained in several imbricate thrust sheets of the Hochwipfel-nappe are termed component C2. The overall mean direction obtained for the magnetisation C2 is Dec=032°, Inc=54° ($\alpha$95=8.7°, k=28.8) in situ (10 sites, 49 samples). Application of the bedding correction, does not improve the grouping of the site mean directions (Fig. 4-14). However, the equal area projection of the mean directions in stratigraphic coordinates form a small circle distribution with a average inclination of 51° (Fig. 4-14, Table 4-5). It is assumed, that this distribution is attributed to various amounts of rotations of the sampled thrust sheets as result of the deformation scenario affected the sampling area.

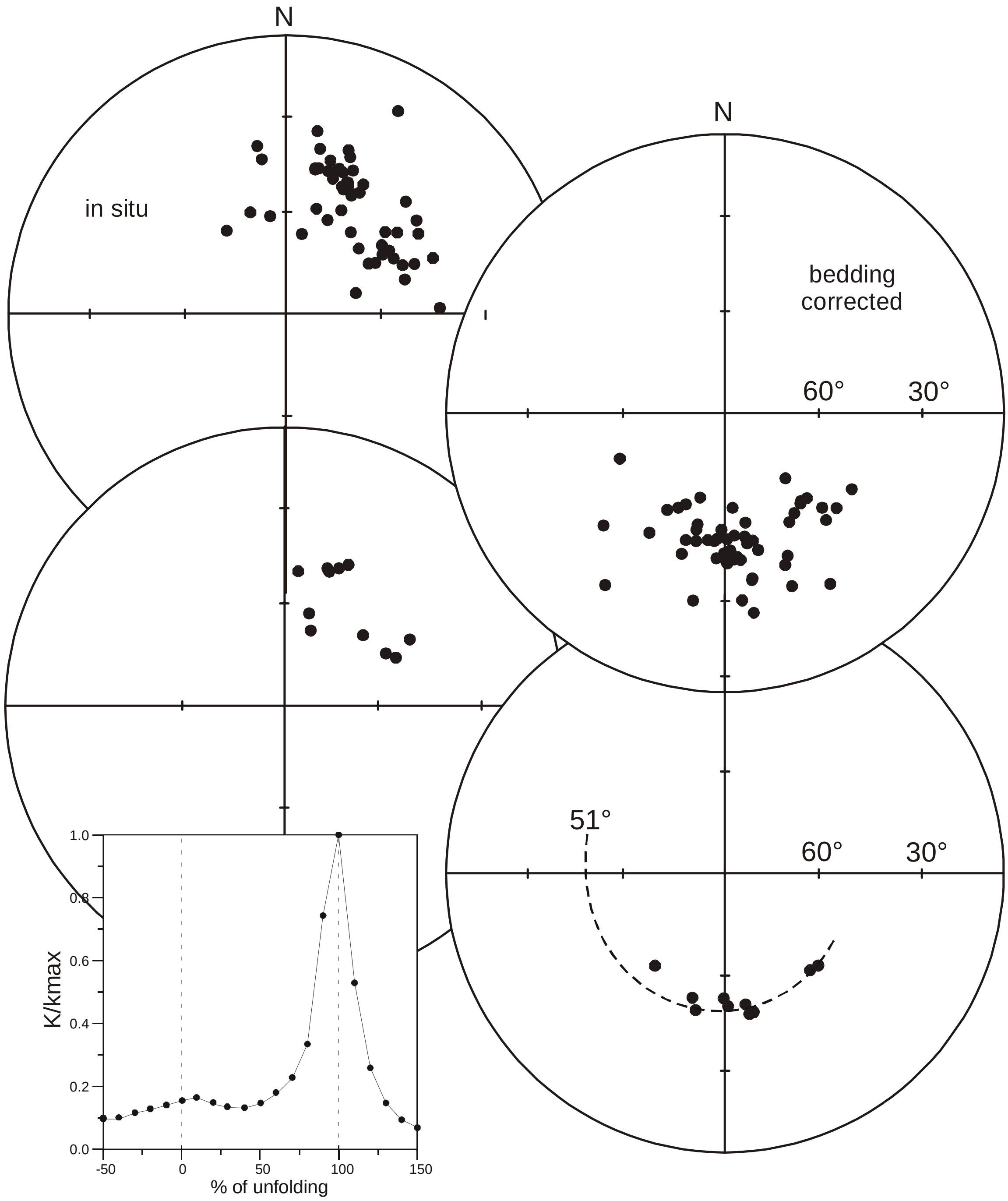



Figure 4-14: Equal area projection of the (a) in situ sample directions (above) and site mean directions (below), and (b) bedding corrected sample directions (above) and site mean directions (below) for component C2 along a small circle segment. (c) The calculated k/kmax ratio vs percentage of unfolding applying the inclination-only test. Notation as in Fig. 4-3.

The statistical distribution of the site mean directions was tested during stepwise untilting the rocks, applying an inclination only test of [*Enkin and Watson*, 1996]. Applying this procedure results in a maximum k of 439.5 at 100 % of untilting (Fig. 4-14, Table 4-5) with k increasing from k = 72.0 (in situ). With that the inclination only fold test of the sampled Lower Devonian limestones within several thrust sheets is positive at the 99 % confidence level [*McElhinny*, 1964] and amounts to an overall inclination value of 51° with $\alpha_{95}$ = 1.7°. Thus it is assumed that the magnetisation C2 is also of primary origin.

INTERPRETATION

The demagnetisation results represented in Figure 4-12 demonstrate a mixed magnetic mineralogy as remanence carrier of the identified magnetic directions. To characterize the carriers of magnetisation several rock magnetic experiments were carried out. Isothermal remanent magnetisation (IRM) acquisition studies of the Lower Devonian limestones (Fig. 4-15) indicate at some samples saturation at about 500 mT. A few samples contain also a higher coercivity phase which is not finally saturated by the maximum field of 1.5 T. During thermal demagnetisation the samples show a steep gradually remove of 80 % - 90 % of magnetisation until temperatures of about 400°C (Fig. 4-15). In the unblocking temperature range above 400°C up to 580°C final demagnetisation occurs. This demonstrate the predominance of magnetite as remanence carrier of the final components C1 and C2, also indicated by the maximum unblocking temperature of the NRM. Sulfides and/or occasionally goethite displaying lower unblocking temperatures are in part the carriers of the secondary components, which are not discussed in more detail.

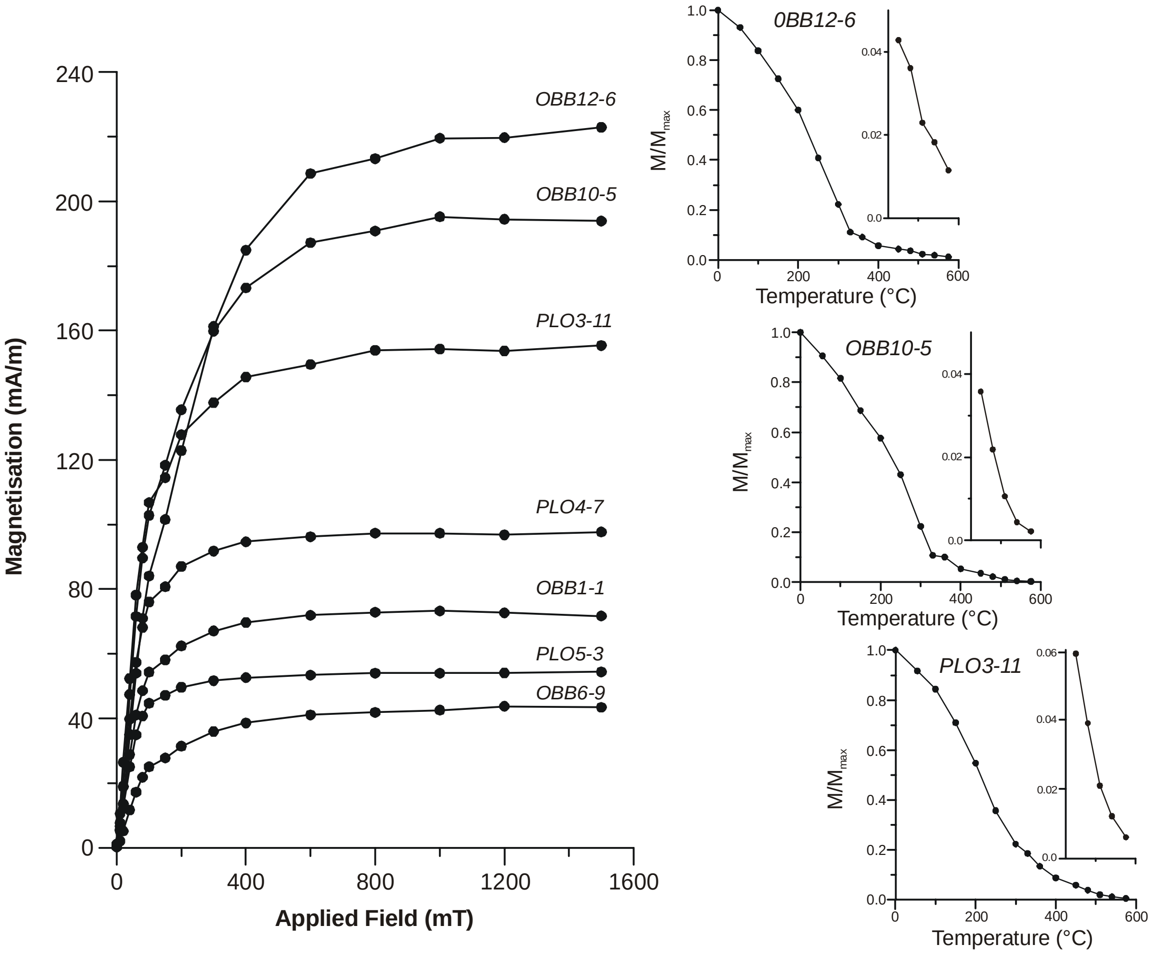



Figure 4-15: Acquisition curves of a monodirectional isothermal remanent magnetization (IRM) of Lower Devonian sediments of the central Carnic Alps and thermal demagnetisation curves.

Evaluating the final components C1 and C2, from the Lower Devonian limestones, reliably identified in 16 sites (67 samples) the tectonic environment of the sampling area must be taken into account. The magnetic directions of each component which belong to distinct nappes pass the fold and inclination only test at the 99 % confidence level [*McElhinny*, 1964] and both amounts to an overall inclination value of about 50° after bedding correction (Table 4-5, Fig. 4-16).

