Karoo-Etendeka Unconformities in NW Namibia and their Tectonic Implications

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ABSTRACT

In north-western Namibia the fills of the Karoo-Etendeka depositories can be subdivided into (1) a Carboniferous-Permian, (2) a Triassic-Jurassic and (3) a Cretaceous megasequence, each recording extensional periods related to successive rifting phases in the evolving South Atlantic. The tectonic environment of the depositories in north-western Namibia changes successively from the coast towards the continental interior, which is reflected by the facies distribution and the position of time-stratigraphic gaps. Close to the present-day coastline synsedimentary listric faults, trending parallel to the South Atlantic rift (N-S), caused the formation of wedge shaped sediment bodies. Here, the Karoo Supergroup is only represented by the Permian succession in the Huab area. A hiatus within the Permian can be recognised by the correlation with the main Karoo Basin in South Africa and the Brazilian Paraná Basin. This stratal gap correlates with a pre-Beaufort Group unconformity in the main Karoo Basin that might be related to an orogenic pulse in the Cape Fold Belt. The Permian succession itself is unconformably overlain by the Lower Cretaceous Etendeka Group. This hiatus extending from the Upper Permian to the Lower Cretaceous has probably been induced by a combination of rift shoulder uplift and additional crustal doming associated with Etendeka flood volcanism. The enhanced tectonism during the Early Cretaceous controlled accommodation space for the alluvial-fluvial and aeolian deposits of the lower Etendeka Group.
Disconformities within those deposits and the overlying lava succession attribute to distinct phases of tectonic and volcanic activity heralding the South Atlantic breakup.

Towards the south-east, the Karoo succession becomes successively more complete. In the vicinity of Mt. Brandberg Early Triassic strata (Middle Omingonde Formation) follow disconformably above the Upper Permian/Lowermost Triassic Doros Formation. The sedimentation there was essentially controlled by the SW-NE trending Damaraland Uplift.
South of the Damaraland Uplift the SW-NE trending Waterberg-Omaruru Fault zone is interpreted as a sinistral oblique-slip fault that compartmentalised the South Atlantic rift. This fault controlled accommodation space of the entire Triassic Omingonde Formation and the Early Jurassic Etjo Formation in its associated pull-apart and transtension structures. A locally well developed angular unconformity defines a hiatus between the two formations. Correlation with the main Karoo Basin in South Africa confirms that this gap is of a regional extent and not only a local, fault induced feature. Furthermore, it might also correlate with an orogenic pulse of the Cape Fold Belt.

In general, the Mesozoic megasequences record the long-lived history of the southern Atlantic rift evolution. Rifting has been controlled by orogenic pulses derived from the Samfrau active margin throughout the Mesozoic. The associated intracratonic E-W extension caused the formation of grabens and conjugated oblique-slip zones. The generation of voluminous flood basalts marks the climax of intracratonic extension that was accompanied by enhanced uplift of the rift shoulders.

Zusammenfassung

Das Mesozoikum in Nordwest-Namibia gliedert sich in (1) eine karbonisch-permische, (2) eine triassisch-jurassische und (3) eine kretazische Megasequenz, welche die Entwicklung des südatlantischen Rifts widerspiegeln. Die tektonische Position der karoo-zeitlichen Ablagerungsräume Nordwest-Namibias verändert sich sukzessive von der Küste bis ins Innere des Kontinents, welches sich in der Verbreitung und Größenordnung mehrerer Hiaten ausdrückt. So dominieren in der Küstenregion listrische Störungen, die dem N-S Trend des südatlantischen Rifts folgen. Dabei zeigen keilförmige Geometrien assoziierter Sedimentkörper den syn-sedimentären Charakter dieser Störungen an. In der küstennahen Region ist die Karoo Abfolge nur durch permische Sedimente in der Huab Region überliefert. Eine Diskordanz innerhalb des Perms läßt sich aus der Korrelation mit zeitäquivalenten Ablagerungen im Großen Karoo Becken Südafrikas und dem Paraná Becken Südamerikas herleiten. Wahrscheinlich hängt dieser Hiatus mit einem orogenen Impuls im Kap-Ventana Faltengürtel zusammen.

Die permische Abfolge ist wiederum diskordant überlagert von der unterkretazischen Etendeka Gruppe. Dieser Trias und Jura umfassende Hiatus ist vermutlich durch die Anhebung der südatlantischen Riftschulter verursacht worden. Die damit verbundene hohe tektonische Aktivität drückt sich deutlich in der Fazies- und Mächtigkeitsverteilung der unterkretazischen sedimentären Ablagerungen und Laven aus. Dabei spiegeln mehrere Diskonformitäten einzelne Phasen tektonischer und vulkanischer Aktivität wider, welche die bevorstehende Öffnung des Südatlantiks ankündigen.

Südöstlich der Huab Region wird die Karoo Abfolge zunehmend vollständiger. So ist im Bereich des Brandberges bereits die triassische Omingonde Formation vertreten, während noch weiter im Südosten die unterjurassischen Sandsteine des Waterberg Gebietes hinzukommen. Die Sedimentation in diesem Raum wurde maßgeblich durch das SW-NO streichende Damaralandhoch beeinflußt. Südlich des Damaralandhochs stellt die prominente SW-NO streichende Waterberg-Omaruru Störungszone eine Transferzone dar, die das Südatlantikrift unterteilt. Entlang dieser Transferzone erreicht die triassische und jurassische Abfolge ihre höchsten Mächtigkeiten, welches durch eine erhöhte Subsidenz innerhalb verschiedener Extensionsbecken begründet ist.

Eine entlang der Transferzone zu beobachtende Diskordanz zwischen der triassischen Omingonde Formation und der unterjurassischen Etjo Formation ist nicht nur von lokaler Bedeutung, sondern korreliert mit einer entsprechenden Diskordanz im Großen Karoo Becken Südafrikas. Letztere läßt sich wiederum einem orogenen Impuls im Kap-Ventana Faltengürtel zuordnen.

Chapter 1: Introduction

The Karoo Supergroup embraces Late Carboniferous to Early Cretaceous strata, which are widespread in various depositories across the Gondwana supercontinent (Fig. 1-1). Of these depositories, two major basin types are developed in southern Africa (Fig. 1-2): (A) the main Karoo Basin in South Africa that developed as a foredeep in front of the northwardly prograding Cape Fold Belt and (B) the elongate grabens and halfgrabens in southern and particularly eastern Africa that reactivated pre-existing basement shear zones (Daly et al., 1989).

In South America, the Paraná Basin represents the main depocentre for Karoo-equivalent deposits. In both, the main Karoo Basin and the Paraná Basin, Karoo-equivalent strata attain maximum cumulative thicknesses of 10 000 m (Visser, 1996) and 6 000 m (Zalán et al., 1990), respectively. This contrasts with the Karoo in Namibia were only a maximum thickness of 700 m is reported. Gaps within the Namibian Karoo Supergroup were presumed by Porada et al. (1996) and Miller (1997). The most obvious gap between Permian and Cretaceous strata in the Huab area has been used by Milner et al. (1994) as a basis for stratigraphic separation of the Carboniferous-Jurassic Karoo Supergroup and the Cretaceous Etendeka Group. Stollhofen (1999) and Stanistreet & Stollhofen (1999) relate these contrasting stratigraphic records to the development of major unconformities expressing successive phases of extension and thermal uplift along the early southern South Atlantic rift zone. The latter ultimately led to oceanic onset and continental separation of South America from Africa during the Early Cretaceous.

Other publications dealing with Mesozoic time stratigraphic gaps concern the main Karoo Basin in South Africa (Turner, 1999) and the South American Paraná Basin (Milani et al., 1994; França, et al., 1995), but more detailed studies for northern Namibia are lacking.

It is the aim of this thesis to analyse these gaps in the stratigraphic record and to place them in a tectonic context. The study focuses on the Huab and Waterberg areas in north-western Namibia (Fig. 1-3), because of the advantages of good exposure and the availability of outcrops in different tectonic settings. The Huab area represents a marginal rift-basin setting at the eastern side of the evolving South Atlantic rift, whereas the Waterberg area is placed along an intracratonic fault zone compartmentalising this rift. A major approach of this thesis is a comparison of the anatomy, time-stratigraphic relevance and spatial extent of Mesozoic unconformities in either tectonic settings. Another approach is to evaluate the regional significance of these unconformities by correlating the sequences of the study area with time-equivalent sequences in the South African main Karoo Basin and the Brazilian Paraná Basin. Finally, the genesis of the Mesozoic unconformities is discussed in the context of local and regional tectonic events.

Fig. 1-1

Palaeozoic to early Mesozoic reconstruction of Gondwana including the distribution of Karoo deposits (compiled from De Wit et al., 1988; De Wit & Ransome, 1992 and Stollhofen, 1999). Ages of the axes of successful continental separation are from Harland et al. 1990.

Fig. 1-2
The distribution of Karoo deposits in southern Africa with distinction between volcanic and sedimentary rocks (modified from Miller & Schalk, 1980 and Stollhofen, 1999).

Fig. 1-3
Overview map showing the Karoo and Etendeka outcrops and the main fault systems in north-western Namibia. Map compiled from Miller & Schalk (1980) and Milner (1997). The frame gives the location of detailed maps given in Figures 1.2-3, 2.1-4 and 5.3-11.

1.1 PHYSIOGRAPHY

The investigated areas are located between 13°00'-18°00' East in longitude and 19°30'-21°30' South in latitude. This section extends from the South Atlantic coast to 500 km inland and follows the climatic zonation from the hyper-arid Namib Desert in the west to the vegetated semi-desert environment in central Namibia.

The coastal area is marked by the Great Escarpment, which comprises a step-wise relief of more than 1000 m from the present coast to the continental plateau (Fig. 1.1-1). A gap within this escarpment extends from Swakopmund in the south to the northern margin of the Huab River valley. Within this zone the Albin Ridge (Fig. 1-3) is the only distinct elevation running parallel to the coast.

The Huab region subdivides into the pediment, that, starting from the coast, continuously raises until 600 m altitude over a distance of 30 km and, farther inland, into a deeply incised area characterised by numerous table mountains. The upper catchment of the Huab River is located in the eastern table mountain region and the easterly attached areas.

The southern margin of the Huab area partly drains into the Ugab River, which incises deeply into basement rocks. Farther south the Goboboseb Mts. occur as isolated hills standing on a wide plane. There the scenery is dominated by Mt. Brandberg, which is the highest mountain in Namibia with 2573 m altitude (Fig. 1.1-1).

Another large mountainous complex, the Erongo Mts., appears 100 km farther south-west. Towards the north-east several table mountains occur within an approximately 100 km wide strip, the most remarkable of them are Mt. Etjo and the Große Waterberg Plateau.

Localities in the coastal region providing sufficient outcrops exposing sedimentary and volcanic strata are the Albin Ridge, an incised section at the Sanianab River and a small area 20 km north of Terrace Bay (Fig. 1-3). The Huab area provides good outcrops along the slopes and cliffs of numerous table mountains and in the vicinity of Bloukrans 512 Farm. The outcrop conditions of the Goboboseb Mts. are relatively poor, but a good section is located at the south-western slope of Mt. Brandberg.

The Lions Head peak area at the south-eastern margin of the Erongo Mts. exposes a well preserved section, as do the slopes of Mt. Etjo, Mt. Kleiner Waterberg and Mt. Großer Waterberg. Worth mentioning is Dinos Farm (a portion of Otjihaenamapereo 92 Farm) in the vicinity of Mt. Klein Etjo, where a portion of the Waterberg Fault zone is clearly exposed (Fig 1-3).

Fig. 1.1-1
Topography of the study area in north-western Namibia based on a digital elevation model (data obtained from the US-Geological Survey, public database online).

1.2 GEOLOGICAL BACKGROUND

1.2.1 STRATIGRAPHY

The study area exposes a Carboniferous-Permian, a Triassic-Jurassic and a Cretaceous megasequence resting on folded Precambrian basement (Fig. 1.2.-1). The first two megasequences belong to the Karoo Supergroup, the Cretaceous megasequence is the Etendeka Group. Each megasequence is confined by distinct disconformities which reveal time-stratigraphic gaps.

The basement rocks belong to the Precambrian Damara orogen (Miller, 1988), comprising chlorite-muscovite-schists, meta-grauwackes and marbles. They host syn- and post-orogenic granite complexes of the Salem-Suite (Porada, 1989).

The Carboniferous-Permian megasequence comprises diamictites, mudstones, limestones and sandstones, which record the transition from a glacio-marine (Horsthemke et al., 1990), via a fluvio-marine stadium (Holzförster & Stollhofen, in press) to an entirely continental-lacustrine environment (Horsthemke et al., 1990; Stollhofen et al., 2000). The latter two facies are separated from one another by a hiatus. The Triassic-Jurassic megasequence is completely continental and is dominated by fluvial deposition with aeolian deposits on top. Within this megasequence another hiatus is assumed at the Triassic-Jurassic boundary.

The Cretaceous megasequence includes the basaltic to intermediate volcanics of the Etendeka Group with minor fluvio-alluvial and aeolian sediments at their base (Jerram et al., 1999b). Several disconformities have been observed within both, the sedimentary and the volcanic section of this megasequence, but none of them reveals a significant time-stratigraphic gap.

1.2.2 BASEMENT FABRICS

Southern Africa comprises several cratonic platforms being surrounded by orogenic belts (Fig. 1.2-2). In western Namibia and eastern Brazil the late Proterozoic closure of the proto-South Atlantic resulted in the NNW-SSE striking Damara-Ribeira belt (Porada, 1979), of which the coastal branch in north-western Namibia is termed the Kaoko belt (Miller, 1983). The attached SW-NE trending inland branch of the Damara orogen in Namibia formed by almost contemporaneous (600-535 Ma; Hoffmann, 1991) collision of the Kalahari Craton with the Congo Craton (Miller, 1983; Porada, 1989). The corresponding NE trend of the Proterozoic basement can possibly be traced into the area of the East African rift system, where N-S, NW-SE and NE-SW trends prevail.

In southern Namibia south-westerly fabrics of the Gariep belt represent the southern part of the Pan-African mobile belt, which was also formed during the Damara orogenic phase (Porada, 1979).

The study areas in north-western Namibia are either located directly on the SW-NE trending Damara inland branch (Erongo Mts., Mt. Etjo, Waterberg area, Goboboseb Mts.) or in the transitional zone where the inland branch is juxtaposed to the NNW-SSE orientated Kaoko belt (eastern Huab area). Only the outcrops in the coastal region (western Huab, Sanianab, Terrace Bay) lie directly within the Kaoko belt.

1.2.3 REACTIVATION OF NEO-PROTEROZOIC STRUCTURES

Daly et al. (1989) emphasise the influence of reactivated Precambrian basement anisotropies on the Phanerozoic rift basin evolution in southern and eastern Africa. Phanerozoic rifts are chiefly accommodated along cratonic margins, highlighting the reactivation of Pan-African suture zones. Daly et al. (1989) distinguish two modes of reactivation: (1) Dip-slip reactivation of gently dipping fabrics and structures and (2) strike-slip reactivation of steeply dipping fabrics and structures. Both modes controlled Mesozoic structures in the study areas: The Waterberg-Omaruru Lineament for example, refers to steeply dipping basement fabric, which became reactivated as a transfer fault in the Early Triassic. The Purros Fault zone (Ambrosiusberg Fault in the Huab area) acted as a thrust fault during Pan-African orogeny and became later reactivated as a normal fault. Prior to continental breakup the Purros zone confined the rift shoulder to the South Atlantic rift during its long-lasting influence on Mesozoic sedimentation (Stollhofen, 1999).

1.2.4 THE STRUCTURAL FABRIC OF THE STUDY AREAS

The major Waterberg-Omaruru Fault zone and the Autseib-Otjohorongo Fault system (Fig. 1-3) trace the SW-NE trend of the Damara inland branch (Fig. 1.2-2). Therefore Miller (1983) attributes the SW-NE orientation of these fault zones to deep crustal basement anisotrophies that follow concordantly the general basement fabric. A Mesozoic reactivation of these inherited anisotrophies controlled the accommodation space for the Mesozoic sedimentary strata in the vicinity of the Waterberg-Omaruru and Autseib-Otjohorongo lineaments (Fig. 3.4-1) (Porada et al., 1996; Holzförster et al., 1999).

Another important SW-NE trending structure, the Huab Fault, is situated just south of the Cretaceous Etendeka Plateau and is partly covered by it. This fault has been proposed by Mountney et al. (1998) to form the tectonically active north-western margin of a halfgraben-shaped depository that contains rather thick Permian and Early Cretaceous stratigraphic records, but with the Triassic-Jurassic sequence missing. NNW-SSE trending basement structures were most probably reactivated by prominent sets of NNW-SSE striking, westerly dipping normal faults, which occur in a belt subparallel to the present coastline and extending up to about 60 km inland. The major structures are referred to as the Ambrosiusberg, Uniab, Wêreldsend, Bergsig and Twyfelfontein Faults (Fig. 1.2-3). The western two faults are generally associated with an antithetic block-rotation, that caused the development of wedge-shaped sediment bodies. The other faults farther east show almost pure vertical displacements with only subtle block-rotations. This set of five fault systems caused a tectonic zonation, which is inflected by the N-S orientation of facies belts of late Palaeozoic and Mesozoic strata.

The Bloukrans Graben towards the eastern Huab outcrop area (Fig. 1-3) represents a subordinate, roughly E-W trending structure characterised by an extraordinary thick Lower Permian sequence. In contrast, Karoo outcrops in the vicinity of the Brandberg and Messum intrusive complexes do not reveal a relationship to any obvious tectonic structure on first sight. However, the inline array of the post-Karoo alkali intrusive complexes Messum, Brandberg and Okenyenya (Fig. 1-3) and the tectonic influence on Triassic (?) Karoo deposition in the Okenyenya area (Miller, 1980) may suggest a reactivation of a deep crustal NE-SW trending basement anisotropy.

1.2.5 THE MESOZOIC MEGASEQUENCES IN A GONDWANA CONTEXT

The decay of the Gondwana Supercontinent commenced with the late Palaeozoic rifting that ultimately led to the Jurassic breakup of West Gondwana (South America and Africa) from East Gondwana (India, Antarctica and Australia) followed by the Lower Cretaceous separation of Africa from South America and India from Antarctica (Harland et al. 1990). In south-western Gondwana continental breakup was heralded by the Jurassic (Duncan et al., 1997) and Lower Cretaceous (Milner et al., 1995b) effusions of voluminous flood basalts.

The complex interplay between sedimentation and extensional tectonics is recorded by corresponding Upper Carboniferous to Lower Cretaceous deposits. In southern Gondwana initial plate tectonic control on Mesozoic basin development may be viewed in terms of the Samfrau active margin south of Gondwana which developed a flexural deformed foreland basin (Tankard et al., 1982). The NW-SE and NE-SW trending elongated graben and half-graben structures of southern Africa and South America, however, define rift systems which show widespread evidence for incremental basin extension (Burke, 1976; Daly et al., 1989). In these depositories the Mesozoic sequences accumulated in widely varying thicknesses and with a variable completeness of the stratigraphic record (Stollhofen, 1999).

In southern Africa the foredeep north of the Samfrau subduction zone is represented by the main Karoo Basin of South Africa, that accumulated an Upper Carboniferous to Lower Jurassic sedimentary sequence of 10 000 m cumulative thickness (Visser, 1996). In Namibia the Karoo depositories reveal rift-basin structures that hosted only a few hundred metres thick successions (Stollhofen, 1999). The main depositories are the Karasburg, Aranos, Waterberg, Ovambo and Huab basins (Fig. 1.2-4).

Deposition of the Karoo Supergroup in southern Africa initiated during the Permo-Carboniferous deglaciation of Gondwana (Smith et al., 1993). Subsequent melioration of the climate culminated in the transgression of the *Mesosaurus* seaway during the Late Permian (Fig. 1.2-5) (Williams, 1995, Santos et al., 1996). During the Triassic and Jurassic continental deposition established during repeated phases of extension (Cairncross et al., 1995; Porada et al., 1996; Turner, 1983; 1999), which ultimately climaxed in the extrusion of the voluminous Jurassic flood basalts (Duncan et al., 1997). A second major magmatic event created the Etendeka-Paraná Flood Basalt Province, immediately prior to the opening of the South Atlantic (Milner et al.,1995b).

Fig. 1.2-1

Comparison of the completeness of Karoo and Etendeka sections and correlatives along an E-W traverse covering (A) the Huab, (B) the Brandberg and (C) the Waterberg areas in north-western and central Namibia. Numerical time scale is after Gradstein & Ogg (1996).

Fig. 1.2-2
Map of southern Africa displaying Precambrian cratonic and orogenic zones. The distribution of Phanerozoic rift basins is almost restricted to the orogenic belts, indicating the tectonic control of Precambrian structures on Phanerozoic rift basin development. Map modified from Daly et al. (1989).

Fig. 1.2-3
Structural map and cross section of the Huab area. The main faults and dykes show NNW-SSE, SW-NE and subordinate NW-SE and E-W trends. A decrease in block tilting from W to E is apparent in the cross section. Furthermore it shows a rapid thickness increase of the Gai-As Formation to the west across the Uniab Fault. The Ambrosiusberg Fault might indicate the trace of the Lower Cretaceous rift shoulder. The definitions of western, central and eastern Huab area is given in the foot bar. Map compiled from Miller & Schalk (1980) and Miller (1988).

Fig. 1.2-4
Distribution of the Karoo Supergroup and the Etendeka Group in Namibia (modified from Miller & Schalk, 1980 and Stollhofen, 1999). Only deposits known from outcrops and suboutcrops have been considered. The real distribution of subsurface Karoo deposits is probably more extensive in north-eastern Namibia.

Fig. 1.2-5
Distribution of the "Mesosaurus Seaway" during the higher Lower Permian (after Williams, 1995 and Santos et al., 1996) and the distribution of synsedimentary active fault systems (after De Wit & Ransome, 1992 and Tankard et al., 1995). Map is redrawn and simplified from Stollhofen (1999).

1.3 RESEARCH HISTORY

The first description of Karoo sediments in the Huab region was made by Krause (1913). A stratigraphy for these sediments was first proposed by Reuning & Von Hühne (1925) according to the discovery of *Mesosaurus* fossils. Subsequent research followed by Stahl (1932), Reuning & Martin (1957) and Martin (1953, 1973, 1975, 1981).

The eastern Huab area was mapped in a 1:100 000 scale by Hodgson (1970, 1972). In 1980 the South African Committee for Stratigraphy (SACS, 1980) dedicated the South African formation names Dwyka, Prince Albert and Whitehill to corresponding formations in Namibia.

A biostratigraphic correlation of the Permian strata in the Huab region with related deposits in the Brazilian Paraná Basin and the main Karoo Basin in South Africa was constructed by Oelofsen (1981, 1987) and Oelofsen & Araujo (1983) on the basis of the nektonic reptile *Mesosaurus tenuidens*. More recent studies on the Karoo stratigraphy, facies and correlation has been done by Horsthemke et al. (1990), Horsthemke (1992) and Ledendecker (1992). The above lying Etendeka Group was first dated by Siedner & Mitchell (1975), who obtained radiometric Lower Cretaceous ages (K/Ar isochron study). More detailed geochemical and stratigraphic work is published by Milner et al. (1994; 1995a) and Jerram et al. (1999b).

Official geological maps covering the Huab-Goboboseb area are sheet 2013 Cape Cross; 1:200 000 (Miller, 1988) and sheet 2114 Omaruru; 1:250 000 (Milner, 1997).

An early description of the sedimentary rocks in the Waterberg-Erongo area was given by Cloos (1911). The term "Etjo beds" was introduced by Reuning & Von Hühne (1925). More detailed studies on the stratigraphic framework followed by Gürich (1926a), Gevers (1936), Keyser (1973) and Hegenberger (1988). A description of the Karoo subsurface geology is given by Gunthorpe (1987), who observed coals in several cores. More recent finds of tetrapods improved the stratigraphic framework (Pickford, 1995; Holzförster et al., 1999). The syn-depositional importance of the Waterberg Fault was first mentioned by Reuning (1923) and later emphasised by Hegenberger (1988) and Porada et al. (1996).

This thesis is part of the research of the Karoo Tectonics Research Group of the University of Würzburg. The Group focuses on the tectonic context of Namibian Karoo deposits, besides improvement of the stratigraphic framework. During ongoing research the group released several publications focusing on Karoo geology in Namibia (Bangert et al., 1998, 1999, in press; Gerschütz, 1996; Grill, 1997; Holzförster et al., 1998, 1999, 2000; Holzförster & Stollhofen, in press; Jerram et al., 1999a+b; Mountney et al., 1998; Stanistreet & Stollhofen, 1999; Stollhofen, 1999; Stollhofen et al., 1998, 2000; Wanke et al., 1998, 2000, in press; Warren et al., in press; White et al., 2000).

1.4 METHODS

The data base for this thesis was mainly obtained by field techniques such as logging sedimentary and volcanic sections, mapping and sampling. Particular attention has been dedicated to sedimentary structures, fossil contents and tectonic features. Structural maps have been compiled with the use of satellite images and by stereoscopic interpretation of aerial photographs.

Samples have mainly been collected for stratigraphic purposes. Accordingly fossils have been determined and tuff-beds have been dated on the basis of the isotopic composition of juvenile zircons.

1.4.1 TUFF-BED DATING

The zircon content of the sampled tuff-beds was low and individual zircon crystals small (< 90 µm). Therefore a sufficient analysis required the SHRIMP technique (Sensitive High Mass-Resolution Ion Microprobe; Compston & Williams, 1984), which is capable of analysing individual zircon grains of less than 20 µm in diameter.

In order to separate suitable zircons for SHRIMP dating the sample procedure was the following:

First the samples have been crushed and milled. With subsequent sieving the grain-size fractions 0.045 - 0.18 and 0.18 - 0.9 mm were separated and subsequently washed with demineralised water. From both fractions the heavy minerals have been extracted using heavy liquid (poly-tungstenate solution) and a paramagnetic separator (Franz techniques). A preliminary zircon concentrate was then hand-picked under a binocular microscope. For further processing preferably clean, elongate zircon crystals, which illustrated well defined pyramidal geometries, were selected.

Final selection, analysis and age calculations have been carried out by Dr. Richard Armstrong at the Research School of Earth Sciences (RSES) at the Australian National University, Canberra. There the final zircon concentrate was again hand-picked under a binocular microscope and the selected grains mounted in epoxy together with the zircon standard AS3 (Duluth Complex gabbroic anorthosite; Paces & Miller, 1989) and the RSES standard SL13. The grains were sectioned approximately in two halfs, polished and photographed. All zircons were then examined by cathodoluminescence (CL) imaging on a Scan

Electron Microscope (SEM). Through CL imaging of the zircons, hidden and complex internal structures can be more accurately determined than in normal reflected or transmitted light. Consequently, the target area for the final measurement can be more reliably selected, although inheritance in the form of complete grains washed into the ash horizon can only be assessed by analysing many grains.