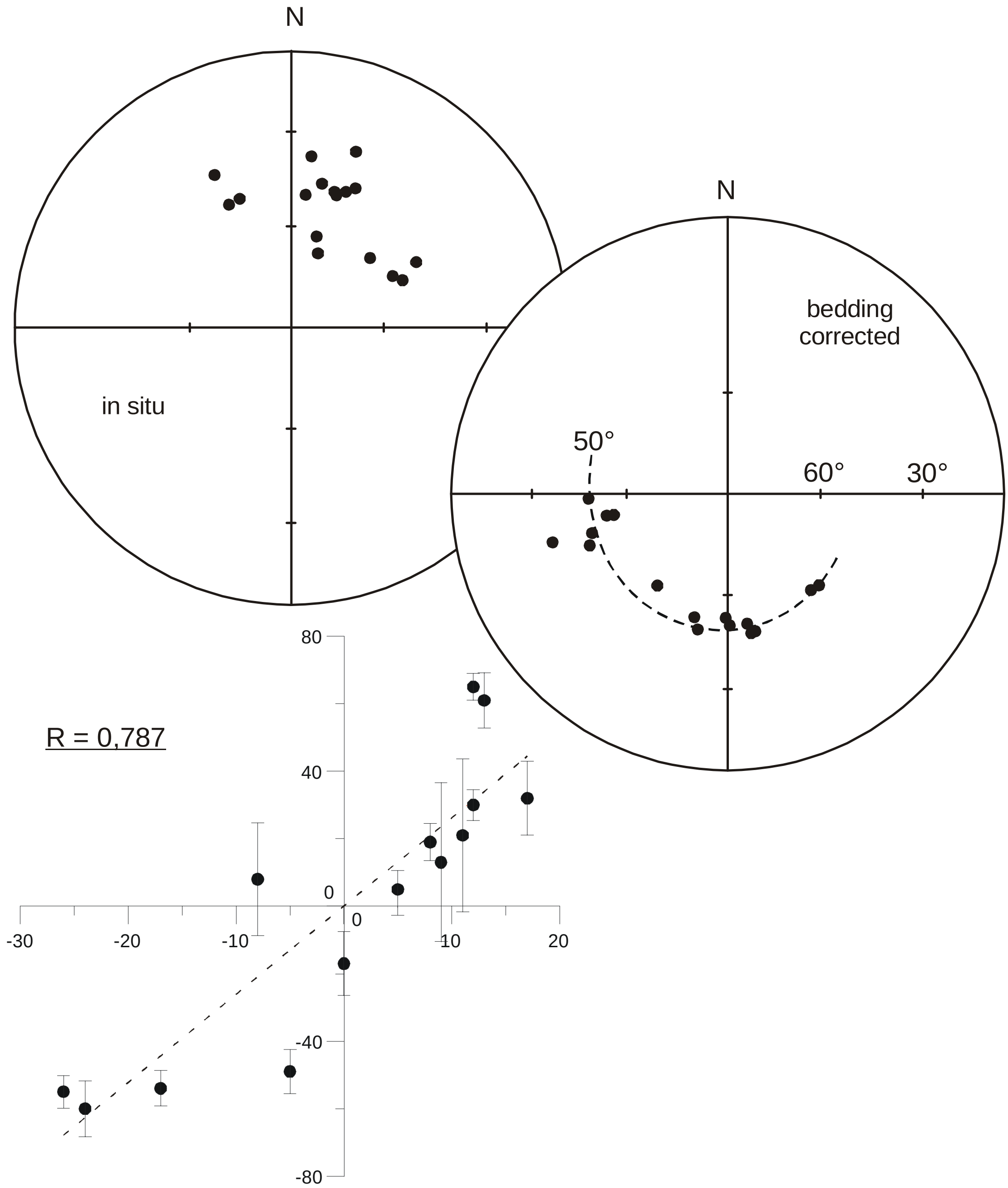



Figure 4-16: Equal area projection of the (a) in situ site mean directions, and (b) bedding corrected site mean directions for component C1 and C2 along a small circle segment. (c) Diagramm showing the strike deviations relative to declination deviations for the sampled nappe units in central Carnic Alps. Sr is the reference strike, So is the observed strike, Dr is the reference declination and Do is the observed declination. R is the correlation coefficient. Error bars corresponds to the calculated $\alpha_{95}$ of each site. Dashed line is a best-fit line calculated by linear regression.

To quantify the structural and tectonic influence on the palaeomagnetic directions of the single sites of both nappe-units a linear regression technique –strike test- is useful, in which the differences between the observed palaeomagnetic declinations ($D_o$) to a reference declination ($D_r$ – $D_o$) are plotted on the ordinate versus the corresponding differences of strikes ($S_r$ – $S_o$) on the abscissa [*Eldredge et al.*, 1985; *Schwartz and Van der Voo*, 1983]. The line of best fit is defined by its correlation coefficient (R). The combined data from the Cellon-Kellerwand nappe and the Hochwipfel nappe are plotted in Figure 4-16. The linear regression results in a significant correlation coefficient of 0.79 as well as a regression line that is approaching ideal oroclinal bending [*Eldredge et al.*, 1985]. Whereas, if the gradient of the regression line is bigger than 1, [*Eldredge et al.*, 1985] postulated that the orogenic belt contains parallel, nearly linear, thrust fronts, while it consists of thrust sheets that underwent various amounts of rotations. This is in good agreement with the data set represented in Figure 4-16 which point to a gradient of about 2 and the above described tectonic environment.

Taking all discussed aspects into consideration it is assumed that the determined magnetisation direction termed C1 and C2 yield a primary inclination value. Using the reversed polarity option the new inclination value for the sampled sedimentary rocks translate into a palaeolatitude for the sampling area and the NBT of about 31°S (+ 1.3° and – 1.7°, upper and lower 95% confidence level) in Lower Devonian times.

# Chapter 5 Palaeozoic Palaeomagnetic results of the Malopolska Block

## Chapter 5.1 Palaeomagnetism of Ordovician carbonate rocks from Malopolska Massif, Holy Cross Mountains, SE Poland- Magnetostratigraphic and Tectonic Implications

INTRODUCTION

Along the Trans-European Suture Zone a number of different Palaeozoic terranes have been identified, based on geological and geophysical studies which have been carried out [*Pozaryski*, 1990] [*Pozaryski et al.*, 1992] [*Berthelsen*, 1992]. The Lysogory, Malopolska and Upper Silesian Massif are fault bounded crustal units at the rezent border of the Baltic Shield in Southern Poland (Fig. 1-13).

Traditional palaeogeographic interpretations have regarded these units as sedimentary realms along the Tornquist margin of Baltica at least since the beginning of the Cambrian period [*Bergström*, 1984] [*Vidal and Moczydlowska*, 1995]. Likewise the Lysogory Unit and Maloposka Massif were recently interpreted as fragments of Baltica's Precambrian crust [*Dadlez*, 1995] [*Pharaoh and al.*, 1996]. Primarily this was based on records of Early Cambrian trilobites in Malopolska and Upper Silesian diagnostic of the Baltic zoogeographical province [*Orlowski*, 1975] [*Orlowski*, 1985]. However there are records of Cambrian brachiopods and trilobite trace fossils of certain Gondwanan affinity from Lysogory [*Jendryka-Fuglewicz,* 1992] [*Jendryka-Fuglewicz*, 1998] [*Seilacher*, 1983]. In contrast to this [*Belka et al.*, 2000] lined out, that in some palaeogeographic reconstructions [*Cocks and Fortey*, 1998] [*Cocks et al.*, 1997] the subdivison of the Holy Cross Mountains (HCM) was overlooked and the faunal records from the Lysogory Unit and the Malopolska Massif (juxtaposed in the HCM) were considered as biogeographical data from one single palaeoplate. Therefore, based on isotopic informations and biogeographical studies [*Belka et al.*, 2000] tried to solve the accretionary scenario for these terranes, with the result, that the Maloplska Massif were separated from the surrounding of the northern margin of Gondwana in Early Cambrian and reached a position close to Baltica to share Baltic province trilobites. The co-occurrence of brachiopods of Avalonian affinity in Malopolska indicate still the proximity of the margins of Gondwana. [*Belka et al.*, 2000] propose the collison of the Malopolska Massif with the Baltic margin after Mid Cambrian and before Tremadoc times. Additional palaeomagnetic data allow various interpretations. [*Lewandowski*, 1995] [*Lewandowski*, 1993] assumes large-scale displacement of the Malopolska massif along the

TESZ during the Palaeozoic and rotations with respect to the EEP up to Variscan times, because the data deviate significant from the Baltic Apparent Polar Wander Path (APWP). On the other hand a almost fixed model for the Malopolska Massif in its present position since Silurian times is favoured by [*Nawrocki*, 2000], who published a Late Silurian palaeomagnetic pole which shows a congruence to the Baltic APWP and concludes no significant post Silurian tectonic displacements of the Malopolska Massif relative to Baltica have occurred. Due to the unsolved discrepancies the mode of accretion and consolidation of this terrane with respect to the Baltic shield remains unclear.

## SAMPLING AND LITHOLOGY

Ordovician rocks in the Holy cross Mountains are generally poorly exposed and there are only a few localities where surface exposures can be found (see Chapter 1.4.3). The aim of sampling was the Mid Ordovician Mojcza limestone whose type locality is exposed in in an old quarry at Mójcza near to the city Kielce (Fig. 1-14). The section dips towards NE under an angle of about 38° and the exposed profile has a stratigraphic thickness of about 8 m (Fig. 5-1). The very basal part of the profile comprises arenaceous limestones, and becomes organodetrital with phosphatic nodules and ferruginous ooids above a basal discontinuity. The main part of the section is dominated by carbonate sediments with a thin bentonite layer in the central part of the profile (Fig. 5-1). In the uppermost part of the profile, about 3.5m above the bentonite layer, the sequences consists of argillaceous limestones with interbedded marls [*Dzik and Pisera*, 1994]. The most characteristic inorganic components of the Mójcza Limestone are ferruginous ooids. The occurrence of the ooids seems to be controlled by delivery of iron into the sedimentary basin, possibly in connection with volcanism as documented by the bentonite layer, or transported from a source area [*Dzik and Pisera*, 1994].

Samples for palaeomagnetic analysis were collected from the condensed continuous sequence of organodetritic limestone beds which span the Llanvirn and Caradoc with a stratigraphic thickness of some 4.5 m (Fig. 5-1). The stratigraphic range of this part of the formation is precisely determined using conodont biostratigraphy [*Dzik*, 1978] [*Dzik*, 1994]. Numerous analysis of the conodont colour alteration indices (CAI) from palaeozoic rocks of the Malopolska Massif show CAI values between 1 and maximum 2 [*Belka*, 1990] [*Belka and Siewniak-Madej*, 1996]indicating a maximum palaeotemperature below 150°C [*Epstein et al.*, 1977] and thus the suitability of the section for palaeomagnetic analysis. Samples were collected at ca. 5-10 cm intervals in the central part of the section (Fig. 5-1), covering the time interval from the upper Llanvirn *Eoplacognathus robustus* subzonesto the Cardocian *Amorphognathus superbus* zone. All samples were taken with a portable, gasoline-powered drill and oriented in the field using a standard magnetic compass.