The U-Th-Pb analyses were performed on the SHRIMP II at the RSES and the data have been reduced in a manner similar to that described by Compston et al. (1992) and Williams & Claesson (1987). U/Pb in the unknowns were normalised to a $^{206*}Pb/^{238}U$ value of 0.1859 (equivalent to an age of 1099.1 Ma) for AS3 using an empirical power law calibration. The U and Th concentrations were determined relative to those measured in the SL13 standard. U/Pb ages were calculated from their radiogenic $^{206*}Pb/^{238}U$ ratios with correction for common Pb made by using the measured $^{207}Pb/^{206}Pb$ and $^{206}Pb/^{238}U$ values following Tera & Wasserburg (1972) and as described in Compston et al. (1992).

All age calculations and statistical assessments of the data have been done utilising the geochronological statistical software package Isoplot/Excel (version 2.00) of Ludwig (1999). Both inheritance and radiogenic Pb loss ñ resulting in older and younger apparent U/Pb ages respectively, can complicate U-Pb zircon dating of tuffs. These effects can be quite subtle, because inherited older ages derived from xenocrystic zircons are difficult to detect, especially when those xenocrysts are not much older than the particular ash bed under investigation. Xenocrystic zircons are even more difficult to detect when they have very similar morphological and geochemical characteristics to the magmatic zircons deposited from the eruption. For this study zircons which have high common Pb contents or unusually high U (or Th) contents are rejected from age calculation, as well as analyses which have been identified as outliers via a modified 2σ set of criteria. The final ages on pooled data sets are reported as weighted means with 95 % confidence limits.

1.5 CO-OPERATION

Fossil determinations, facies analyses and petrographical analyses have been partly carried out in co-operation with other researchers. Parts of the field work has been done together with Frank Holzförster (University of Grahamstown, SA) and Dougal Jerram (University of Durham, UK). Especially field work in the Waterberg and Erongo region and subsequent interpretation of the results has been accomplished together with Frank Holzförster. Research on the Permian strata in the Huab area has been carried out with minor co-operation, whereas the research on the stratigraphy of the Etendeka Group in the Huab Outliers area was in co-operation with Dougal Jerram.

The following list documents which parts of research have been carried out in co-operation:

Co-operation with Frank Holzförster:

-Measuring and interpreting sections No. H-P3 (Fig. 2.2-1), D-P6 (Fig. 2.2-3), H-P2 (Fig. 2.2-6), D-P14 (Fig. 2.2-7), B-P1 (Fig. 2.2-9, Fig. 3.1-4), W-P4 (Fig. 3.1-6), W-P3 (Fig. 3.2-1).

-Interpretation of the XRD analyses of tuff beds (Fig. 2.2-12).

-Taking photographs of thin sections of tuff bed V (Fig. 2.2-13).

-Interpretation of aerial photographs of the Bloukrans Graben and construction of cross sections (Fig. 2.4-2).

-Interpretation of thin sections of palaeosols (chapter 2.2.2.3)
-Determination of trace fossils in the Late Carboniferous-Early Permian megasequence (chapter 2.1 and 2.2.).
-Collecting of fossil vertebrates and molluscs in the Gai-As Formation (chapter 2.2.3).
-Collecting of fossil tetrapods in the Omingonde Formation (chapter 3.1.4).

Co-operation with Dougal Jerram:
-Facies interpretations of the Lower Etendeka Group in the area of the Huab Outliers.

Co-operation with Steven White (University of Würzburg, Germany):
-Interpretation of the tectonic features observed on Dinos Farm (chapter 3.4.2.1).

The following fossil determinations have been carried out by the following researchers:
-*Ann Warren and Bruce Rubidge (University of Witswatersrand, Pretoria, SA):* Amphibians in the Gai-As Formation (chapter 2.2.3.5) and tetrapods in the Triassic-Jurassic megasequence (chapter 3.1.4 and 3.2.4)
-*Marion Bamford (University of Witswatersrand, Pretoria, SA):* Fossil wood from the Permian sequence (chapter 2.2.3.5).
-*Rosemarie Rohn (University of Sao Paulo/Rio Claro, Brazil):* Bivalves in the Gai-As Formation (chapter 2.2.3).
-*Ian Stanistreet (University of Liverpool, UK) and Franz Fürsich (University of Würzburg, Germany):* Trace fossils in the Permian sequence (chapter 2.2.2 and 2.2.3).

Dating of tuff-beds was carried out in co-operation with Richard Armstrong (Research School of Earth Sciences, Canberra, Australia). Sample preparation was done by the author, but taking CL-images, isotopic measurements and data processing has been carried out by Richard Armstrong (chapter 1.4.1).

1.6 LOCALITY DEFINITIONS

Several locality names and area definitions are often used in the following chapters. A brief description of those localities which are not shown on commercial topographic maps is given here.

-*Western Huab area:* Western portion of the Huab area (Fig. 1-3) between the present Atlantic coastline and the Wêreldsend Fault system (Fig. 1.2-3).
-*Central Huab area:* Portion of the Huab area between the Wêreldsend Fault system and the Bergsig Fault system (Fig. 1.2-3).
-*Eastern Huab area:* Portion of the Huab area east of the Bergsig Fault system including the area around Twyfelfontein (Fig. 1.2-3).
-*Bloukrans Graben:* E-W trending valley running through Austerlitz 515 Farm, Bloukrans 512 Farm and Rooiberg 517 Farm.
-*Gai-As:* Ruins of a colonial German police station located at a spring in the central Huab area (S20°46´05´´/E14°01´08´´).
-*Klein Gai-As:* Ruin and seasonal spring located approximately 5 km east of Gai-As.

-*Mt. Bruin:* Prominent peak in the western Huab area south of the Huab River (S20°44´50´´/ E13°53´15´´; summit is 585 m above see level).

-*Rhino Section:* South-western end of a SW-NE trending incised valley, 10 km north of Doros Crater (S20°40´39´´/E14°11´20´´).

-*Poiki Locality:* Isolated N-S trending ridge in the Skeleton Coast Park north of the Huab River (S20°46´35´´/E13°44´26´´).

-*Tafel Section:* Prominent exposure at the south-western flank of Mt. Brandberg, beneath a small plateau, which is labelled "Tafel" on topographic maps (S20°09´38´´/E14°24´34´´).

-*Ambrosiusberg Locality:* Southern bank of the Huab River bed, approximately 9 km east of the present coastline (S20°54´12´´/E13°33´59´´).

-*Sanianab Locality:* Southern bank of the western Sanianab River bed, approximately 500 m east of the adjacent dune field (S20°01´58´´/E13°18´07´´).

Chapter 2: The Late Carboniferous-Permian Megasequence and associated Unconformities

In north-western Namibia the Karoo Supergroup in spatially unequally distributed: In the coastal area the Karoo Supergroup is only represented by the Late Carboniferous-Permian megasequence, which is overlain unconformably by the Lower Cretaceous Etendeka Group there. Farther south-east in the Gobeboseb-Otjongundu and Erongo-Waterberg basins (Fig. 3.4-1) the Karoo Supergroup becomes successively more complete with parts of the Triassic-Jurassic megasequence being present (Fig. 1.2-1). The Late Carboniferous and Permian deposits are only exposed in the Huab-Goboboseb area and partly at Albin Ridge and at a small incised section along the Sanianab River (Fig. 1-3). Good outcrops provide numerous cliffs of the table mountains south of the Huab River. The most complete exposures are located at Mt. Bruin, at the scarps located 7 km north of Doros Crater and at the northern flank of Doros Crater (Fig. 2.1-4). More outcrops of Late Carboniferous and Permian deposits are found in the Goboboseb Mts. and at the south-western slope of Mt. Brandberg. Permian deposits of the Erongo-Waterberg area are only known from boreholes (Gunthorpe, 1987).

This study of the Late Carboniferous-Permian megasequence focuses on the Huab-Goboboseb area because of good exposures and its direct relation to the Paraná Basin in Brazil: The Huab area was part of a large embayment of the Paraná Basin (Ledendecker, 1992) that covered at least an area of 1 000 000 km² (Milani & Zalán, 1999). An up to 250 m thick condensed marginal facies developed in the Huab-Goboboseb area, which correlates with the fully developed facies, several thousand metres thick, in the central Paraná Basin (Horsthemke et al., 1990; Porada et al., 1996) (Fig. 2-1, Fig. 2.1-3).
In north-western Namibia the Late Carboniferous-Permian megasequence comprises (1) the glaciogenic Dwyka Group followed by (2) a fluvio-marine system illustrating a successive continental to marine transition and (3) a lake system recording successive shallowing and drying out (Fig. 2.2-7). The later two successions are separated from one another by a significant hiatus, which is expressed by a relatively inconspicuous para-unconformity (chapter 2.3.3).

Fig. 2-1
Stratigraphic sections comparing the Karoo and Etendeka Group in the Huab area, the Eastern Cape region of the main Karoo Basin (after Cairncross et al., 1995) and the Paraná Basin (compiled from Bigarela, 1970; Zalán et al., 1990; Milani et al., 1994; Rohn, 1994; França et al., 1995; Milner et al., 1995a).

2.1 THE LATE CARBONIFEROUS SEQUENCE

Scattered relicts of the Late Carboniferous (Bangert, et al. 1999) glaciogenic Dwyka Group have been reported from north-western Namibia (Reuning & Martin, 1957; Hodgson, 1972; Horsthemke et al., 1990). These remnants are restricted to deep U-shaped glacial valleys, incised into Damara basement (Martin, 1975, 1981). According to Martin (1961a, 1973) these valleys were incised by pre-glacial rivers draining to the west and subsequently overmodeled by ice-flow in the same direction. In the Huab area the up to 12 m thick Dwyka Group comprises rare diamictites, but dropstone-bearing deposits including rhythmites (Horsthemke, 1992) which accumulated in periglacial lacustrine and subordinate marine environments (Holzförster & Stollhofen, in press) (Fig. 2.1-1).

The Dwyka Group is exposed at a few spatially restricted places in the Huab area only. Beside section 21 (Fig. 2.1-1) only two good exposures have been found in the Huab area. One locality occurs along a shallow incised river bed, situated 2 km south of Klein Gai-As (S20°50´10´´/E14°04´51´´), the other remarkable outcrop is located at the eastern flank of a table mountain on Bloukrans 517 Farm (S20°24´43´´/E14°21´54´´).

2.1.1 LITHOLOGY OF THE DWYKA GROUP IN THE HUAB AREA

The lower part of the Dwyka succession comprises massive homogeneous and layered diamictites (Plate I-3). The clasts are subangular or slightly rounded with diameters between 5 and 100 cm. In the layered units clasts reach 30 cm in diameter and sometimes show imbrication. Interlayers of immature sandstones, and matrix- and clast-supported conglomerates occur. Further up-section medium grained sandstone bodies appear, which are wavy laminated, slightly channelised and characterised by small-scale slump folds and abundant dewatering structures. Towards the top they grade into rhythmically bedded, sandy flagstones, occasionally with thin mudstone layers draping their surfaces. Pebble-bearing horizons occur rarely. The bedding surfaces of the flagstones exhibit a rich ichnofossil assemblage including *Isopodichnus* isp., *Diplichnites* isp., *Umfolzia sinuosa*, *Gluckstadtella cooperi* and *Punctichnium namibiense*, which Ledendecker (1992) related to a glaciolacustrine prodelta environment. However, new finds of *Helminthopsis* isp., *Chondrites* isp., *Thalassinoides* isp. and possibly *Teichichnus* isp. may indicate marine influence similar to those reported from the Aranos basin in southern Namibia that comprises the same marine ichnofossil assemblage (cf. Bangert et al., in press).

2.1.2 DWYKA GROUP FACIES

The massive diamictites are interpreted as lodgement tills (Reuning & Martin, 1957), but the layered diamictites in combination with conglomerates and thin sandstone interlayers are typical for subaquatic outwash fans and massflow deposits. The sandstones and rhythmites represent glaciofluvial to glaciolacustrine environments with minor marine incursions that are reflected by the marine ichnofauna assemblage (Holzförster & Stollhofen, in press).

2.1.3 PALAEO-GEOGRAPHIC INTERPRETATION OF THE DWYKA GROUP

It is assumed that the Huab Basin was a major depression between two glaciated highlands: (1) The Ponte Grossa High in the north, which was probably completely glaciated by the Kaokoveld lobe (c.f. França & Potter, 1988) and (2) the Uruguay Shield-Windhoek Highlands in the south (Fig. 2.1-2), both contributing glaciers in westerly direction (Martin, 1961a), namely into the Paraná Basin. Retreat of the ice during the Late Carboniferous led to the infilling of the valleys with glacial and periglacial sediments in a predominantly lacustrine environment. Only during see-level highstands, which coincide with peaks in deglaciation, the glacial valleys were eventually reached by marine incursions (Holzförster & Stollhofen, in press). After deglaciation the sediments were partly removed in north-western Namibia due to isostatic rebound, while in the central parts of the Paraná Basin the correlative glaciogenic Itararé Group has not been subject of erosion and is therefore preserved with considerable thickness (maximum 1400 m: Milani et al, 1998; Zalán et al., 1990).

2.1.4 DWYKA GROUP CORRELATIVES AND TIME CONSTRAINTS

The Dwyka Group is known from several depositories in South Africa and Namibia. In the main Karoo Basin of South Africa the Dwyka Group is represented by two major facies associations: An up to 800 m thick platform facies association consisting mainly of diamictites and a more heterogeneous valley fill facies association comprising diamictites with interleaved periglacial conglomerates and shales (Visser, 1986). Visser (1997) identified four deglaciation sequences in the western part of the main Karoo Basin which he correlated with the Dwyka succession in southern Namibia. The second deglaciation sequence in south Namibia is capped by the Ganigobis mudstones, that have been dated by tuffbed zircons at 302±3 Ma (Bangert et al., 1999). This places the lower part of the Dwyka Group into the Kasimovian according to the time scale of Gradstein & Ogg (1996) despite older palynological datings placing them at the Carboniferous/ Permian boundary (Anderson, 1977). The third deglaciation sequence in Namibia correlates with the Eurydesma Transgression which is placed in the Sakmarian according to palynological dating (Visser, 1990). Instead, Bangert et al. (1999) obtained early Asselian (288.0±3.0 and 289.6±3.8 Ma) radiometric dates from tuff beds at the base of the Prince Albert Formation from the Western Cape province. Thus, according to the age determinations of Bangert et al. (1999) the Dwyka Group encloses a time span that at latest started at about 302 Ma and ended at about the Carboniferous/Permian boundary, 290 Ma before present.

In Namibia, the Dwyka Group attains a maximum thickness of 440 m in the Asab area in the south-east of the country (well Vreda; see Fig. 1.2-4 for location) (Heath, 1972). In the north it is restricted to less than 100 m thick valley-floor fills (Martin, 1981).

In the Paraná Basin of Brazil the up to 1400 m thick Itararé Group begins with continental red beds at the base becoming entirely marine towards the top (Zalán et al., 1990). Continental conditions prevailed throughout the Carboniferous in the northern half of the basin (Aquidauana Formation) whereas the southern half became successively marine influenced. Diamictites interpreted as tillites, and associated sandstones document glaciogenic sedimentation (Zalán et al., 1990). Ledendecker (1992) considers the Itararé Group of the Paraná Basin as the direct equivalent to the more marginal facies of the Dwyka Group in the Huab Basin (Fig. 2.3-1).

Fig. 2.1-1
Section of glaciogenic Dwyka Group deposits in the Huab area. Legend

Fig. 2.1-2
Permo-Carboniferous distribution of glaciated highlands in southern Gondwana. Distribution of glaciated highlands from Ledendecker (1992), distribution of faults compiled from De Wit & Ransome (1992) and Tankard et al. (1995).

Fig. 2.1-3
Isopach map of the Late Permian Passa Dois Group of the Paraná Basin in South America. Its thickness exceeds 1000 m, which contrasts to the rather condensed Upper Permian Gai-As and Doros formations of the Huab area, where only a maximum of 200 m are attained. Map redrawn from Mühlmann (1983).

Fig. 2.1-4
Simplified geological map of the Huab area. The relevant localities and the positions of the measured sections in Figures 2.1-1, 2.2-1, 2.2-2, 2.2-3, 2.2-4, 2.2-6, 2.2-7, 2.2-8 and 2.3-3 are indicated.

2.2 THE PERMIAN SEQUENCE

2.2.1 THE PERMIAN FLUVIO-MARINE SYSTEM

The fluvio-marine system of the Permian sequence is subdivided into three formations by Horsthemke (1992) (Fig. 1.2-1):

1) Verbrandeberg Formation

The deposits of the Verbrandeberg Formation are the first Karoo deposits that occupied the postglacial landscape. Sandy point bar and coal bearing floodplain deposits reflect a widespread meandering river system in a swampy flood basin environment.

2) Tsarabis Formation

Following the deposition of the Verbrandeberg Formation a widespread but not very pronounced fluvial incision took place, which is ultimately followed by the deposition of the laterally amalgamated braided fluvial sandstones of the Tsarabis Formation. The fluvial sandstones interfinger laterally and upwardly with marine, wave dominated shoreface deposits (Holzförster & Stollhofen, in press).

3) Huab Formation

The overlying Huab Formation is built of shallow marine sediments including stromatolite bioherms with flat pebble breccia formed in tidal channels between the bioherms. The entire succession became buried by landward onlapping marine mud- and marlstones documenting the peak in the overall transgressive development (Holzförster & Stollhofen, in press).

The above described lithologies and facies associations are only strictly valid for the central Huab area (Figs. 1.2-3 and 2.1-4) between Gai-As and the Bergsig Fault system, as the facies architecture of the entire Permian succession varies considerably in cross-section from the coast to the continental interior.

2.2.2 STRATIGRAPHY AND FACIES OF THE FLUVIO-MARINE SEQUENCE

2.2.2.1 Verbrandeberg Formation

<u>Verbrandeberg Formation Lithostratigraphy</u>

The Verbrandeberg Formation (Figs. 2.2-1 & 2.2-2) attains a maximum thickness of about 70 m in the area immediately north-east of the Huab Outliers.

The lower half of this succession starts with subhorizontally bedded, grey to white shales and siltstones. Intercalated coal seams and dark pelitic horizons reflect enhanced contents of organic carbon, but only sporadically plant fragments, e.g. leaves of *Glossopteris* sp. and *Noeggerathiopsis* sp. (Adendorf, written comm. 1998; Ledendecker, 1992) and lepidophyt remains (Ledendecker, 1992) are preserved. About 80 % of the succession is pedogenically modified displaying autobrecciation, rhizoliths and few mudcracks. Ferrugineous nodules, up to 20 cm in diameter, and ironstones, consisting of up to 12 cm thick hematite layers, preferentially occur close to the base of the Verbrandeberg Formation.

The upper half of the succession is dominated by shales, interleaved with thin, 5-30 cm thick channels of intraclast-bearing coarse sandstone with slightly scoured bases. Up to 45 cm thick sandy flagstones are

interleaved. They consist of thinly bedded medium grained sandstones with planar bounding surfaces and they display apparent normal grain size grading.
Towards the top, pedogenically modified horizons become abundant in the eastern area (Verbrandeberg, Twyfelfontein) and in the Goboboseb Mts. (Tafelkop-, Brandberg area). They contain abundant rootlets, ferrugineous nodules and calcareous concretions, up to 20 cm in diameter, which often record silica precipitation. Sandy flagstones are highly silicified. In the vicinity of Mt. Brandberg (e.g. Tafel Section, Fig. 2.2-9) the pelitic rocks underwent an extraordinary intensive autobrecciation.

The whole succession interfingers sporadically with up to 15 m thick coarse-grained arkosic sandstone bodies, which usually comprise up to 30 cm thick lens-shaped cross-bedded horizons of several tens of metres lateral extent. The coarse, usually trough cross-bedded bases often comprise concentrations of cobble-sized rounded quartz and alkali feldspar besides shales and coal. Up-section medium grained sandstones follow, which corresponds to a well developed upward-fining trend. The sandstones display lamination with well defined planar bedding and ripple cross-bedding. Trough and ripple cross-bedding reveals highly variable palaeo-current directions with northerly to north-westerly directions being favoured (Fig. 2.2-5).

<u>Verbrandeberg Formation Facies</u>

Bedding and grain size characteristics and relatively high contents of organic carbon point to a deposition in a vegetated floodbasin environment. The sandy parts of the Verbrandeberg Formation represent an arkosic point bar facies association, which interfingers with a coal-bearing floodplain facies association.
Palaeosol development and the decrease in organic carbon reflect an increasingly better drainage of the area towards the upper boundary of the Verbrandeberg Formation.
As a whole, the association of coal bearing floodbasin deposits partly interfingering or being overlain by coarse point bar deposits indicates a swampy meandering river system under a cool-temperate post-glacial climate. This facies is comparable to the coal bearing, mainly terrestrial Rio Bonito Formation, which is assumed as a stratigraphic equivalent in the Paraná Basin (Ledendecker, 1992) (Fig. 2.3-1).

<u>The Verbrandeberg volcanics</u>

A restricted, but conspicuous scoriaceous deposit, interleaved with the coal-bearing floodbasin deposits, is the only evidence for Permian volcanism. A good exposure provides the Verbrandeberg locality, which is directly located at the end of road D-3254 (S20°37´21´´/E14°25´07´´). The layered scoria deposit can be traced approximately 10 km to the south-west and north-east and hence it provides a useful marker horizon for correlation.
The scoriaceous deposit comprises 0.1-1.2 m thick layers of tongue-like shape. At places thin, pelitic sediments are interleaved. The original components are almost completely replaced by hematite and various clay minerals. The shape of phenocrystic pseudomorphs indicates that the original components have been plagioclase feldspar and pyroxene, pointing to a primary basaltic composition. Thus, the scoriaceous layers are interpreted as weekly welded, basaltic scoria-fall deposits, that derived from various small proximal scoria cones.

Fig. 2.2-1
Section of the Lower Permian Verbrandeberg Formation showing a dominant floodplain facies and a subordinate point bar facies (compiled together with Holzförster). Legend

2.2.2.2 Tsarabis Formation

Tsarabis Formation Lithostratigraphy
The up to 125 m thick Tsarabis Formation overlies concordantly the Verbrandeberg Formation, but at many places with a sharp, locally erosive contact. The whole succession consists of clastic material and is subdivided into three characteristic facies associations (Holzförster & Stollhofen, in press): (1) An amalgamated sheet sandbody facies association (Fig. 2.2-2) which is followed either by (2) an upward-coarsening facies association (Fig. 2.2-3), or (3) by a hummocky cross-bedded facies association (upper part of Fig. 2.2-2). In general all sandstone units become thinner and decrease in grain-size towards the western Huab area, where their lithologies become successively similar to the fine clastics of the Huab Formation. West of Mt. Bruin the Tsarabis and Huab formations show almost identical lithologies. There the Huab Formation can be separated from the Tsarabis Formation on the basis of *Mesosaurus* remains only (chapter 2.2.2.6).

Amalgamated sheet sandbody facies association
The amalgamated sheet sandbody facies association characterises the eastern domain of the Huab area, east of the Huab Outliers. There sheet-like sandbodies, up to 10 m thick, display large scale planar and trough cross-bedding within individual layers. The trough cross-bedded zones comprise numerous erosive channel bodies, 12-40 m wide, a few metres thick and highly variable in orientation. The planar cross-bedded zones comprise up to 3.5 m thick units showing low-angle planar foresets in variable orientations.
In general this facies association displays vertical and lateral amalgamation beside an upward-thinning and an upward-fining trend. The sandstones are commonly coarse-grained, immature, feldspar rich and slightly calcareous in parts.
Accumulations of large silicified tree trunks (up to 30 m long and 1.5 m in diameter) turn up in the upper part of the unit (e.g. at the Petrified Forest National Monument, for location see Fig. 2.1-4).
Up-section this unit grades either in the upward-coarsening or hummocky cross-bedded facies association.

Upward-coarsening facies association
In the eastern domain of the Huab area the sheet-like sandstone facies association is succeeded by an upward-coarsening facies association, but westwards it shows a lateral transition to the latter. The up to 30 m thick succession of upward-coarsening cycles comprises siltstones and fine to very coarse sandstones. The lower half of a cycle is dominated by plane-bedded siltstones, which exhibit intense bioturbation, dewatering structures, slump-folds and load casts. Large scale steeply northwardly and westwardly dipping foresets (20-35°) characterise the upper half of a cycle, that comprises well sorted medium grained sandstones with coarse, gritty sandstones on top. The latter are incised by trough cross-bedded coarse-clastic channel sandstone bodies up to 20 m wide and 8 m thick in the eastern domain of the Huab area, in the vicinity of Twyfelfontein and in the Bloukrans Graben (Fig. 2.1-4) .

Hummocky cross-bedded facies association

Hummocky cross-stratified up to 20 m thick sandstones occur at many places in the central and eastern Huab area. They usually follow up-section of the amalgamated sheet sandbody facies association, but in places also above the upward-coarsening facies association. Discontinuous horizons of green sandstones are sporadically intercalated. Their green colour probably originates from considerable glauconite contents. The hummocky cross-stratification is usually of large scale with wavelengths between 1 and 3.5 m. It appears in upward-coarsening units comprising siltstones and fine to coarse grained sandstones. An intense bioturbation caused by organisms of a marine ichnofacies including *Skolithos* isp., *Diplocraterion* isp., *Monocraterion,* isp., *Planolites* isp., *Arenicolites* isp. and possibly *Teichichnus* isp. strongly point to a shallow marine environment (Holzförster & Stollhofen, in press).

Large tree trunks of up to 1,2 metre in diameter appear at the top of the sandy units (Plate I-2). They are usually oriented parallel to the strike of the hummocky cross-beds, which is SE-NW in the central Huab area east of Klein Gai-As.

Western environments

In the western domain of the Huab area, west of Mt. Bruin, the three facies associations are less conspicuous and their appearance is similar to the fine clastics of the Huab Formation: The approximately 35 m thick succession comprises shales and siltstones with a thin, but characteristic conglomerate in the middle, that might indicate a maximum sea-/lake level highstand between a transgressive and regressive cycle (Fig. 2.2-4).

<u>Tsarabis Formation Facies</u>

According to Holzförster & Stollhofen (in press) the sheet sandbody facies association illustrates a broad, mobile braided stream system, as indicated by the high degree of amalgamation and lack of floodplain sediments.

The distinct upward-coarsening cycles of the second facies association are attributed to a deltaic setting (Horsthemke et al., 1990). Holzförster & Stollhofen (in press) specify a Gilbert-type delta due to the abundance of soft sediment deformation and the development of relatively steep foresets.

The hummocky cross-bedded facies association is typical for storm wave generated deposits in shoreface successions (Duke et al., 1991). Their close interrelationship to other facies and the marine ichnofossil assemblage record a facies-type that is transitional to the shallow marine Huab Formation of the central and western Huab area (chapter 2.2.2.4). The concept of a dominant nearshore position might be supported by the often observed uniform alignment of petrified tree trunks, that probably trace palaeo-shore lines.

Although the general palaeo-current directions are highly variable (Fig. 2.2-5), a general transport to a depocenter in the west can be concluded from the E-W directed facies zonation: Proximal fluvial conditions with coarse sand- and gritstones occur in the east. They change successively into deltaic sandstones and nearshore deposits until in the western domain of the Huab area shallow marine conditions with shales and siltstones prevail. The palaeo-geographic reconstruction suggests an environment in which westwardly flowing rivers deposited their clastic load in the coastal environment of the Huab Basin whereas fine-grained suspension load was distributed farther west over the entire Paraná area. There the pelagic Palermo Formation forms a correlative to the Tsarabis Formation (Ledendecker, 1992) (Fig. 2.3-1).