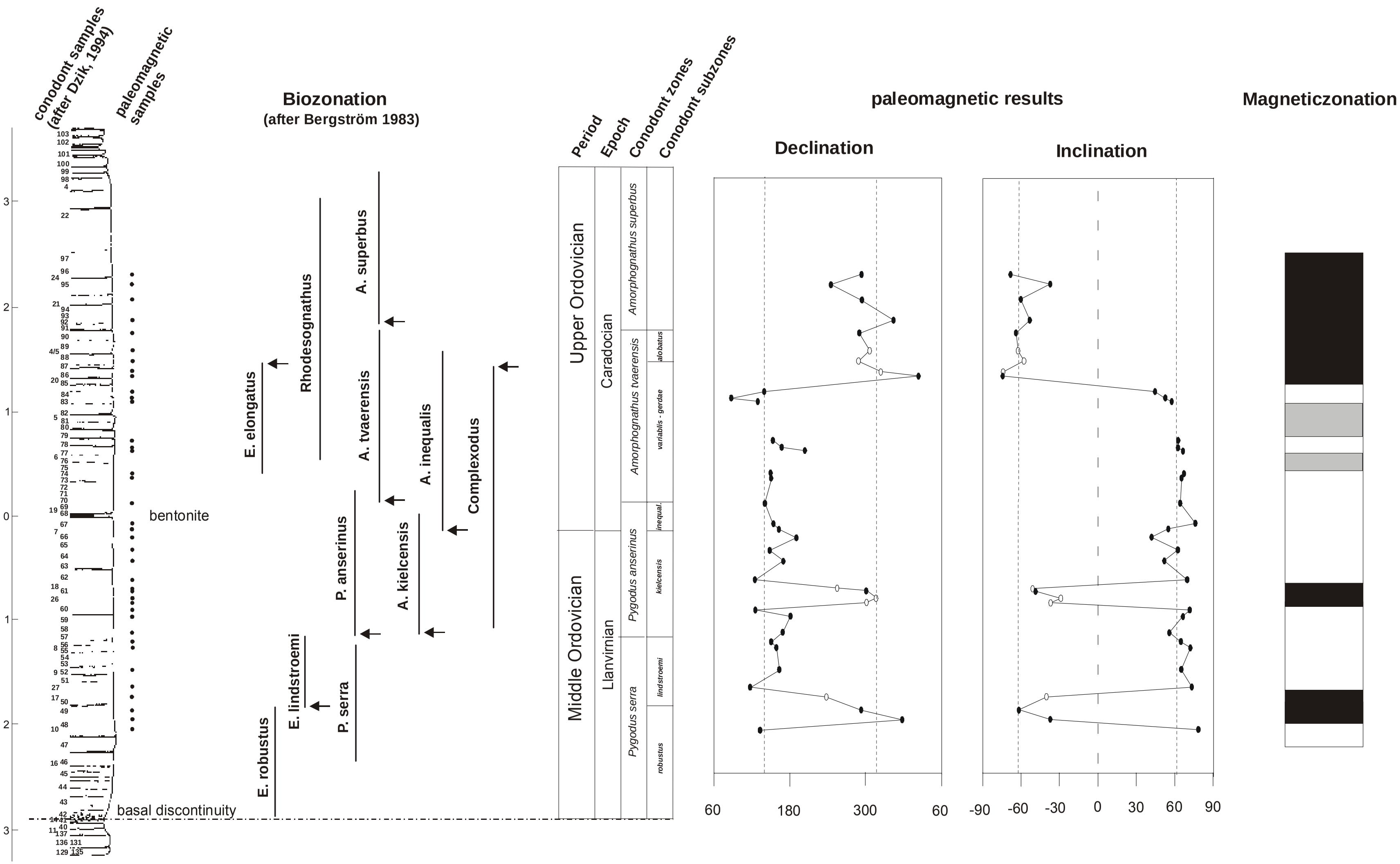



Figure 5-1: Stratigraphic column of the sampled section of the Mójcza limestone with sample locations for faunal [*Dzik*, 1994] and palaeomagnetic studies. The correlation with the global stratigraphy(compiled by Dr B.D. Webby in consultation with Drs R.A. Cooper, S.M. Bergström and Florentin Paris for the IGCP Project No. 410) was done using the biozonation for platform conodonts by [*Bergström*, 1983]. Arrows mark important FAD and LAD of conodont species, which were used for correlation. The palaeomagnetic results indicate at least five polarity changes. Filled circles mark results, where the direction of magnetisation was identified ba linear segments and stable endpoints, open circles indicate results from great circle analysis. Black (white) intervals represent normal (inverse) polarity of the Earth magnetic field. The grey parts indicate intervals of about 15 to 30 cm length where no samples were taken.

## BIOZONATION

A characteristic feature of this formation is the absence of complete macrofossils, these were destroyed during prolonged periods of exposure on the sea bottom due to the very slow sedimentation rates [*Dzik and Pisera*, 1994]. The biostratigraphically most important fossils in the Mójcza Limestone are pelagic conodonts [*Dzik and Pisera*, 1994]. The conodont fauna in the Mojza Limestone is dominated by the platform-conodont genera including the lineages of *Eoplacognatus*, *Pygodus* and *Amorphognatus* [*Dzik and Pisera*, 1994]. All conodont apparatuses of sixty seven species of thirty six genera are described by [*Dzik*, 1978] [*Dzik*, 1994]. For a precise biostratigraphical zonation of the sampled section, the evolutionary lineages and relationships of Ordovician platform-conodonts given by [*Bergström*, 1983] were applied here (Fig. 3). The arrows in the biozonation section of Fig. 5-1 mark the “First-“ and “Last Appearance Dates” which are indicative for biozone bounderies. The record shows a stratigraphic completeness, with only one significant hiatus associated with the discontinuity at the lower part of the section, and corresponding to the subzones *E. pseudoplanus* to *E. reclinatus*. Samples for palaeomagnetic analysis were collected upwards from the discontinuity, in the conodont zones ranging from the Pygodus serra zone in the Llanvirnian, to the Amorphognathus superbus zone in the Caradocian. The more or less continuous replacement of one lineage by another allows a detailed documentation of conodont zones and subzones in this part of the section [*Dzik*, 1994]. No significant faunal changes are observed across the bentonite layer (Fig. 5-1). It seems that this event of volcanic activity had no direct influence on the organic life in the Malopolska area. It is not possible to correlate the Mójcza bentonite bed with any other in the Baltic area [*Dzik and Pisera*, 1994].

The conodont fauna is believed to be of shallow and warm water conditions, but also cold water genera like *Sagittodontina*, *Scaccardella* and *Hamarodus*, as well as some of Welsh affinity such as *Complexodus*, *Phragmodus* and *Rhodesognathus* are dominant in places. In general, the conodont faunas show a mix of Welsh and Baltic affinities, but given their more pandemic nature conodonts are poor biogeographic indicators. Perhaps more significant is the lack of certain typical Baltic genera in the section, such as *Peridon, Eoplacognathus* and *Scalpellodus* are rare [*Dzik and Pisera*, 1994]. The diversity of Ordovician conodont assemblages seems to depend more on local ecologic factors and bathymetry than on climate [*Dzik*, 1983].

Correlation of this sub-section with the global stratigraphy (compiled by Dr B.D. Webby in consultation with Drs R.A. Cooper, S.M. Bergström and Florentin Paris for the IGCP Project No. 410) was done using the biozonation for platform conodonts of [*Bergström*, 1983]. The sampled section comprise about 9 million years (from the uppermost Pygodus serra zone - Eoplacognathus lindstroemi subzone up to the lowermost Amorphognathus suberbus zone) [*Dzik*, 1978] [*Dzik*, 1994] [*Bergström*, 1983], indicating very low sedimentation rates (ca. 0.5 mm/1000a).

ROCK MAGNETISM

During IRM experiments two types of behaviour can be observed (Fig. 5-2). In the group 1 samples, specimens show a strong increase of their magnetic remanence up to 150 mT and then a slight but continuous increase in intensity at higher fields without reaching saturation up to the maximum applied field of 1.5T, indicating a mixture of both low and high coercivity minerals.

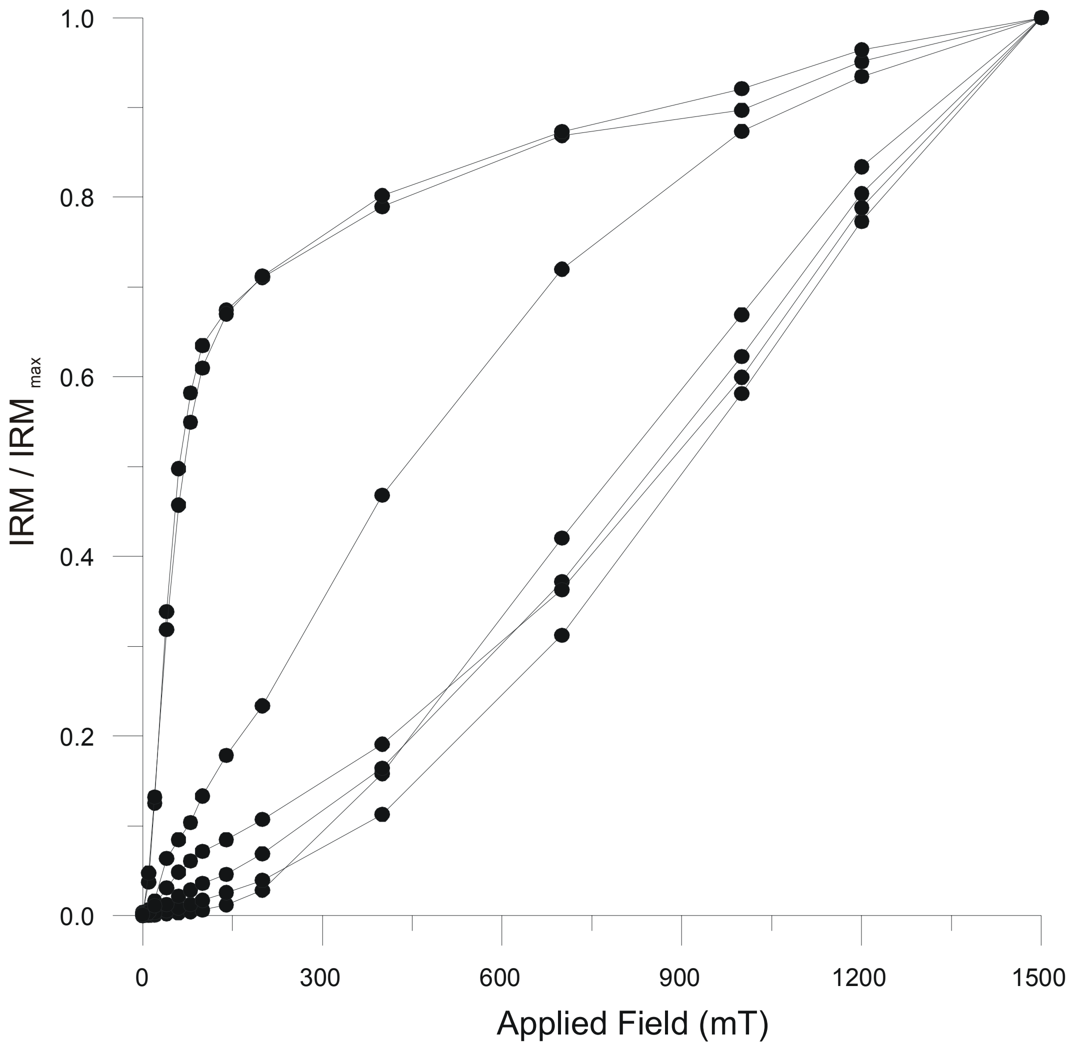



Figure 5-2: Normalized acquisition curves of monodirectional isothermal remanent magnetisation (IRM) of Ordovizian limestones of the Mojcza quarry.

The other group of samples shown in Fig. 5-2 indicate the presence of only high and very high coercivity minerals by the continuous and gradual increase of magnetic intensity up to the maximum applied field. Thermal demagnetisation up to 680°C of a composite IRM, containing three orthogonal axis of IRM’s (1,5 T, 0,5 T, 0,2 T, method as described by [*Lowrie*, 1990]) was also carried out. Group 1 type samples (Fig. 5-3a) show a magnetisation mainly carried by magnetite with low coercivity and an unblocking temperature of approximately 580°C, with minor amounts of goethite (the high coercivity fraction with low unblocking temperatures), sulphides (intermediate coercivity fraction and a inflection in the intensity decay curve at 300°C – 350°C), and hematite (high coercivity and high unblocking temperatures).