2.2.2.3 Huab Formation

Huab Formation Lithostratigraphy

The facies architecture and thickness of the Huab Formation vary considerably in a cross-section from the present day coastline to the easternmost occurrence of the Huab Formation at Twyfelfontein and in the Bloukrans Graben. In the western Huab area maximum cumulative thickness attains ca. 75 m. There, the Huab Formation succeeds the fine clastics of the Tsarabis Formation concordantly with a gradual transition, rather than with a distinct change (Fig. 2.2-4). In the eastern domain, the Huab Formation interfingers with the Tsarabis Formation with rather reduced thicknesses particularly at the basin margins.

Holzförster & Stollhofen (in press) identified 4 major facies associations: (1) A slightly calcareous, laminated shales facies association (Fig. 2.2-7), which is interbedded with (2) a clastic flat-pebble conglomerate facies association (Fig. 2.2-7; Fig. 2.2-6). Towards the west both are replaced by several belts of (3) a stromatolitic bioherm carbonate facies (Fig. 2.2-4; Fig. 2.2-7). Finally the entire succession was covered by (4) landwards onlapping marine mud- and marlstones recording the peak in the overall transgressive development.

Slightly calcareous laminated shales

Outcrops of the up to 20 m thick slightly calcareous shales succession are restricted to the area east of the Uniab Fault system (central Skeleton Coast Park, for location see Fig. 2.1-4). The succession is fully developed in the eastern domains of the Huab and Goboboseb areas, where it underwent intense palaeosol development. The shale facies association occurs all over the outcrop area, but it is thinning and fining towards the west.

The shales are laminated and consist of silty and fine-sandy material. Bedding surfaces are plane, but minor wavy lenticular massive beds are interleaved. Subordinate interbeds of micritic, slightly kerogenous limestone, up to 25 cm thick, reveal planar laminated and minor domal microbial-mat structures. Pebbly sandstone interbeds are characterised by erosional bases, basal trough cross-bedding, normal grain size grading and laminated top parts.

Clastic flat-pebble facies

The clastic flat-pebble facies association is restricted to the area east of the Skeleton Coast Park with good exposures available at the Bergsig Fault system and in the north-western Goboboseb Mts. In the western domain the flat pebble conglomerates form 1 to 3 well defined horizons, each less than 20 cm thick. In the central domain the clastic flat-pebble facies consists of either tabular massive flat-pebble horizons, or more locally channelised units. The latter are usually located in depressions between bioherm bodies of the stromatolitic bioherm facies association (see below). In the eastern Huab area the flat-pebble facies exhibits a 3 to 15 m thick alternation of coarse quartz- and carbonate-detritus dominated sandstones, thin mudstones and distinct flat-pebble conglomerate horizons. A general upward-fining trend and at places planar and trough cross-bedding, documenting fluvial control, are abundant. The top is characterised by conspicuous iron-oxide cemented rudaceous gritty sandstones forming amalgamated, slightly channelised tabular sheets with abundant root marks.

The usually erosive based conglomerates are chiefly clast supported, with the clast components resembling the intraformational lithologies of the shales, laminated limestones, oolites and stromatolites of the corresponding facies associations. Additionally small silicified branches and tree trunks, as well as vertebrate remains are abundant. The poorly sorted matrix comprises ooids, quartz grains and mud granules. In the eastern domain, e.g. at Doros Crater (Fig. 2.2-8), dark iron-oxide cements occur in the upper conglomerate horizons.

The top of the flat-pebble facies, either in the flat-pebble units of the western and central domains, or in the gritty sandstone unit of the eastern domain, is characterised by a bonebed containing abundant abraded ribs, vertebrae and other bones of the crocodile-like reptile *Mesosaurus tenuidens* Gervais 1865 (Ledendecker, 1992; Oelofsen & Araujo, 1987). In places, e.g. at Doros Crater, this bonebed splits into two bonebeds separated by an approximately 25 cm thick layer of sandy shales (Fig. 2.2-8).

Stromatolitic bioherm carbonate facies

In the Huab and Goboboseb areas stromatolitic bioherms occur in several, roughly N-S oriented belts. In the Huab area these belts are located in a 35 km wide zone reaching from 3 km east of the Uniab Fault to a 2 km east of the Bergsig Fault system (Fig. 2.4-1). The belts can be extrapolated into the western Goboboseb Mts. where they gradually curve to the SW. The bioherms interfinger with the above described facies associations with the flat pebble conglomerates often underlying them and/or forming channelised patches between them. Individual bioherms commonly form elongated domal bodies, with their long axis perpendicular to the belts (Plate I-1). In the Huab area bioherms are up to 4 m high and 50x20 m in plan view, whereas the dimensions in the Goboboseb Mts. decrease to bioherms of about 7x2 m in plan view and less than 1 m in height.

Most bioherms consist of a series of laterally attached domal stromatolites at the base, that are overgrown by more continuous laminae towards the top. Thus, the basal parts refer to the SH-type of Logan et al. (1964), the upper parts to the LLH-type (Horsthemke, 1992). Individual laminae may be more than 1 cm thick, containing coarse sand sized siliciclastic and calcareous material. Many stromatolites display multiple undulating erosion surfaces, which are occasionally marked by vertical boring structures of the *Trypanites*-type.

Marine mud- and marlstone facies

The marine mud- and marlstone facies association onlaps the other facies associations landwards. Maximum thicknesses exceeding 20 m are attained in the western domain of the Huab area, whereas the unit pinches out east of the Bergsig Fault system. The unit comprises similar lithologies as the underlying calcareous laminated shales facies association, but is distinct from it because of the dominance of silty to medium sand size grained platy shales with only few interbeds of sandstones. The shales display a planar lamination, but often the top-surfaces of individual layers are wavy or small-scale rippled. A high concentration of synaeresis-cracks characterise those shales in the central Huab area between the Wêreldsend and Bergsig Fault systems. The rare, up to 25 cm thick sandstone interbeds reveal slightly erosive bases but indistinct oscillation-rippled tops.

Pedogenic features in the eastern domain

The above described facies associations underwent intense pedogenic modification in the eastern domains of the Huab (e.g. at Verbrandeberg, Doros Crater [Fig. 2.2-8], Twyfelfontein) and Goboboseb area (e.g. northern Goboboseb Mts., Mt. Brandberg [Fig. 2.2-9]). The shaly facies associations are mainly affected, but where present, also the flat pebble conglomerates and stromatolitic bioherms.

Palaeosols are characterised by calcareous concretions of up to 1.5 m in diameter, concretionary horizons up to 2.5 m thick, desiccation and shrinkage cracks, rootlets, calcretes, silicretes, ferricretes and autobrecciation. In thin-section, fine root tubes are sometimes associated with the heteromorphous symbiotic micro-organism *Microcodium* (cf. Freytet & Plaziat, 1982), but pedogenic corrosion of quartz grains is not apparent.

Patches of the upper units are enriched in Uranium and Thorium (Roesener & Schreuder, 1998), that might be related to the composition of the source rocks, but additionally a pedogenic enrichment is suggested (cf. Davies & Elliot, 1994).

2.2.2.4 Huab Formation Facies

The Huab Formation records a continuation of the overall transgressive development that was initiated with the deposition of Tsarabis Formation. Towards the end of the Huab Formation this overall transgressive trend was ceased by a marked regression, that is indicated by a widespread bonebed (Holzförster & Stollhofen, in press).

Melioration of the climate is mirrored by stromatolites, ooids and the intense palaeosol formation. The poor fauna, kerogene content of the carbonates and syneresis cracks indicate restricted living conditions due to high salinity and dysaerobe conditions (Horsthemke, 1992) in a lagoonal environment.

The facies associations in the western domain refer to a shallow marine nearshore environment, whereas the eastern domain was frequently under aerial exposure as indicated by the ubiquitous palaeosol development.

Western and central domain

Stromatolitic bioherms acted as a barrier towards the open sea in the west. Their preferred E-W orientation coincides with tidal- and wave induced currents perpendicular to the palaeo-coastline (Horsthemke, 1992). Lagoonal environments were established landwards of the bioherm belts, in which laminated shales developed. The lamination is due to suspension fallout in slightly agitated water. Erosive based interbeds of coarse to gritty sandstones formed during storm events and are therefore interpreted as tempestites. West of the bioherm belts the fair weather wave basis reached the ground, expressed in oscillation-rippled surfaces. Channels between bioherm ridges have been interpreted as tidal channels by Horsthemke (1992). They host flat-pebble conglomerates, which are characterised by a high ooid content. These ooid bearing flat-pebble conglomerates support the facies concept of shallow, wave dominated water in the vicinity of the bioherms: Ooids form in very shallow, fairly agitated, carbonate saturated water probably under warm climate (Tucker & Wright, 1990). Furthermore the flat pebbles derived from desiccated layers of unlithified cohesive fine clastics that had been reworked (cf. Richter & Füchtbauer, 1981), preferentially under the influence of storm waves. Tidal flats or lagoons provide the conditions for the generation of both, ooids and flat-pebble conglomerates.

Eastern domain

Shales and mudstones besides algal laminites represent deposition in a slightly agitated, shallow water body with relatively intensive carbonate production. Oscillating water-levels resulted in the formation of palaeosols, which were reworked in flat pebble conglomerates after flooding. Autochtonous breccias give clear evidence for frequent sea level fluctuations, that refer to palustrine soils, which underwent periodically subaerial exposure (cf. Freytet & Plaziat, 1982).

Coarse sandstones, partly channelised, reveal fluvial influence. Probably the coarse material derived from a nearby delta that was temporarily flooded during storm events. The dominance of those coarse sandstones towards the top of the eastern flat-pebble facies association might indicate progradation of the delta (Horsthemke, 1992). The iron-oxide cements are interpreted as ferricretes, that either formed autochtonous or derived from adjacent reworked ferricrete crusts.

2.2.2.5 Huab Formation base level cyclicity

Sections of the above described facies associations reveal an architecture that refers to two transgressive-regressive cycles (Horsthemke, et al., 1990; Horsthemke, 1992). Best noticeable are the two cycles in the facies belts of the stromatolitic bioherms (e.g. in the vicinity of the Bergsig Fault system, Fig. 2.4-1). There the lower cycle begins with algal laminites or stromatolitic bioherms associated with thin flat-pebble conglomerates and slightly calcareous laminated shales. In this lower cycle the bioherms reach their largest dimensions. The upper half of the cycle is dominated by the landwardly onlapping marine mud and marlstone facies.

The upper cycle again begins with algal laminites and bioherms that may be associated with flat-pebble conglomerates, laminated shales, mudstones and oolites. In the western domain of the upper cycle the marine mud and marlstone facies is present, whereas the eastern domain is dominated by the clastic flat-pebble facies with the *Mesosaurus* bonebed on top.

2.2.2.6 Palaeobiological constraints and correlation

The fossil content of the Huab Formation is of special interest for palaeo-environmental interpretation and furthermore it enables correlation to time-equivalent deposits in the main Karoo Basin and Paraná Basin.

The laminated mudstones and marlstones contain a diverse ichnofauna including *Skolithos* isp., *Planolites* isp., and large exemplars of *Rhizocorallium irregulare*, the latter is exclusively known from marine environments (Fürsich & Mayr, 1981).

In the upper cycle abundant abraded bones of *Mesosaurus tenuidens* occur. This nektonic reptile is a strong tool for correlation, as it represents a short-lived species that is also found in the Vryheid and Whitehill Formation of the main Karoo Basin and the Irati Formation of the Paraná Basin (Ledendecker, 1992) (Fig. 2.3-1). A borehole log from the central Paraná Basin documents a prominent gamma-ray peak in the middle part of the Irati Formation (Milani et al., 1998; Fig. 7). Such strong gamma-ray signals are often associated with key horizons, such as maximum flooding zones (cf. Davies & Elliot, 1994). Thus the prominent gamma-ray peak in the Irati Formation could be a correlative to the upper transgression-regression cycle found in Namibia. As a whole, the correlation of the Huab Formation into the Paraná Basin is well constrained. This is an essential requirement for identifying a hiatus between the Huab Formation and the overlying Gai-As Formation (chapter 2.3.3.2).

Fig. 2.2-2

Section of the Verbrandeberg, Tsarabis and Twyfelfontein formations. The basal gritstones of the Lower Permian Tsarabis Formation conformably incise into the floodplain fines of the underlying Verbrandeberg Formation. The Tsarabis Formation is dominated by an amalgamated sheet sandbody facies association. The Permian sequence is unconformably succeeded by the Lower Cretaceous Twyfelfontein Formation. Legend

Fig. 2.2-3

Section of the Lower Permian Tsarabis Formation displaying a dominant upward-coarsening facies association (compiled together with Holzförster). Legend

Fig. 2.2-4

Section of the western environments of the Lower Permian Tsarabis and Huab formations. The contact between the two formations is lacking any erosional features. Legend

Fig. 2.2-5

Palaeo-current analyses of the Verbrandeberg and Tsarabis formations in the Huab area. In the western and central Huab area westerly directions are favoured, whereas in the eastern Huab area the palaeo-current patterns are multi-directional: Northerly directions dominate within the Verbrandeberg Formation, whereas the palaeo-current vectors of the Tsarabis Formation are rather variable with the main vector being WNW.

Fig. 2.2-6

Section of the Lower Permian Huab Formation showing a clastic flat-pebble facies association and a stromatolitic bioherm facies association (compiled together with Holzförster). Legend

Fig. 2.2-7

Section of the Permian Huab, Gai-As and Doros formations and the basal Etendeka Group (compiled together with Holzförster). The Huab Formation developed a clastic flat-pebble facies besides algal stromatolites. The Gai-As Formation comprises marginal lacustrine deposits, whereas the following Doros Formation displays a pronounced fluvial control with a superimposed shallowing and drying out trend. The Permian sequence in unconformably capped by the mass-flow, aeolian and volcanic deposits of the Etendeka Group.
Fossil findings and the stratigraphic positions of tuff beds are indicated in the column. Legend

Fig. 2.2-8

Section at the northern rim of Doros Crater comprising the Lower and Upper Permian sequence. Marginal deltaic facies associations and pedogenic overprinting occur in all formations. Legend

Fig. 2.2-9

Section at the south-western slope of Mt. Brandberg (compiled together with Holzförster). The Permian sequence shows intense pedogenesis. The Upper Gai-As Formation follows erosively above the intensely pedogenically modified Huab Formation. The Lower Gai-As Formation is missing. The Gai-As and Doros formations developed gritstones of a marginal fan delta facies. Legend

Fig. 2.2-10

Palaeo-current analysis of the delataic deposits of the Upper Gai-As and Doros formations at Doros Crater and Mt. Brandberg. North-westerly directions dominate clearly.

2.2.3 THE PERMIAN GAI-AS LAKE SYSTEM

2.2.3.1 Introduction

Initially, the term Gai-As Formation was introduced by Horsthemke (1992) for predominantly lacustrine sediments in the Huab area, but a subsequent subdivision by Stanistreet & Stollhofen (1999) separates the Doros Formation from the underlying Gai-As Formation due to the overall coarsening and drying out trend recorded in the uppermost 35 m of the stratigraphic column (Fig. 2.2-7).
The Gai-As lake deposits provide the most widespread depositional units in the study area. They can be regarded as a final phase of complete overall basin infill, which explains its aggradation and consequently widespread occurrence.

Gai-As Formation

The Gai-As Formation is separated from the underlying marine Huab Formation by a significant hiatus. The sequence consists of fine clastic lacustrine red beds and rare microbial limestones and tuff-beds.

Doros Formation

The Doros Formation comprises dominantly sheet-like fluvial sandstone units with thin lacustrine siltstones and limestones interleaved. Mudcracks, pseudomorphs after evaporite crystals and rare reworked aeolian sands preserve evidence of a rapid lowering of the lake water level.

2.2.3.2 Gai-As/Doros Formation Lithostratigraphy

The succession is up to 170 m thick and is subdivided into a pelitic lower unit and a sandy upper unit, both are best developed in the central outcrop area. The change from the Lower to the Upper Unit is always transitional. Either units comprise four lithofacies in different proportions: (1) reddish to violet siltstones and quartzitic sandstones, (2) coarse-grained maroon gritstones, and (3) calcareous laminites and stromatolites, and (4) ash-fall tuff-beds.

Gai-As Formation Lower Unit

The Lower Unit of the Gai-As Formation is widespread and it attains 35 m in thickness in the central Huab and western Goboboseb area, but it becomes rather condensed towards the east (e.g. at Verbrandeberg and Doros Crater [Fig. 2.2-8], section 25 and 26 of Fig. 2.3-3) and it is entirely missing in the vicinity of Mt. Brandberg (e.g. at Tafel Section, Fig. 2.2-9).

The strata comprise reddish to violet, mostly laminated argillaceous to silty shales, containing 1-3 cm thick tabular interbeds of normally graded medium grained sandstones and a few, 10-50 cm thick, laminated limestone horizons. Discontinuous concretionary horizons, containing disc-shaped calcareous concretions of up to 1 m in diameter, occur. The shales usually exhibit lamination and rarely oscillation ripples. Sandstones show climbing ripples, small-scale hummocky cross-bedding and sometimes oscillation ripples on top-surfaces.
At least two widespread shale horizons are associated with conspicuous concentrations of articulated and disarticulated shells of a *Terraia* molluscan fauna, which refers to the Leinzia similis assemblage *sensu* Rohn (1994). Concentrations occur in several layers, up to 4 cm thick, with a dominance of concave-down shells reflecting the influence of traction.

Gai-As Formation Upper Unit

The widespread Upper Unit of the Gai-As Formation attains 20 m in thickness in the central Huab- and western Goboboseb area, but its thickness rapidly increases west of the Wêreldsend Fault system (e.g. at Mt. Bruin) where 70 m are exceeded. Towards the south-east (e.g. at Doros Crater [Fig. 2.2-8], Tafel Section at Mt. Brandberg [Fig. 2.2-9]) the otherwise dominant lacustrine red-bed facies fringes successively with coarser fan delta deposits.

The Upper Unit of the Gai-As Formation comprises laminated reddish-violet mudstones with interbeds of reddish-violet limestones and tabular red-brown sandstones. The limestones are laminated or small-scale cross-bedded and partly stromatolitic. They occur mainly in the lower half of the unit and their beds are lenticular or tabular in shape and up to 35 cm thick. They developed micritic limestone/calcareous sandstone alternations displaying rhythmic wavy and lenticular bedding with isolated sandy current or oscillation ripples. The 1-10 cm thick, fine to medium grained sandstone beds are usually normally graded and may show load casts at their bases and oscillation-rippled tops. They are arranged in a pronounced upward-thickening and upward-coarsening architecture. Limestones occur in the lower part of the unit and they are characterised by domal stromatolites (up to 25 cm high) with interdomal areas infiltrated by oolites and oncolites.

Evidence of intense bioturbation is preserved in places by burrows of *Planolites* isp., *Skolithos* isp., *Beaconites* isp., *Plaeophycus tubularis* and *Rosselia* isp., besides larval and nematode traces. Concentrations of coprolites (some of them up to 30 mm in diameter) and remains of palaeoniscid fish (Bender, pers. comm. 1998) including scales and teeth of *Namaichthys* and *Atherstonia*, as well as disarticulated skeletal fragments of mastodonsauroid or matoposaurid amphibians occur in several sandstones and limestones (Warren et al., in press). In one widespread single limestone/sandstone couplet, shells of a *Leinzia* molluscan fauna are concentrated. Flora is preserved with imprints of plants and rootlets and small (up to 0.5 m in length) fragments and trunks of silicified wood, identified as *Araucarioxylon* sp1. and *Podocarpoxylon* spp. (Bamford, 1998).

The unit further comprises up to five tuff-beds, each 2-5 cm thick, showing a poorly developed normal grading. Tabular geometry of the tuff-beds and their conspicuously bright colour make them important chronostratigraphic markers with the lowermost and uppermost tuff-beds coinciding broadly with the base and top of the Upper Gai-As Formation.

Doros Formation

The up to 30 m thick Doros Formation succeeds the Upper Unit of the Gai-As Formation transitionally. It is restricted to hangingwall traps, where post Permian erosion was not as good facilitated as in adjacent footwall areas. The unit displays almost the same lithologies and fossil content as the Upper Gai-As Formation, but it is dominated by sheet-like sandstone beds (Fig. 2.2-7). These make up 60-80 % of the succession and individual beds are up to 2 m thick. The medium to coarse grained sandstones show either planar or low-angle scoured basal contacts and may contain low-angle accretion surfaces, plane or small-scale hummocky cross-stratification. Hummocky beds are 5-25 cm thick with wave lengths ranging between 20 and 130 cm. Several of the upper surface contacts display oscillation ripples and rare current ripples.

Fine grained clastic beds, e.g. mudstones, siltstones and limestones, are rarely interlayered. They contain abundant desiccation and shrinkage cracks and carbonate nodules. Pseudomorphs after evaporite minerals, such as halite and gypsum, were found in the top part of micritic limestone layers. The limestones are partly dolomitic and silification is common (Horsthemke, 1992).

Towards the top of the formation bright, up to 1.5 m thick lenticular beds occur sporadically. At Gai-As they display an apparent medium-scale planar cross-bedding with foreset dip angles of 35° and an unidirectional dip direction to WSW. In other places trough-cross bedding occurs. The medium size grained sandstones are very well sorted and the majority of the grains are well rounded. Thus, the rounding and sorting characteristics of the sandstones correspond to aeolian sands, suggesting that the material has an aeolian origin.

South-eastern gritstone deposits

Coarse deposits correlating with the Upper Gai-As and Doros formations occur in the eastern Goboboseb Mts. in the vicinity of Mt. Brandberg (Fig. 2.2-9). Corresponding to their conspicuous appearance in this region Porada et al. (1996) termed this unit "Goboboseb Formation" or "Sedimentary Unit III". A fairly good exposure provides the Tafel Section at the south-western slope of Mt. Brandberg. There, compared with the Rhino Section (Fig. 2.2-7), the sequence coarsens and shows pedogenic overprinting: At the base the approximately 32 m thick unit contains abundant palaeosol horizons displaying pedogenic features like mudcracks, thin rootlets and calcareous nodules. In the basal part they are interleaved with violet, mica-rich sandstones and white, fine arkosic conglomerates up to 4.5 m thick. They form channelised bodies with erosive bases, in which reworked deposits of the underlying Huab Formation occur as intraclasts. Up-section the sandstones grade into roughly subhorizontally stratified conglomerate beds, forming up to 5.5 m thick individual upward-coarsening cycles, which often show clast imbrication. Their clast spectrum embraces poorly rounded pebbles and cobbles of granite, quartz, shale, quartzite, amphibolite, limestone, abundant fine grained reddish sandstone and siltstone, abundant tourmaline and reworked clasts from the underlying strata. Palaeo-currents, deduced from clast imbrication and trough cross-beds reveal WNW directed palaeo-currents (Fig. 2.2-10).

2.2.3.3 Facies development of the Gai-As lake system

The Gai-As and Doros formations represent an overall upward-shallowing lake sequence. During the Lower Gai-As Formation a quiet far-offshore ("hemi-pelagic") lacustrine environment is recorded by laminated shales that formed from suspension fallout. The interleaved sandstone and limestone beds deposited during storm-initiated density-flows transporting siliciclastic or carbonate sands into the offshore region. Major storm events seem to have affected this part of the sequence rarely, as only a few tempestite- and turbidite layers occur in that part of the sequence.

The Upper Gai-As Formation documents an apparently upward-shallowing lake with tempestites, which is documented by the increasing amount of medium grained sandstones. They are interpreted as tempestites/turbidites due to their sedimentary structures, e.g. load casts and oscillation rippled tops. Deposition happened preferentially at water depths below fair weather wave base but above storm wave base as indicted by shallow water features such as wavy and lenticular bedding and oscillation rippled surfaces. Stromatolitic limestones reflect phases of high microbial activity, but reduced clastic input.

The succeeding Doros Formation records repeated phases of subaerial exposure and a more fluvially controlled deposition. The micritic, partly dolomitic limestone layers with halite and gypsum crystals on top are interpreted by Horsthemke (1992) as evaporites that formed in alkaline-saline lake environments. Concentration of fish remains in several layers might reflect mass mortalities during phases of very high salinity and/or oxygen shortage. Subsequent drying out is documented by abundant mudcracks. The sandstone units are interpreted as regressive sheetsands (cf. Heward, 1981) and poorly channelised fluvial sheetflood deposits. These were emplaced in a shallow, wave dominated lake margin/shoreline environment. The aeolian material, found in flat sandstone-lenses, derived probably from patchy subaerial accumulations of aeolian sands on small islands or sandbars.

Subaerial exposure features such as evaporite pseudomorphs and desiccation cracks together with the presence of aeolian material suggest a significant lowering of the lake level under a warm, relatively dry climate.

The basin margins are either characterised by condensed lake deposits in the east (e.g. at Verbrandeberg) or marginal fan deltas in the south-east (e.g. at Mt. Brandberg). The easterly condensed facies was highly influenced by lake-level changes, which resulted in numerous pedogenic horizons and widespread shoreline deposits such as hummocky cross-stratified sandstones. In contrast to this, at the south-eastern margin (e.g. at the Tafel Section of Mt. Brandberg and at Doros Crater) marginal fan deltas developed on the pedogenically modified facies, as indicated by the bedding characteristics and architecture of the upward-coarsening gritstone units. The fine grained fraction was transported farther offshore, where it fringed with the lacustrine sediments of the Upper Gai-As and Doros formations as documented north of Doros Crater (e.g. at the Rhino Section; Fig. 2.2-7). Other more distal deltaic sequences might be present at Bloukrans 512 Farm, Gudaus Hill, Sanianab Locality and at Albin Ridge (cf. Stollhofen et al., 2000).

2.2.3.4 Gai-As/Doros Formation Radiometric time constraints

The Permian lake deposits contain at least seven ashfall tuff-beds. The lower three are sporadically preserved in sections west of Mt. Bruin. Up-section two conspicuous tuff-beds appear, which can be traced from south-west of the Huab Outliers (section DP 14, Fig. 2.2-7) until Klein Gai-As and possibly farther westwards until the Uniab Fault system (for locations see Fig. 2.1-4). These two tuff-beds have a vertical distance from one another of about 2 metres at the Bergsig Fault system and of about 2,5 metres in the vicinity of the Uniab Fault system. Their position defines the transition from the Gai-As Formation to the Doros Formation (Fig. 2.2-7). In the upper half of the Doros Formation further two tuff-beds occur, which have been confidently identified in section DP 14 (Fig. 2.2-7) only.

Two of the seven mentioned tuff-beds (sample B18 and 30/98) have been dated by analysing the isotopic composition (U; Pb; Th) of juvenile volcanic zircons with the SHRIMP technique. The calculated ages are 265.5±2.2 Ma and 272.7±1.8 Ma. Although these two dates do not conform very well, they give a much better time constraint than biostratigraphy (chapter 2.2.3.5).

The tuff-beds (Fig. 2.2-11) are 1-7 cm thick and show a poorly developed simple or sometimes multiple grading. The groundmass is micro-crystalline, primary volcanic minerals are up to 0,3 mm in size. In places, the tuff-beds display an internal plane- and cross-lamination and admixture of sandy siliciclastic material. Grain-size and bedding characteristics of the tuff beds suggest a distal source and only minor reworking after fallout.