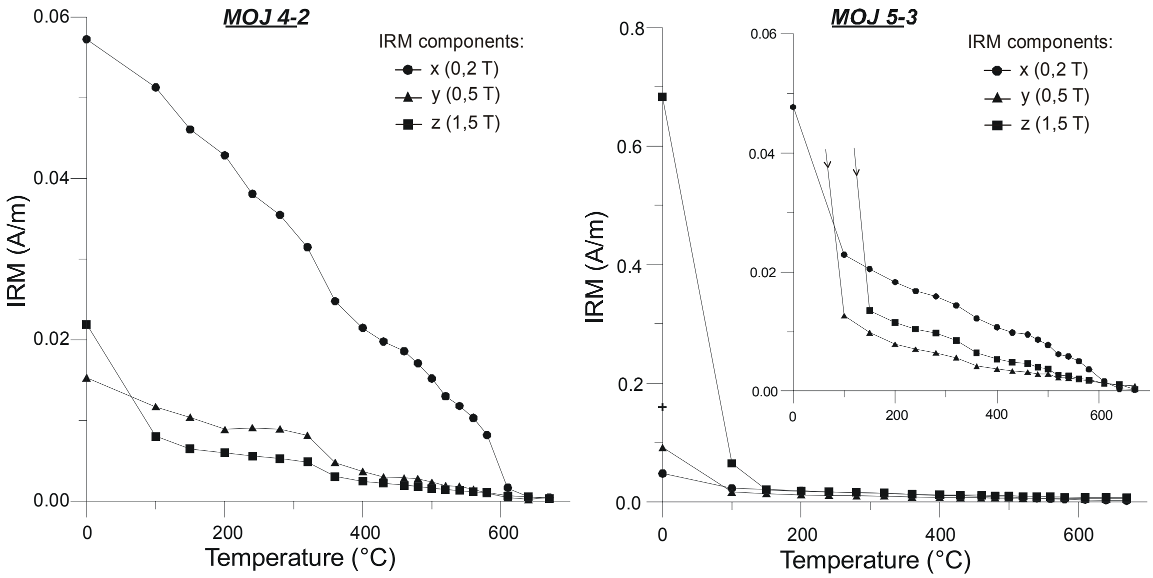



Figure 5-3: Thermal demagnetisation of triaxial differntial IRM (method described in [*Lowrie*, 1990]) in (a) of speciment MOJ4-2 and in (b) of speciment MOJ5-3. The blow-up in 5-3(b) shows a detail of the demagnetisation curve.

Group 2 type samples (Fig. 5-3b) show a magnetization dominated by goethite (high coercivity fraction and low unblocking temperatures) smaller amounts of iron sulphide, magnetite with maximum $T_{ub}$ of 580°, and hematite. The increase in susceptibility after heating to 400°C is most likely due to oxidation of sulphides and production of magnetite. The hysteresis parameters $M_{rs}/M_s$ and $H_{cr}/H_c$ , plot either in PSD field or with a trend to higher $H_{cr}/H_c$ ratios in the typical Day diagramme [*Day et al.*, 1977]. This is typical for samples with mixed magnetic-mineralogy and varying grainsizes.

THERMAL DEMAGNETIZATION

The palaeomagnetic results discussed in this study are based on the magnetic behavior of samples during stepwise thermal demagnetization experiments, as alternating field method failed in most cases to reveal the full character of the remanent magnetisation. The natural remanent magnetisation (NRM) intensities are low and range from 0.2 – 4 mA/m. Most of the samples show multivectorial behaviour, sometimes with overlapping unblocking temperature ($T_{UB}$) spectra, leading to a great circle behaviour in the demagnetisation trajectories. However, in most samples a high $T_{UB}$ stable endpoint direction can be identified.

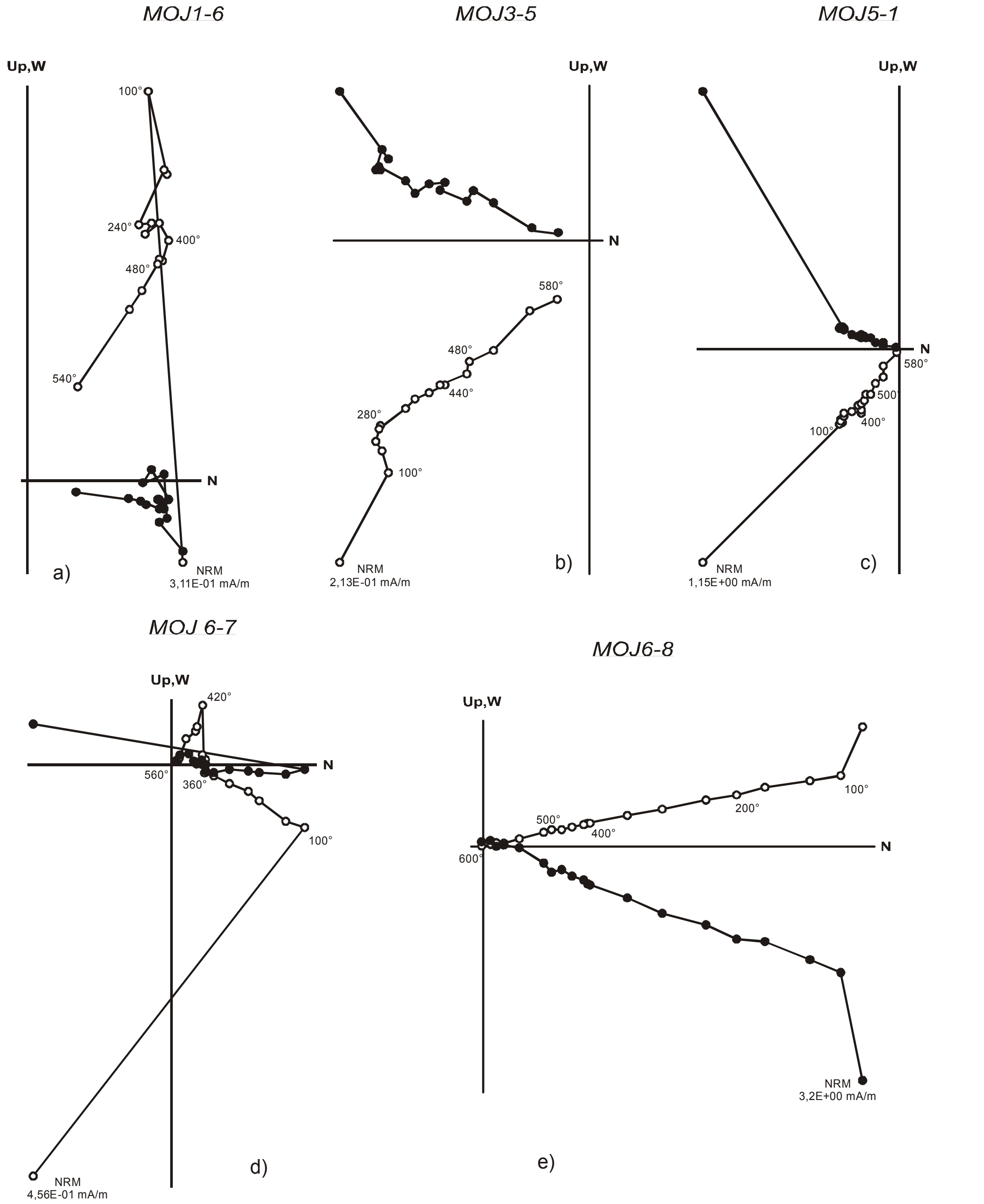



Figure 5-4: Orthogonal projection of thermal demagnetisation behaviour of samples of the Mójcza limestone. All directions are plotted in geographic coordinates and solid (open) symbols represent the horizontal (vertical) component respectively. Specific demagnetisation temperatures are given in degree Celsius. After a removal of a viscous overprint one component of magnetisation is observed, which is stable up to 580°C. The component is of bipolar nature with northerly (southerly) declination and medium to steep negative (positive) inclination in situ.

Different kinds of behaviour during thermal demagnetisation can be observed on vectorial diagrams (Fig. 5-4a-e). Many samples show relatively simple behaviour, after removal of a low temperature direction A of random orientation and carried by goethite, a stable end point direction C with unblocking temperatures up to 580°C can be readily identified (Figs 5-4a & 5-4c). Some samples then either show rather noisy behaviour up to approximately 400°C (Fig. 5-4a) or a direction B with intermediate unblocking temperatures (maximum $T_{ub}$: 420°, Fig. 5-4d) can be identified prior to isolation of the stable end point C direction at higher temperatures.

Occasionally, however the demagnetisation trajectories of some samples are not directed toward the origin at higher temperatures and the samples show great circle behaviour. Stereographic projections (Fig. 5-5) demonstrate that the demagnetisation trajectories are quite well defined for each sample (Figs. 5-5a and 5-5b) and show two distinct great circle paths. The first great circle between 20°C and 320°C trending from northerly positive to southerly shallow directions, and the second, between 360°-380°C and 540°C, moves towards a north/northwesterly negative direction (Figs. 5-5a and 5-5c). When all the higher temperature great circle data are compared, a number of different paths are described, but in most cases, in combination with the stable end point data, the great circle data could be confidently interpreted using the analytic procedure of [*McFadden and McElhinny*, 1988]

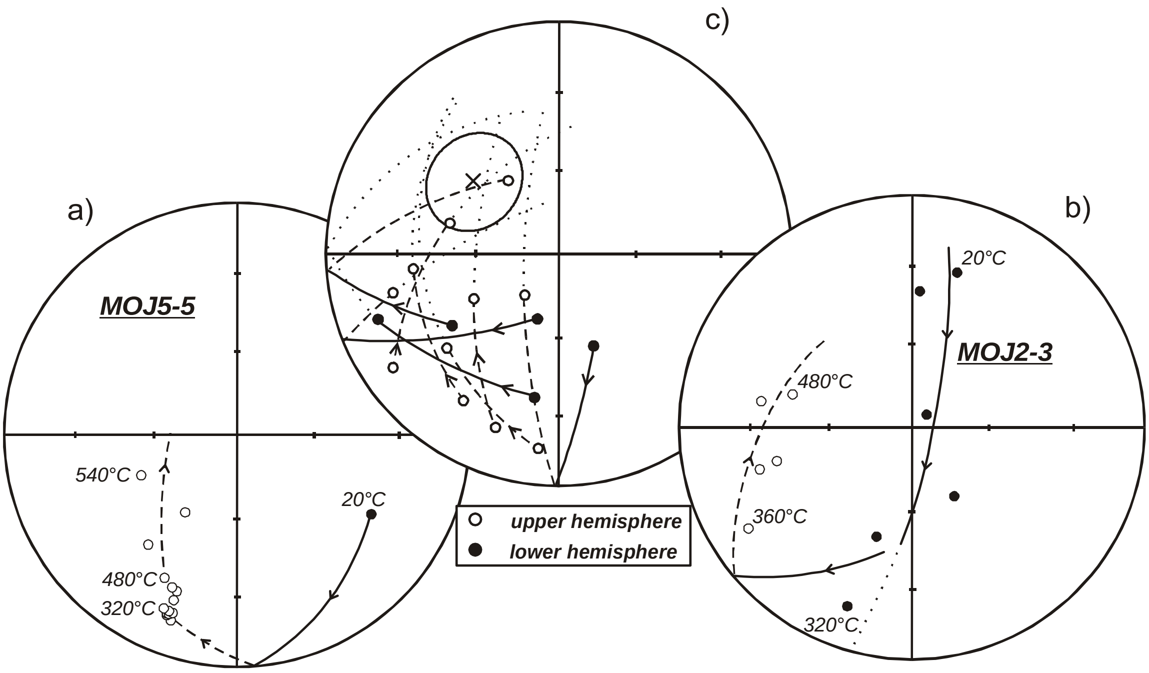



Figure 5-5: Equal area stereographic projection of thermal demagnetisation results of samples of the Mójcza limestone in stratigraphic coordinates, in which stable endpoints are not reached. In (a) two great circle paths can be identified, the first from 20°C to 240°C and the second from 320°C to 540°C resulting from removal of overlapping magnetizations. Also in (b) two great circle paths can be identified. At temperatures higher than 480°C and 540°C the magnetization directions become random due to mineralogical alteration and no further reliable mesurements could be made.

MAGNETIC DIRECTIONS

Direction B can be identified in a total of 12 samples but the directions show no consistency between samples and thus no mean direction can be calculated. This component is probably carried by magnetite and/or pyrrotite and the erratic nature of the direction probably results from overlapping blocking temperature spectra of different magnetization directions. Most of the lower temperature great circle paths identified, however, move towards shallow southerly directions and are thus most likely indicative of a weak Permo-Carboniferous overprinting. In general, the directions identified up to approximately 400°C are arbitrary in orientation with no significant directional distribution and, therefore, can not be evaluated in more detail.

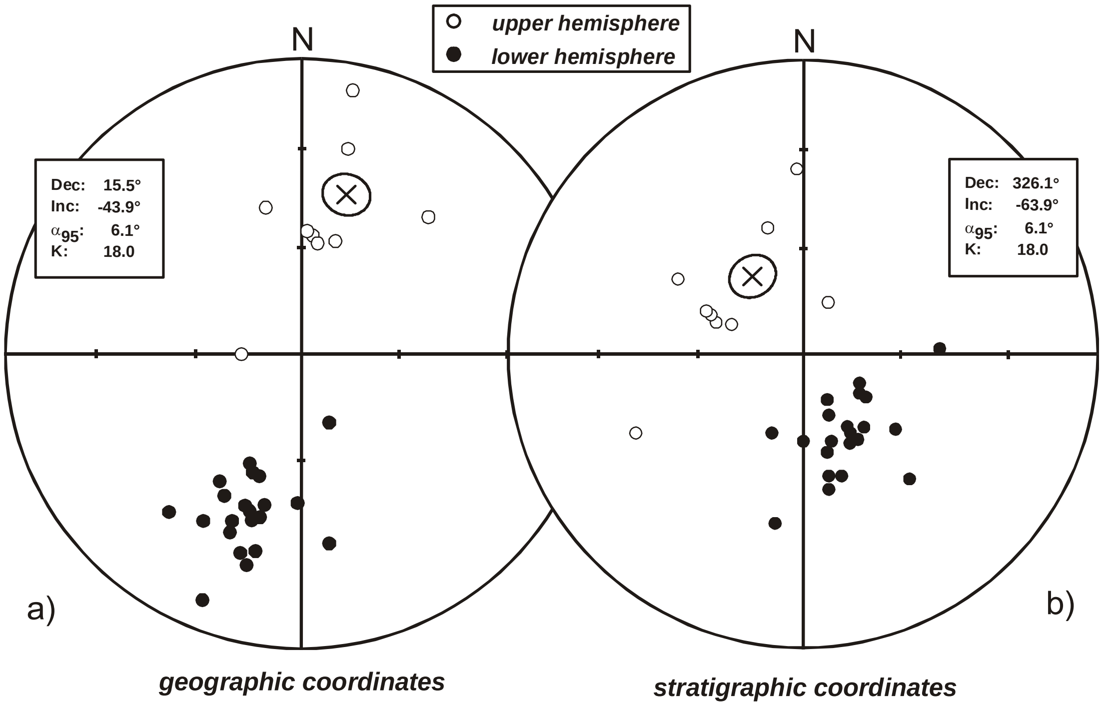



Figure 5-6: Equal area stereographic projection of single sample directions (data from linear segments and stable endpoints only) in geographic (a) and stratigraphic (b) coordinates of the bipolar component observed in the Mójcza limestone. The direction in stratigraphic coordinates does not resemble any younger magnetisation direction from the region and the reversal test is positive (class C), indicating the primary nature of the magnetisation. Also shown is the cone of 95% confidence and the overall sample mean direction.

The highest unblocking temperature C direction, however, is carried by magnetite and can be clearly identified in most samples with maximum unblocking temperatures of 580°C. Is it identified as the stable end point direction in a total of 32 samples and with dual polarity (Fig. 5-6). In 7 samples C could not be isolated as stable endpoints whereas great circles are defined at intermediate to high temperatures (Fig. 5-5) and used for great circle analysis. For magnetostratigraphic interpretations the endpoint of the final great circle segment of each sample were included. This lead to transitional directions particularly at the declination values (Fig. 5-1). Although intermediate directions obtained in the stereographic projection can represent transitional directions.

Table 5-1. Paleomagnetic results from the Mójza quarry

| | | | | | In Situu | | Bedding Corrected | | | | Pole | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Lithology | Polarity | Age | bedding | N | Dec | Inc | Dec | Inc | k | $\alpha_{95}$ | Lat.(°N) | Lon.(°E) | dp/dm |
| limestone | N 1 | Car | 053/38 | 9 | 13.5 | -55.7 | 298.0 | -66.7 | 13.5 | 14.8 | 23.2 | 59.4 | 20.1/24.4 |
| limestone | R 1 | Car / Llanv | 053/38 | 15 | 196.3 | 41.1 | 152.0 | 62.9 | 24.2 | 7.9 | 8.2 | 40.5 | 9.8/12.5 |
| limestone | N 2 | Llanv | 053/38 | 4 | 337.5 | -46.6 | 297.9 | -42.7 | 12.9 | 36.0 | 3.2 | 74.2 | 27.5/44.5 |
| limestone | R 2 | Llanv | 053/38 | 7 | 203.4 | 43.1 | 155.5 | 68.3 | 62.7 | 7.7 | 14.5 | 36.1 | 10.9/12.9 |
| limestone | N 3 | Llanv | 053/38 | 3 | 355.3 | -35.0 | 322.4 | -46.2 | 5.9 | 65.6 | -4.9 | 53.6 | 53.8/84.0 |
| | ΣN | | | 16 | 1.6 | -50.3 | 305.2 | -58.4 | 10.4 | 12.1 | 11.9 | 61.1 | 13.2/17.9 |
| | ΣR | | | 23 | 198.6 | 41.8 | 153.0 | 64.7 | 30.1 | 5.7 | 10.2 | 39.2 | 7.4/9.2 |
| | N + R | Car / U. Llanv | 053/38 | 39 | 13.6 | -44.9 | 322.5 | -63.3 | 16.6 | 5.8 | 11.0 | 46.8 | 7.2/9.1 |

| Pol in situ | | |
|---|---|---|
| Lat.(°N) | Lon.(°E) | dp/dm |
| -11.7 | 8.3 | 4.6/7.3 |

R, N, Polarity: Reversed, Normal; N, number of samples; Dec, Inc, declination, inclination of Polarity Zones in degree; k, precision parameter [*Fisher*, 1953]; $\alpha_{95}$, half angle of the cone of 95% confidence in degree [*Fisher*, 1953]; Bedding is given in terms of direction of dip/dip in degrees; Age, stratigrahic age (Car, Caradocian; Llanv, Llandvirnian; U, Upper); Lat./Lon., Pole Latitude and Longitude in degree; dp/dm, semi-axis of the oval of the 95 % confidence about the mean pole; Site location (N 50,84°, E 20,69°) the pole calculations are based on.

## PALAEOMAGNETIC IMPLICATIONS

All samples of the whole section plot in two clusters. A well defined component of reversed polarity is observed in 23 samples (Table 5-1). After bedding correction this component has a direction of 153.0°/64.7°, k=30.1, $\alpha_{95}$=5.7° (Fig. 5-6b and Table 5-1). The other characteristic direction of normal polarity is observed in 16 samples in stable endpoints and great circles (Fig. 5-6b, Fig. 5-5c and Table 5-1). To calculate the overall mean direction in stratigraphic coordinates 305.2°/-58.4°, k=10.4, $\alpha_{95}$=12.1° (Table 5-1) great circles (Fig. 5-5c) and stable endpoints (Fig. 5-6b) were combined using the method of [*McFadden and McElhinny*, 1988]. The reversal test of [*McFadden and McElhinny*, 1990] was applied to the bedding corrected sample directions over the whole well defined stratigraphic section (Table 5-1). The deviation of the antipode is less than the critical angle of 13.1° at the 95% confidence level, thus the directions pass the reversal test with classification C. This observation support the interpretation that the obtained magnetization direction is primary and linked to the well known biostratigraphic age of deposition. This yield to the overall mean direction after bedding correction of 322.5°/-63.3°, k=16.6, $\alpha_{95}$=5.8°. Based on this results a south pole position of 11.0°N, 46.8°E (Table 5-1) could be calculated. It is considered to be representative for the Malopolska Massif of the Holy Cross Moutains from Lanvirnian to Caradocian times.

## CONCLUSIONS

The above described ancient magnetic direction is of bipolar nature with northwesterly (southeasterly) declination and medium to steep negative (positive) inclination in stratigraphic coordinates. The mean direction in geographic coordinates (Table 5-1) resembles no younger magnetisation direction for the region. The positive reversal test (Classification C) is regarded as proof, thus this component is considered to be primary in origin and closely linked with the rock age. The palaeomagnetic results indicate at least five polarity changes between Middle Llanvirnian (Pygodus serra zone) and Upper Caradocian times (Amorphognathus superbus zone) and a long interval of inverse polarity from Upper Llanvirnian to Middle Caradocian (Fig. 5-1). However this interval include two sampling gaps and the presence of transitional directions (Fig. 5-1) can point at a not documented reversal. The results from the Mójzca limestone are compared with published Ordovician magnetostratigraphies [*Torsvik and Trench*, 1991] [*Trench et al.*, 1991], which are mainly based on graptolite biostratigraphies (Fig. 5-7). The conodont and graptolite biozones were tied together using a new correlation, compiled by Dr B.D. Webby in consultation with Drs R.A. Cooper, S.M. Bergström and Florentin Paris for the IGCP Project No. 410 (Ordovician diversity-Edition 2001). The results of this study are in an exellent correlation between two intervals of normal polarity in the Upper Llanvirnian (Fig. 5-7). A third normal polarity interval in the lower Caradocian, which is given by [*Trench et al.*, 1991] might have been missed during sampling, also due to the strongly condensed sampled Ordovician section. In the Middle and Upper Caradocian the obvious differences between the Polish magnetostratigraphy and the compilation of [*Trench et al.*, 1991] and [*Torsvik and Trench*, 1991] could reflect uncertainties in correlation within and between the various studied profiles, based on the different stratigraphic biozones and uncertainties in comparible biostratigraphic age control. However the magnetostratigraphy of the sampled section is in good accordance with other existing global Ordovician data and again an evidence for the primary character of the identified magnetic direction.