XRD analysis (Fig. 2.2-12) and thin section microscopy (Fig. 2.2-13) of samples revealed quartz, albite, orthoclase, sanidine, clinopyroxene and illite-montmorillonite. Some tuff-beds contain analcime, calcite, dolomite and ankerite due to early diagenetic alteration in an alkaline-saline environment (Horsthemke et al., 1992).

Heavy mineral separates comprise primary volcanic minerals like zircon, hornblende, clinopyroxene, apatite, monazite and titanite, beside secondary iron oxides/hydroxides.

Fig. 2.2-11

Photograph of tuff-bed IV from a scarp immediately south of the Huab River in the Skeleton Coast Park area (S20°51´12´´/E13°41´16´´). The tuff-bed is dislocated by a steep, N-S striking normal fault. Vertical distance to tuff-bed V (stratigraphic equivalent of sample B18) is 120 cm (not shown).

Fig. 2.2-12

XRD Analyses of tuff-beds IV and V from different locations. Identified components are quartz, feldspar (albite, orthoclase, sanidine) and illite-montmorillonite. Some samples reveal carbonates and analcime. XRD spectrum has been interpreted together with Holzförster.

Location A: Scarp immediately south of the Huab River in the Skeleton Coast Park area.

Location B: Southern flank of an isolated hill north of the Huab River in the Skeleton Coast Park area.

Location C: Scarp north of the Huab River in the Skeleton Coast Park area (ca. 6 km west of the park border).

Location D: Eastern flank of a small incised valley in the vicinity of the Bergsig Fault system (central Huab area, south of the Huab River). Sample B 18 was collected at this location.

Fig. 2.2-13

Thin section of tuff-bed V (sample B 18). Images A, B, D, E, F and G show glass fragments. Their shape partly illustrates aerial transport. Image C shows an amphibole, clinopyroxene is shown in image H. An euhedral zircon crystal is indicated in the left side of image I (thin section preparation carried out together with Holzförster, photographs taken by Holzförster).

Fig. 2.2-14

Cathodoluminescence images of zircons used for SHRIMP analyses. Zircons of both samples (B18 and 30/98) have variable euhedral forms and they all show a well developed magmatic compositional and sector zoning. Photographs taken by Armstrong at the RSES.

Sample localities

Sample B18 was collected from a tuff-bed near the top of the Gai-As Formation in an outcrop location in the vicinity of the Bergsig Fault system south of the Huab River (S20°44´38´´/E14°08´21´´). There two conspicuously bright tuff-beds occur in approximately half a meter distance to one another. The low zircon content of the lower one (tuff-bed IV of Holzförster, 2000) was inadequate for analysis, but the upper one (tuff-bed V of Holzförster, 2000) was sufficiently zircon bearing. Sample B18 was collected from the upper tuff-bed.

Sample 30/98 derived from an outcrop north of Doros Crater (Rhino Section at S20°40´55´´/E14°11´32´´) and was stratigraphically located in the upper half of the Doros Formation (tuff-bed VI).

It should be noted, that the here used correlation differs from Holzförster (2000). The tuff-bed from which sample 30/98 derived is in Fig 2.2-7 labelled tuff-bed VI despite to Holzförster (2000), who correlates the position of sample 30/98 with tuff-bed IV of the previously described section at the Bergsig Fault system.

Results for tuff sample B18

The zircons have variable euhedral forms from acicular to more equi-dimensional and squat grains, and are generally clear, transparent and only lightly coloured. Cathodoluminescence images show well-developed magmatic compositional and sector zoning (Fig. 2.2-14).

Eighteen U-Th-Pb SHRIMP analyses were done, each on a different zircon. The zircons have high Th/U ratios (0,68 - 3,18) and variable U and Th concentrations and are plotted on a Tera-Wasserburg U-Pb concordia diagram (Fig. 2.2-15). Although most of the analyses plot as a group near the concordia, the data do show a tendency to spread along the concordia curve with a slight increase in common Pb (i.e. away from the concordia curve) suggesting some Pb-loss. Three analyses (1.1, 11.1 and 13.1) have relatively elevated common Pb concentrations (when compared to the bulk of the population), and were therefore excluded from the final calculation of the mean age. However, this does not significantly change the age, that would be calculated from all 18 analyses. From the remaining 15 analyses a final weighted mean $^{206*}Pb/^{238}U$ age of 265.5±2.2 Ma (n=15; MSWD=1.03, probability=0.42) is calculated.

Fig. 2.2-15

Tera-Wasserburg concordia diagram (uncorrected for common Pb) of sample B18 (tuff-bed V). Analyses 1.1, 11.1 and 13.1 are outliers and have been rejected from age calculation. Raw data from Armstrong, written report (1999).

Results for tuff sample 30/98

The zircons from this sample are generally 100-150 µm in length, clear, and euhedral. Most show strong compositional (magmatic) zoning when viewed with cathodoluminescence (Fig. 2.2-14). A number of rounded cores were noted, indicating the possibility of an inherited component within some grains.

The data as plotted in Figure 2.2-16 show that the 25 analyses have some appreciable scatter on a Tera-Wasserburg concordia plot (plotted uncorrected for common Pb). This indicates some heterogeneity and complexity in the age pattern for the zircons from this tuff. Nevertheless within this scatter, the majority of the analyses plot as a definite group near the concordia, and these analyses are interpreted to represent the magmatic population of this rock. The statistical treatment of the data and the final calculation of the emplacement age are $^{206*}Pb/^{238}U$ ages. Pb-loss is manifested by a tendency of the data to spread to the right in a Tera-Wasserburg plot, and clearly this is the cause of some of the scatter of data in Figure 2.2-16.

Calculation of a weighted mean $^{206*}Pb/^{238}U$ age (for all the data) using Isoplot/Excel software package gives 267.2±3.9 Ma, but even with 3 analyses rejected as outliers (7.1, 12.1 and 21.1) on Isoplotís statistical criteria, there is clearly still excess scatter in the data set as a whole (MSWD=6.1; probability=0.00). Both analyses 12.1 and 21.1 have exceptionally high common Pb contents and appear to have suffered significant Pb-loss. Analysis 7.1 appears to be too old and is therefore identified as a xenocryst. As mentioned above, some rounded cores were observed in these zircons, and this could indicate inheritance of too old ages (the cores themselves were not analysed), and it is possible that some discrete or whole grains are also xenocrystic. Both grains 7.1 and 17.1 have relatively low U (bright cathodoluminescence) and are also identified as outliers from the main group. The final weighted mean $^{206*}Pb/^{238}U$ age calculated for the remaining 15 analyses gives 272.7±1.8 Ma (MSWD=0.65; probability=0.83).

In summary, the zircons of sample 30/98 reveal more indications for Pb-loss and inheritance of too old ages compared to those of sample B18. Furthermore this sample derived from a stratigraphic higher position than sample B18 and should therefore give an at least slightly younger age than sample B18. As the opposite is the case, the 265.5 Ma age obtained from sample B18 seems to be more reliable than the 272.7 Ma age of sample 30/98. For this reason the 265.5 Ma datum will be used for further stratigraphic considerations only.

Fig. 2.2-16
Tera-Wasserburg concordia diagram (uncorrected for common Pb) of sample 30/98 (tuff-bed IV). Analysis 7.1, 12.1, 17.1 and 21.1 have been identified as outliers and are excluded from age calculation. Raw data from Armstrong, written report (1999).

2.2.3.5 Gai-As/Doros Formation Biostratigraphy

In general, the fossil content of the Gai-As and Doros formations is relatively poor and the identified species are weak tools for biostratigraphy.

Von Huene (1925) described an isolated tooth from the Doros Formation as *Archaeotherium reuningi*, but the specimen with this sample designation has been shown to be part of an amygdale (Hopson & Reif, 1981) derived from overlying basalts. The supposed presence of a dinosaur vertebra within the Gai-As to Doros Formation red beds (Keyser, 1973) is also proved to be based on misinterpretations.

Horsthemke (1992) and Ledendecker (1992) determined an Upper Permian age by using the endemic *terraia* molluscan fauna, which roughly coincides with the radiometric ages. Silicified wood found in the Upper Gai-As and Doros formations is identified as *Araucarioxylon* sp1. and *Podocarpoxylon* spp. (chapter 1.5; Bamford, 1998). The latter suggest an age comparable with the Upper Beaufort Group of the main Karoo Basin, which is Lower Triassic (Bamford, 1998).

Amphibian remains collected at the Rhino Section (see chapter 1.6 for location) have been determined by Ann Warren and Bruce Rubidge (chapter 1.5). Their stratigraphic position is even more problematic, as their bones refer more to a Triassic species despite to the teeth, that have a strong affinity to Late Permian species (Warren at al., in press): The vertebrae were determined to derive from temnospondyl amphibians. Among those the specimen probably belongs to the *Stereospondyli*, the temnospondyl clade which originated during the Late Permian and to which most Mesozoic temnospondyls belong to (Yates & Warren, in press). However, none of the described Permian members of the *Stereospondyli* have stereospondylous vertebrae, which evolved only in the Middle Triassic (Yates & Warren, in press). The structure of the stereospondylous intercentra suggest that the specimen is a capitosaurid from the Middle or Late Triassic, a conclusion supported by the structure of the exoccipital condyle. Similarly shaped intercentra have not been reported from the Permian or Early Triassic, although an intercentrum fused to pleurocentra is described in *Dvinosaurus* from the Late Permian of Russia (Bystrow, 1938).

In the teeth the simple infolding of the dentine is close to that of the Late Permian *Rhinesuchidae*, and to *Dvinosaurus* (Warren & Davey, 1992), but unlike that of the *Capitosauridae* which are complexly folded. Dark dentine has not been found in rhinesuchids. Therefore Warren et al. (in press) suggest that the material belongs to a member of the *Rhinesuchidae* with stereospondylous intercentra, or to a previously undescribed clade of *Temnospondyls*. In any case the stereospondylous intercentra of the Namibian specimen must have developed in parallel to those of Middle-Late Triassic S*tereospondyls*.

In summary, the specimen found in the Namibian Gai-As lake is unlike other described amphibian, when considering all taxonomic criteria and therefore it is not a sufficient tool for biostratigraphy.

The discrepant stratigraphic relationships of the fossil content can only be unravelled when inferring that the Upper Permian SHRIMP ages derived from tuff-bed zircon separates are correct. This would imply, that the traditional biostratigraphy of the fossil wood must be extended into the Permian and furthermore it would prove that amphibians with stereospondylous intercentra already evolved during the Upper Permian.

2.2.3.6 Correlation

The Namibian Gai-As Formation contains endemic bivalves of the *Terraia altissima* biozone only, characterised by *Leinzia similis* which confidently constrains correlation with the Serrinha Member of the Rio do Rasto Formation in the Brazilian Paraná Basin (Fig. 2-1, Fig. 2.3-1).

The 265.5 Ma U/Pb date of the Namibian sequence is very close to U/Pb zircon ages of 261 Ma (Browring, pers. comm. 1998) derived from tuffs of the *Cistecephalus* assemblage zone *sensu* Rubidge et al. (1995) of the Upper Permian Teekloof Formation (Beaufort Group) in the western Cape region of South Africa. The Teekloof Formation includes the biostratigraphic well constrained *Pristerognathus*, *Tropidostoma* and the above mentioned *Cistecephalus* assemblage zones (Rubidge et al., 1995). The fauna of these zones has a strong affinity to the dicynodont *Endothiodon*, that has been found in the Brazilian Morro Pelado Member, that follows immediately above the Serrinha Member of the Rio do

Rasto Formation (cf. Barberena et al., 1991). The Morro Pelado Member correlates most likely with the Doros Formation of the Huab area as it shows a similar shallowing and drying out trend (cf. Rohn, 1994).

2.3 DISCONFORMITIES IN THE PERMIAN SEQUENCE

The Permian sequence in the Huab-Goboboseb area developed four disconformities, which separate the above described formations from one another. Towards the basin margins the disconformities developed well defined bounding surfaces, which are fainting towards the west, viz. the Paraná Basin. All disconformities mark rapid facies changes, but only one implies a significant hiatal gap.

2.3.1 DWYKA GROUP/ VERBRANDEBERG FORMATION DISCONFORMITY

The Dwyka Group/ Verbrandeberg Formation disconformity is well notable south of Klein Gai-As (Fig. 2.3-2, Plate I-3) where clast rich diamictites of the Dwyka Group are succeeded by argillaceous shales of the Verbrandeberg Formation without any transitional unit. At other places, e.g. at section 20 (Fig. 2.2-2) coarse grained crevasse splay deposits incise sandy flagstones of the Dwyka Group. However, other places of the same area are lacking such an apparent disconformity and the base of the Verbrandeberg Formation cannot be clearly located.

The primary thickness of the Dwyka Group is unknown for the Huab area. Due to the palaeo-geographic position of the Huab area (chapter 2.1-3) it is concluded, that primary thicknesses were much smaller than those of the correlative Itararé Formation in the central Paraná Basin.

Probably isostatic rebound commenced at the late stage of the Dwyka Group. Therefore an erosional phase prior to the onset of the deposition of the Verbrandeberg Formation is rather likely. Hence a minor hiatal gap in the order of less than 10 Ma (this is approximately the minimum time span of the Dwyka Group, chapter 2.1-4) can be postulated.

2.3.2 VERBRANDEBERG/ TSARABIS FORMATION CONTACT

Most conspicuous is the contact between the Verbrandeberg and Tsarabis Formation in the area east of the Huab Outliers, e.g. at section 20 (Fig. 2.2-2). There the coarse grained sandstones of the basal Tsarabis Formation follow on argillaceous to sandy shales of the Verbrandeberg Formation with an apparent erosive contact. This contact relation is similar to that between the Dwyka Group and Verbrandeberg Formation, but there is no argument for an depositional break and subsequent erosion. Therefore the contact between the Verbrandeberg- and Tsarabis Formation probably reflects a widespread, but not very pronounced incision rather than a time-stratigraphic gap.