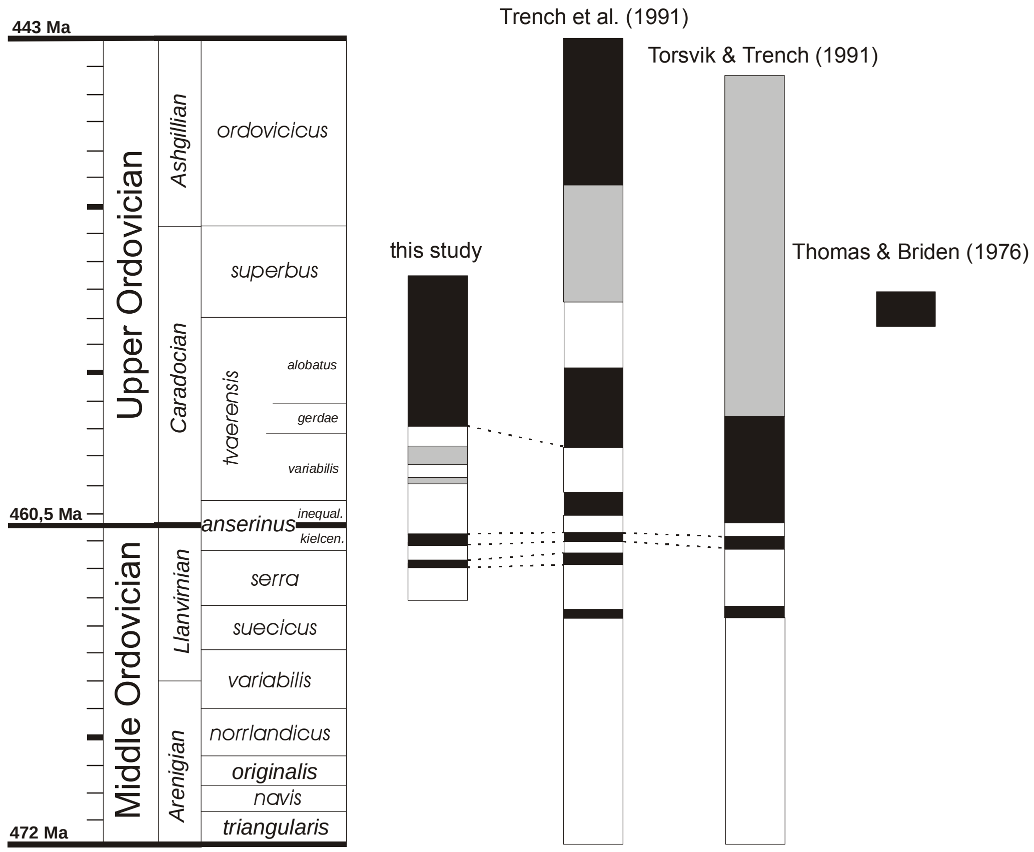



Figure 5-7: Comparsion of the results from the Mójcza limestone with published megnetostratigraphies. Black (white) intervals represent normal (inverse) polarities. Grey colours indicate uncomplete sampling.

The geomagnetic south pole of 11.0°N, 46.8°E (Table 5-1) from the Holy Cross Mountains agrees well with the Llanvirnian/Caradocian segment of the Baltic APWP after [*Smethurst et al.*, 1998] (Fig. 5-8). This Middle/Upper Ordovician palaeomagnetic south pole from the Malopolska Massif in Holy Cross Mountains does not support the hypothesis of post Ordovician large-scale tectonic movements along the TESZ and also this renders major local tectonic rotations of the investigated area rather unlikely. However these palaeomagnetic result point to a closure of the Tornquist sea as it is proposed by [*Torsvik et al.*, 1996] favouring a Mid Ordovician closure of the Tornquist sea. We believe that the attachment of the Malopolska Massif at the border of Baltica was finished before Mid Ordovician times.

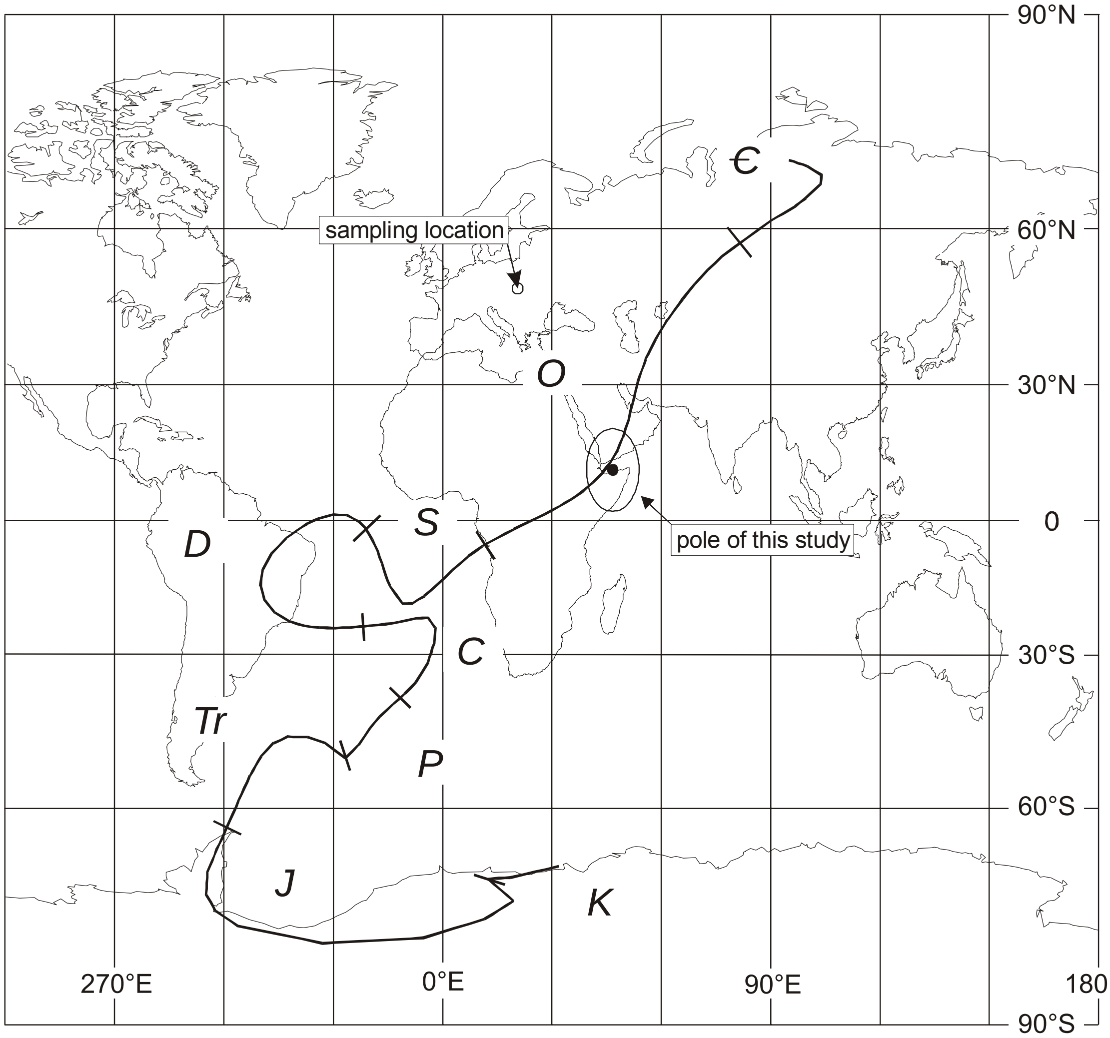



Figure 5-8: Apparent polar wander path of Baltica. The path for the south pole is based on data from [*Smethurst et al.*, 1998]. The new Middle Ordovician south pole with the 95% error oval for the Holy Cross Mountains of this study is displayed in accordance with table 5-1. Ages in the path are as follows: Є=Cambrian, O=Ordovician, S=Silurian, D=Devonian, C=Carboniferous, P=Permian, Tr=Triassic, J=Jurassic, K=Cretaceous.

# Chapter 6 Conclusions

All the new palaeomagnetic data obtained from several palaeozoic sequences in central and southern Europe and discussed in chapters 3 to 5 have important implications for the geodynamic evolution of Variscan Europe. All the identified ancient palaeomagnetic directions pass several field tests, therefore, they are interpreted as being primary and closely linked to their rock age. The resulting overall mean directions obtained in this thesis are summarized in Table 6-1. Due to the complex tectonic environment and differential thrust sheet rotation in the alpine realm, only the overall mean inclination values are quoted. For all sampled tectonostratigraphic units treated in this thesis, it was possible to translate the inclination values into palaeolatitudes which are used for palaeogeographic reconstructions as described below.

Table 6-1: Overall mean directions/inclinations determined in various sampling areas discussed in chapters 3 - 5 and their resulting palaeopole and/or palaeolatitude.

| Tectonostrat.-Unit | Segment | Geology | Age | Mean Direction/ Mean Inclination | | Statistics | | | Palaeopole | | **Palaeolatitude** | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | | | Dec. | Inc. | N | k | α95 | Lat. | Long. | | |
| Malopolska Massif | | Mojzca-limestone | Caradocian/ Llanvirnian | 323° | -63° | 39 | 16.6 | 5.8° | 11°N | 047°E | **44°S** | Chapter 5 |
| Saxothuringian Basin | Franconian Forest | Ignimbrites | 478.2±1.8[#] | 189° | 76° | 24 | 44.7 | 11.6° | 24°N | 007°E | **63°S** | Chapter 3.1 |
| | | Döbrasandstone | Ashgillian | 030° | -58° | 15 | 25.7 | 18.5° | 3°N | 349°E | **38°S** | Chapter 3.1 |
| | Saxonian Synclinorium | Ockerlimestone | Upper Silurian | 226° | 38° | 34 | 38.2 | 9.1° | 08°S | 329°E | **21°S** | Chapter 3.2 |
| Alpine realm | Upper Austroalpine | Klingler-Kar-Formation | Upper Silurian | - | 65° | 24 | 876 | 1.7° | - | - | **47°S** | Chapter 4.1 |
| | | Volcanics | Mid Devonian | - | 42° | 62 | 53.7 | 4.6 | - | - | **25°S** | Chapter 4.1 |
| | Southalpine | Limestones | Lower Devonian | - | 50° | 67 | 345 | 1.6 | - | - | **31°S** | Chapter 4.2 |

Dec./Inc., declination/inclination of site means in degrees; N, number of samples used in calculation of site mean; k, precision parameter [*Fisher*, 1953]; $\alpha_{95}$ , half angle of the cone of 95% confidence [*Fisher*, 1953]; Lat./Long., Pole Latitude and Longitude in degree; # $^{207}Pb/^{206}Pb$ age.

*Malopolska Massif –Holy Cross Mountains; attachment at the border of Baltica*

The interpretation of isotopic data of detrital muscovites and the geological environment from the HCM by *Belka et al.* [2000] demonstrate that the Malopolska Massif bordering the EEP in southern Poland is a Gondwana derived exotic terrane. During Early Cambrian time the

Malopolska Massif was still separated from Baltica. Due to Cadomian K-Ar cooling ages of mica populations out of Lower Cambrian clastics and the faunal affinity to zoogeographical provinces of Gondwana, respectively Avalonia, indicate a palaeogeographic position close to the northern margin of Gondwana. The co-occurrence of Baltic trilobites in Lower Cambrian sediments of the Malopolska Massif indicate the proximity of Baltica to the border of Gondwana (Fig. 6-1a). The motion of Baltica for Early Cambrian to Mid Ordivician times is characterized by anticlockwise rotation (approximately 90°) combined with a slow drift rate to the North [*Smethurst et al.*, 1998a] [*Pharaoh*, 1999]. In Middle and Upper Cambrian sequences in the HCM a progressive influx of Baltic brachiopods can be observed [*Jendryka-Fuglewicz*, 1998] and also the discordant overlaying Ordovician sediments in the Malopolska Massif contain faunas, which belong essentially to the Baltic province [*Dzik et al.*, 1994]. In addition Muscovites from Middle and Upper Cambrian clastic rocks of the Malopolska Massif show no more Cadomian cooling-ages, however K-Ar cooling-ages from 1.7 to 1.8 Ga [*Belka et al.*, 2000]. The source area for this detrital material must be the Svecofennian basement of Baltica [*Belka et al.*, 2000]. These data point to a scenario which favours a quick transit of the Malopolska Massif between Lower and Mid Cambrian times from the Gondwana margin across the Tornquist Sea to the southern margin of Baltica. As demonstrated by palaeomagnetic data, presented in this thesis (chapter 5), the Malopolska Massif had collided with the southern margin of Baltica by Mid Ordovician times (Fig. 6-1b). Continious sedimentation with varying geological environment at the shelfmargin of Baltica took place and characterize the further geological development of the Holy Cross Mountains.

*Palaeogeography of the Armorican Terrane Assemblage and adjacents plates*

The closure of the Tornquist Sea, separating Baltica and Avalonia, and the northern Iapetus Ocean between Laurentia and Baltica, which resulted in the Caledonian orogenic event are now well constrained (see chapter 1.2) and therefore not repeated here. From faunal and palaeomagnetic evidence the tectonic relationships of the various crustal blocks comprising the Armorican Terrane Assemblage (see chapter 1.2) remain in question. Especially it was not clear, if Saxothuringia was part of Ibero-Armorica or independent, and if it was separated from Barrandia. In the following, Saxothuringia is treated as an independent block based on the new palaeomagnetic data presented in chapter 3 of this thesis.

Most elements of Palaeozoic Europe belong to the ATA, now juxtaposed between Baltica, Laurentia and Gondwana, are characterised by Cadomian crust, thus were part of the northern Gondwana margin in the earliest Palaeozoic. Tremadocian faunas of the Armorican Terrane Assemblage also remained classically Gondwanan with the exception of Barrandia whose brachiopod and trilobite genera also show some similarities to those of Baltica [*Cocks*, 2000]. In Arenigian times, Barrandia was at 76±14°S [*Tait et al.*, 1994b], which is in agreement with palaeontological data which show continued faunal migration between Barrandia, Armorica and Avalonia [*Havlicek et al.*, 1994]. Combining the three poles available for the Armorican Massif,

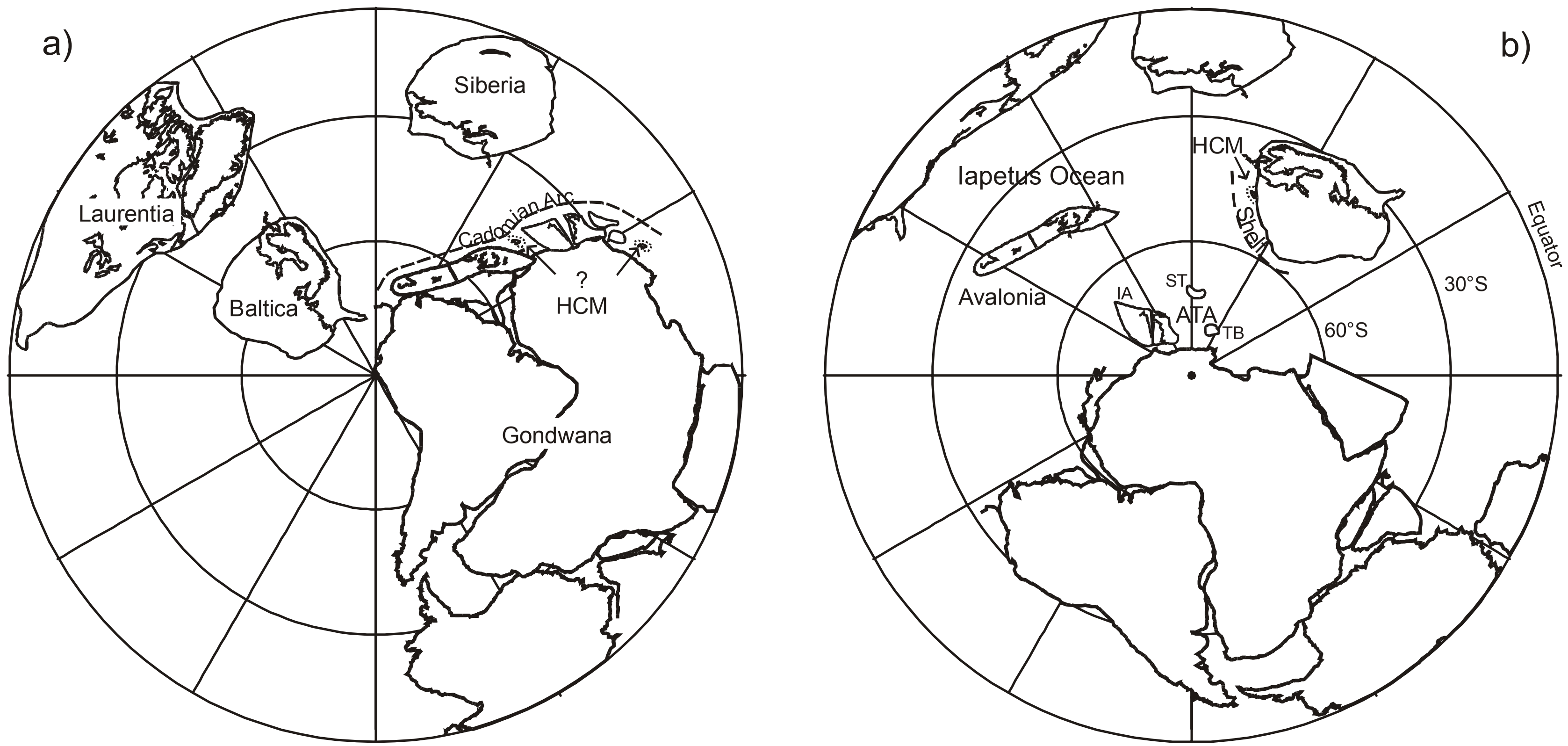



Figure 6-1: Plate reconstruction for Lower Cambrian (a) and Mid Ordovician (b) times. Baltica is positioned according to [*Smethurst et al.*, 1998a] as given in figure 5-8. The orientation of Gondwana is based on data from [*Meert et al.*, 1995] (a) and from [*Lanza and Tonarini*, 1998] (b). For Siberia see data from [*Smethurst et al.*, 1998b] and for Laurentia and Avalonia see [*Torsvik et al.*, 1996] and references therein. Amorican Terrane Assemblage (ATA: TB: Tepla-Barrandian, STB: Saxothuringian Basin, IA: Iberian and Armorican Massifs) based on data mentioned by [*Tait et al.*, 2000a] and data of this thesis (STB). In Figure 6-1b HCM is positioned according the Mójza data of this study (chapter 5).

it is placed at 65±20°S (Fig. 6-2a), i.e., similar to Avalonia and independent of Barrandia. The new palaeomagnetic data for Saxothuringia (chapter 3.1) indicate palaeolatitudes of 63° (+10.4°, -9.3°) S in the Arenig, near the northern margin of Gondwana. It is interesting to point out that this is concordant with available palaeomagnetic data for Ibero-Armorica. In conjunction with geochemical data from the volcanic rocks [*Martin et al.*, 1998] studied in chapter 3.1, which are indicative for rifting related volcanism, the new palaeomagnetic data presented here are taken as evidence for the onset of rifting of Saxothuringia from Gondwana and the start of its relative northward movement.

Caradocian faunas of the ATA show increasing similarity to Avalonia indicating narrowing of the Rheic Ocean (Fig. 6-2b), and by Ashgillian times, Anglo-Baltic faunas can be identified throughout the Armorican terrane assemblage. Late Caradocian faunas of Iberia, however, do show some differentiation to those of Barrandia [*Havlicek et al.*, 1994]. North African faunas remained of low diversity and cold water in character and distinct to those of Armorica and Barrandia. Late Ordovician palaeomagnetic data for the Armorican terrane assemblage are given from Barrandia indicating 40±9.5°S [*Tait et al.*, 1995], the Armorican Massif yielding 76±7°S [*Perroud and Van der Voo*, 1985] and the new data reported here for the Saxothuringian terrane (chapter 3.1) show palaeolatitudes of approximately 38° (+11.4°, -8.9°) S. While the apparently continued high palaeolatitudes for the Armorican Massif in the Late Ordovician is in conflict with faunal evidence, it is fairly clear that by Late Ordovician times, Barrandia and Saxothuringia were moving northwards and away from northern Gondwana. Regarding the faunal record it is tempting to say that the situation for Armorica was the same, as shown in Fig. 6-2b, but until more palaeomagnetic evidence becomes available the position of Armorica must remain debatable. A comparable situation is given for the Iberian Massif for which the faunas show certain degree of endemicity, but for which no reliable palaeomagnetic data have as yet been published.

The Late Silurian paleomagnetic results of chapter 3.2 provide qualitative data to constrain the position of the Saxothuringian basin in this time period. The resulting palaeopole of 329°E, 8°S corresponds to a palaeolatitude of 21° (+3.3°, -3.0°) S and is therefore in good agreement with the palaeolatitudes obtained from palaeomagnetic investigations of Silurian volcanic and sedimentary rocks of the Tepla-Barrandian (23°S) [*Tait et al.*, 1994a] and the paleolatitude (19°S) determined in the Armorican Massif for Early Devonian [*Tait*, 1999]. Thus by the Late Silurian/Early Devonian Bohemia, Saxothuringia, the Armorican Massif and the Catalan terrane [*Tait et al.*, 2000b] were all proximal to the southern margin of Baltica/Avalonia, and formed a chain of continental blocks (ATA) within the latitudinal belt of 20-30°S (Fig. 6-2c). Significantly, however, the new palaeomagnetic data (chapter 3.2) also demonstrate that the Teplá-Barrandian and the Saxothuringian basin did not belong to a coherent microplate. Whereas the magnetic inclinations obtained from these regions are in good agreement, the declination values differ significantly. The Bohemian results indicate large scale anticlockwise rotation of the Bohemian Massif between Latest Silurian and Late Devonian times. Taking into account the present investigations of Saxothuringian carbonates which yield to a mean direction of 46°/-38°, it is now clear that these rotations do not concern the Armorican microplate as a whole. The rotation amount of about 165° also is not recognized in the Early Devonian results of the Armorican Massif [*Tait*, 1999]. Therefore it is supposed that there is a major tectonic discontinuity within the ATA. With it the assumption of a Armorican terrane assemblage proposed by [*Tait*, 1999] is confirmed by these data.

Faunal evidence demonstrates that the closure of the Rheic Ocean between the ATA and Avalonia did not occur until the Mid Devonian (Fig. 6-2d). Final suturing and consolidation of the ATA, closure of the Saxothuringian basin, and rotation of Bohemia into its present day orientation had occurred in Late Devonian times (Fig. 6-2e) [*Tait et al.*, 1997]. Thus, from lithological, faunal and palaeomagnetic data, the accretion of Gondwana derived terranes to the southern margin of Laurussia was essentially an on-going process throughout the Palaeozoic era.

*Palaeogeography of the Proto-Alps and the final formation of Pangaea*

Due to poly-metamorphic and poly-deformational events in the Alps, interpretations of the Palaeozoic palaeogeography and geodynamic evolution of the Alpine realm remain uncertain and still in question. However, the accretion and amalgamation scenario of the Alpine units onto the southern margin of Laurussia would have important implications for the closure of the Palaeotethys with the final collision of Gondwana. Thus the new palaeomagnetic data from Palaeozoic sequences of the Eastern Alps (chapter 4) yield to important implications for the geodynamic evolution of Variscan Europe.

No palaeomagnetic data are as yet available for the Proto-Alps in Ordovician times, however on faunal and lithological considerations it is generally accepted that the Palaeozoic units of Alpine Europe having been part of the northern margin of Gondwana throughout the Cambro-Ordovician [*Antonitsch et al.*, 1994; *Dallmeyer and Neubauer*, 1994; *Handler et al.*, 1997; *Neubauer and Raumer*, 1993; *Schönlaub*, 1992; *v. Raumer*, 1998] (Fig. 6-2a). Rifting related bimodal volcanic activity, documented in the Northern Greywackezone, the Gurktal nappe and in the Carnic Alps [*Läufer et al.*, 2001; *Loeschke*, 1989; *Loeschke and Heinisch*, 1993; *Schönlaub and Histon*, 2000] indicate the beginning of rifting and separation of the southern part of the Proto-Alps (Noric-Bosnian Terrane) from the northern margin of Gondwana in Late Ordovician (Fig. 6-2b).

Using the reversed polarity option the new palaeomagnetic data demonstrate that, for Upper Silurian times, the Noric-Bosnian Terrane (NBT) was at a palaeolatitude of 47° (+2.3°, –2.1°) S. This is in general agreement with the position estimated from palaeobiogeographic evidence [*Schönlaub*, 1992] and documents the gradual northward drift of this Alpine terrane from Cambro-Ordovician to Silurian times (Fig. 6-2c). The Lower Devonian results from the Carnic Alps with palaeolatitudes of 31° (+1.3°, –1.7°) S confirm the continuous movement to the North. In the more northerly Lower and Middle Austroalpine units, Silurian to Devonian magmatic and metamorphic events document northerly directed subduction beneath southern Laurasia [*Neubauer et al.*, 1999]. There is no geological evidence, however, for any pre-Carboniferous orogenic event in the Southern Alps (NBT) where continuous sedimentation from Ordovician to

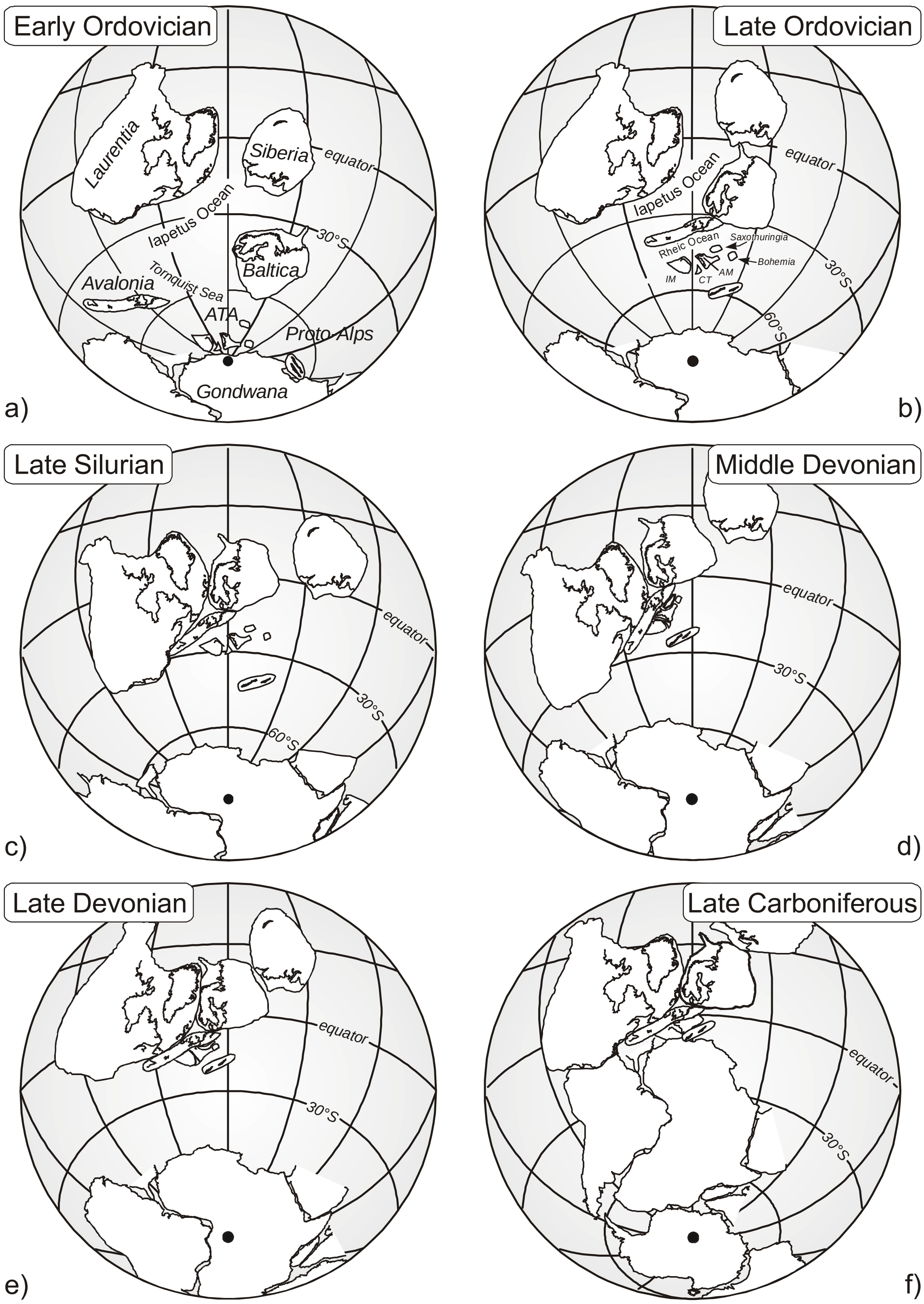



Figure 6-2: Palaeogeographic reconstructions based on palaeomagnetic data. Proto-Alps and ATA based on data mentioned in the text, for Avalonia see data compilation of [*Torsvik et al.*, 1993], for Baltica see review paper of [*Torsvik et al.*, 1992], for Laurentia see [*MacNiocaill and Smethurst*, 1994], and for Siberia see review paper by [*Smethurst et al.*, 1998b]. The palaeopoles used for Gondwana are based on data of [*Corner and Henthorn*, 1978], [*Lanza and Tonarini*, 1998], [*Schmidt and Embleton*, 1990], [*Hurley and Van der Voo*, 1987] and [*Daly and Irving*, 1983] (Late Silurian and Mid-Devonian poles are interpolated from the Late Ordovician [*Schmidt and Embleton*, 1990] and Late Devonian [*Hurley and Van der Voo*, 1987] palaeopoles). IM; Iberian Massif, AM; Armorican Massif, CT; Catalan terrane.

Lower Carboniferous times is recorded [*Neubauer*, 1988; *Neubauer et al.*, 1999; *Schönlaub and Heinisch*, 1993; *Stampfli*, 1996].

For Mid Devonian times the new palaeomagnetic data presented in this study translate into a palaeolatitude of 25°S (+3.7° and –3.4°) for the southern Proto-Alps, and document the continued northward movement of the Proto-Alps relative to Laurasia and narrowing of the intervening ocean (Fig. 6-2d). This scenario is broadly reflected in the lithological record. Deposition of siliciclastic sediments continued in a passive margin type setting [*Heinisch*, 1988; *Schönlaub and Heinisch*, 1993], and the development of carbonate platforms and widespread Late Devonian reef buildups reflect warmer water conditions in the Wildseeloder Unit in the Northern Greywackezone and the Plöcken area in the Carnic Alps [*Schönlaub and Heinisch*, 1993]. Faunal evidence also supports this palaeogeographic scenario, showing an increase in faunal exchange between the southern Alpine region and Laurussia, the Urals, Avalonia and the ATA in the Mid Devonian, but poor exchange of faunas between the Southern Alps and northern Gondwana [*Schönlaub*, 1992] indicating the separation from the northern margin of Gondwana.

Comparing these new data with palaeomagnetic data from the ATA, it is clear that the NBT was at significantly higher palaeolatitudes (Fig. 6-2). The Alpine region, therefore, did not belong to the ATA, but formed an independent microplate, or assemblage of terranes which followed similar drift to the ATA, but were separated by an ocean of some 1000 km width.

The tectonic relationship between the southern Proto-Alps and northern Gondwana, whether or not they remained adjacent to northern Africa throughout the Palaeozoic, is more difficult to ascertain from the palaeomagnetic evidence alone due to the conflicting database for Gondwana. From the existing palaeomagnetic data essentially two contrasting models for the Palaeozoic apparent polar wander (APW) path of Gondwana have been proposed (for discussion of this problem see [*Bachtadse et al.*, 1995; *Tait et al.*, 2000a; *Van der Voo*, 1993]). Details of these models, termed X and Y, are outwith the scope of this thesis, however, the main difference between these models is with regards to the position of Gondwana in Latest Silurian/Early Devonian times. Depending on which model is adopted, the northern margin of Gondwana is either positioned in low (Y) equatorial, or steep southerly (X) latitudes. Combining the Siluro/Devonian data of the alpine region obtained here (chapter 4) with equatorial position of the northern margin of Gondwana (model Y) results in a continental overlap with the Proto-Alps and the European terranes. In order to allow for the equatorial position of northern Africa and to avoid continental overlap, the Alpine terranes have to be rotated along a line of latitude. Whilst this is permissable from the palaeomagnetic data, which cannot constrain the palaeolongitude of the continents, it would require some rather complicated tectonic scenarios to allow this rapid and major relative westward movement of the terranes. There is also no faunal evidence in support of this model. The Mid Devonian faunas of the alpine regions show fairly close similarities to those of the Urals, but very little similarity to Gondwana [*Schönlaub*, 1992]. In conjunction with palaeoclimatic models for Gondwana [*Caputo and Crowell*, 1985; *Scotese et al.*, 1999] this leads to the opinion that the model X which kept the northern margin of Gondwana at higher latitudes is the more likely scenario, but clearly more data for the position of Gondwana have to be required for the pertinent time periods.

During the Lower Carboniferous the southern tectonic elements of the Proto-Alps (Noric-Bosnian Terrane) collided with alpine elements, which were accreted along the northerly adjacent Laurussian [*Neubauer and Handler*, 2000] (Fig. 6-2e). The collisional tectonics are recorded by proximal flysch-type sedimentation in the Southern Alpine units (e.g. Hochwipfel flysch) [*Läufer et al.*, 2001], and final accretion with the margin of Laurussia occurred in late Carboniferous times [*Neubauer and Handler*, 2000] (Fig. 6-2f). North-vergent subduction of the Paleotethys beneath the active Laurussia margin continued until Gondwana finally collided with Laurussia in the Late Carboniferous to Early Permian thus forming Pangaea.

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