Fig. 2.3-1

Stratigraphic correlation panel comparing the Early Permian formations from the Eastern Cape region (South Africa) with those of the Huab (NW Namibia) and Paraná (Brazil) areas. Radiometric and biostratigraphic age constraints are indicated.

Fig. 2.3-2

Photograph of a relictic occurrence of the Dwyka Group deposits (diamictites) and the lowermost Verbrandeberg Formation (dark argillaceous shales). Location is a small incised valley at S20°50´10´´/E14°04´51´´. Compare with Plate I-3.

2.3.3 HUAB/ GAI-AS FORMATION DISCONFORMITY

2.3.3.1 The anatomy of this Permian disconformity

Field evidence for a hiatal unconformity is given by the abrupt environmental change from the marine dominated Huab Formation to the red beds of the entirely continental, predominantly lacustrine Gai-As Formation (chapter 2.2.3.3).

Although an angular unconformity is not clearly developed, phases of non-deposition and erosion are indicated by abundant pedogenic features (autobrecciation, carbonate nodules, root tubes) and erosion of stromatolitic carbonates at the top of the Huab Formation. Those pedogenic features are best developed in the eastern Huab area, whereas erosion features have also been observed as far west as the vicinity of the Bergsig Fault system. However, the boundary between the Huab and Gai-As Formation has much more the character of a para-conformity than a typical unconformity.

The W-E cross-section through the Huab area in Figure 2.3-3 shows the increasing extent of the gap when tracing the stratigraphic boundaries from close to the coast into the eastern Huab area. This section shows that the Gai-As Formation begins with successively higher units, when correlating it from the coast towards the east. As shown in Figure 2.3-3 the Gai-As Formation is lacking more than 60 m of its lowermost units when comparing the eastern parts of the Huab area (sections 25 and 26) with the area around Mt. Bruin (sections Br and 11). The amount of missing stratigraphy increases rapidly across the Bergsig Fault system, indicating its tectonic control on the basin geometry. The successive completeness of the Lower Gai-As Formation towards the west indicates a large scale, eastwardly directed onlap of the Gai-As Formation onto the Huab Formation, which supports the presence of a hiatus. However, this correlation within the Huab Basin can only reveal the considerable amount of missing strata that involves the Gai-As Formation, but it does not allow to estimate the amount of erosion of the Huab Formation.

2.3.3.2 Stratigraphic constraints

The time stratigraphic extent of the gap between the Huab and Gai-As Formation can only be estimated by large scale correlation of Permian strata in the Namibian Huab area with more complete successions in the main Karoo Basin of South Africa and the Paraná Basin of Brazil (Stollhofen et al., 2000). The *Mesosaurus* bearing Lower Permian Huab Formation serves as a marker unit as it is widespread and correlates with the Whitehill Formation of the main Karoo Basin of South Africa and the Irati Shale Formation in the Brazilian Paraná province (Oelofsen & Araujo, 1987; Ledendecker, 1992) (chapter 2.2.2.6) (Fig. 2.3-1). In the Paraná area the Irati Formation is followed by the up to 750 m thick Teresina Formation, which is succeeded by the Serrinha Member, the latter containing the stratigraphically significant *Leinzia similis* assemblage (Rohn et al., 1995).

In Namibia the Gai-As Formation, overlying directly the Huab Formation, contains the same *Leinzia similis* assemblage (chapter 2.2.3.2) as the Serrinha Member in Brazil. Such a correlation implies, that equivalents of the Teresina Formation, characterised by the *Pinzonella illusa* and *Pinzonella neotropica* assemblages are missing in Namibia.

The amount of missing strata increases even more, when correlating the Namibian Gai-As Formation with the Teekloof Formation of the South African Cape Province, as constrained by radiometric dates and vertebrate assemblage zones (chapter 2.2.3.6). Based on this correlation, the gap extends to more than 2000 m of missing strata involving equivalents of the Collingham, Ripon, Fort Brown and Waterford formations that developed in the Eastern Cape province. On a broader scale this stratal gap might correlate with a pre-Beaufort Group unconformity identified in the main Karoo Basin of South Africa by Turner (1999). However, it should be noticed that the latter could also be viewed as a diachronous transgressive bounding surface that has not the character of a typical hiatal unconformity (Hancox, pers. comm. 2000).

2.3.4 GAI-AS/ DOROS FORMATION DISCONFORMITY

The Gai-As/ Doros Formation disconformity is defined by the slightly erosively based sheetsands of the Doros Formation succeeding the Gai-As Formation. Erosion of the underlying strata is best illustrated at the south-eastern basin margins, e.g. at Mt. Brandberg (Tafel Section [Fig. 2.2-9]) where gritstone channels of the Doros Formation slightly incise into the underlying strata. Intraclasts derived from the Gai-As and Huab formations confirm the erosive character of the disconformity. Basinwards the disconformity/unconformity is less pronounced and it grades into a rather transitional contact separating the two formations.

Occasionally, e.g. at the Rhino Section (Fig. 2.2-7), the disconformity is accentuated by a low angle unconformity implying tectonic tilting of the Gai-As Formation footwall prior to the deposition of the Doros Formation (Fig. 2.4-4).

At the Rhino Section no truncation of the uppermost Gai-As Formation has been observed. Hence it is concluded, that the pre-Doros Formation topography has not been apparently affected by erosion. Consequently the Gai-As/Doros Formation disconformity reflects a new phase of block tectonics, rather than a pronounced depositional break with subsequent erosion.

Fig. 2.3-3

W-E cross-section through the Huab area showing stratigraphic gaps in the Permian and the Karoo Etendeka unconformity. The variable extent of the gaps becomes notable when tracing the stratigraphic boundaries from W to E. Correlation is based on Huab Formation maximum flooding surface equivalents. For section locations see Figure 2.1-4.

2.4 PERMIAN SYNSEDIMENTARY TECTONIC ACTIVITY

A pronounced syn-depositional tectonic activity in the Huab area is illustrated by varying sediment thicknesses and facies associations of the Permian sequence across N-S and NNW-SSE trending fault zones.

2.4.1 FACIES INDICATIONS FOR PERMIAN TECTONIC ACTIVITY

Particularly the facies distribution of the Tsarabis and Huab formations illustrates tectonic control on sedimentation: In the western domain a hemi-pelagic facies developed, represented by laminated mudstones and shales, which is confined to the east by the Uniab Fault system. Between the latter and the Bergsig Fault system stromatolitic bioherms of the Huab Formation together with nearshore sandstones of the Tsarabis Formation occur in several, a few kilometres wide, roughly NNW-SSE oriented belts. Farther east no more bioherms are developed, but palaeosols become prominent.

This distinct zonation in a westerly offshore ("hemi-pelagic"), an intermediate nearshore and an easterly pedogenically modified facies reflects clearly a tectonic segmentation of the Huab depository (Fig. 2.4-1).

The correspondence of rapid facies changes across prominent fault zones confirms tectonic control on sedimentation, since more transitional facies changes should be expected from a more continuously deepening of the depository towards the west. An internal segmentation of the nearshore facies zone is shown by the arrangement of the bioherms in at least two belts. Although the location of these belts does not clearly coincide with particular faults, their NNW-SSE trend corresponds with the prominent fault trends.

Fig. 2.4-1

Facies zonation of the Huab Formation in the Huab area. The tectonic control on facies distribution becomes obvious with the align of bioherm belts to corresponding northerly trending fault zones.

2.4.2 SYNSEDIMENTARY GEOMETRIES AND CORRESPONDING THICKNESS VARIATIONS

Another indication for the control of synsedimentary tectonic on deposition is given by pronounced thickness variations, particularly affecting the Tsarabis Formation in the Bloukrans Graben and the Gai-As Formation in the Huab area.

The Bloukrans Graben is a prominent ESE-WNW trending structure that accommodates the Tsarabis Formation with a pronounced thickness. The topography of the graben area strongly corresponds with the tectonic structures: The graben centre is traced by a valley floor that is roughly 200 m below the adjacent plateau. The graben is intersected by several NNW-SSE striking faults. The most prominent of those faults runs through the kraal area of Bloukrans 512 Farm (Fig. 2.4-2). This fault has the attitude of a steep, westwardly dipping normal fault, which causes a maximum displacement of 30 metres. Farther to the east three more NNW-SSE striking faults with a similar attitude follow. This set, comprising four NNW-SSE striking normal faults divides the WSW-ENE trending graben into a western and an eastern segment: The main graben axis of the western segment runs 3 km north of the graben axis of the eastern segment.

Syn-tectonic deposition is viewed by thickness variations across the graben structure and across internal faults within the graben (Fig. 2.4-2). Section A-B of Figure 2.4-2 illustrates a S-N cross-section of the western graben segment. The maximum thickness of the Tsarabis Formation is 60 m and which is attained in the vicinity of a synthetic normal fault, located just east of the road. 1 km farther north, at the cliff to the adjacent plateau, the Tsarabis Formation is only 40 m thick. The total fault induced offset between these two locations is 175 m.

Another example for fault related thickness variations provides a S-N cross-section through the eastern graben segment. In section E-F of Figure 2.4-2 the Tsarabis Formation measures 85 m in thickness in the vicinity of the Petrified Forest National Monument. Crossing three steep, southwardly dipping normal faults, the thickness decreases to 45 m, measured at a cliff 1.6 km farther north. The total vertical offset along this section is 165 m.

The described two S-N sections clearly document, that the Tsarabis Formation thickens from the graben shoulder toward the graben axis, indicating the tectonic control on accommodation space during the Early Permian. Similar thickness variation shows the Huab Formation in section C-D (Fig. 2.4-2), but due to the lack of complete thickness sections this is not very well constrained.

The measured maximum thickness variation is roughly 50 m in S-N sections, which is less than the total vertical displacement of 165 m, measured in section E-F. Therefore the mayor phase of graben subsidence occurred after the deposition of the preserved Karoo strata.

At a smaller scale, the Huab Formation provides a good example for synsedimentary tectonic activity: A stromatolitic body overgrows an E-W trending horst and graben structure that conjugates the N-S trending Bergsig Fault (location at S20°45´43´´/E14°09´03´´). Figure 2.4-3 and Plate V-4 show a pre-kinematically consolidated footwall that consists of horizontally layered marl and limestones. Several steep, north- and southwardly inclined normal faults cut through this section and cause vertical displacements of up to 80 cm. As shown in the detailed view, the pre-kinematic part is succeeded by stromatolitic laminae, which form sigmoidal flexures corresponding to the fault kinematics. These geometries are attributed to a syn-kinematic flexural bending of unconsolidated stromatolite laminae. Up-section the fault induced topography is draped by late-kinematic and post-kinematic onlap of further stromatolite laminae.

The Rhino Section north of Doros Crater, located approximately 6 km east of the Bergsig Fault system (Fig. 2.4-4) provides another outcrop which documents Permian tectonic activity. This section illustrates tectonically induced onlap geometries: The WSW-ENE cross-section of Figure 2.4-4 shows a westerly inclined Gai-As Formation footwall, which is confined to the east by a N-S striking Etendeka Group basaltic Dyke. The dips of the Gai-As Formation layers are almost uniformly 3° to the west. Onlapping sandsheets of the Doros Formation developed wedge shaped geometries by draping the Gai-As tilt-block topography. Up-section the primary footwall-dip of 3° successively decreases until almost horizontal, which indicates that the main tilting phase occurred prior to the deposition of the Doros Formation. Thus, the corresponding tilt-block confining fault has been synsedimentary active during the Permian. Later, it has been intruded by the basaltic dyke, which is now the delimiting structure of the tilt-block.
On a basin wide scale, the pronounced westwards directed thickness increase of the Gai-As Formation from roughly 40 m (without the Doros Formation) east of the Bergsig Fault system (e.g. at the Rhino Section DP14, Fig. 2.2-7) to more than 100 m at Mt. Bruin indicates the influence of synsedimentary faults (e.g. the Bergsig and Wêreldsend Fault systems) on the basin geometry.

Fig. 2.4-2
Tectonic interpretation of aerial photographs of the Bloukrans Graben in the eastern Huab area (compiled together with Holzförster). The Bloukrans Graben is a prominent E-W trending structure intersected by several N-S trending fault systems. Sections across the graben reveal pronounced thickness variations of the Lower Permian Tsarabis Formation, indicating syn-depositional tectonic subsidence of the graben. See Figure 2.1-4 for location.

Fig. 2.4-3
Detailed section of a synsedimentary, E-W trending horst structure in the Lower Permian Huab Formation. The thickness variations and flexural deformation indicate fault activity during stromatolite growth and sedimentation of micritic, partly conglomeratic limestone. The outcrop is located at the eastern flank of a small incised valley in the vicinity of the Bergsig Fault system (S20°45´43´´/E14°09´03´´). Compare with Plate V-4.

Fig. 2.4-4
Detail of onlap geometries within the Gai-As and Doros formations in the vicinity of a N-S trending dyke (location at S20°40´39´´/E14°11´20´´, Fig. 5.3-11). The dyke comprises Etendeka Group basalt that intruded a Permian normal fault. Synsedimentary geometries are indicated by the successive decrease of the dip-angle of the Doros Formation when tracing individual horizons up the section. Compare with Plate VI-2.

2.5 PERMIAN TECTONIC AND CLIMATIC EVOLUTION OF THE HUAB-GOBOBOSEB AREA

The Permian succession of the Huab area in north-western Namibia records both, climatic and tectonic signatures. After repeated sea level highstands during deposition of the Late Carboniferous Dwyka Group (Visser, 1997), isostatic rebound after deglaciation resulted in the establishment of cool terrestrial conditions of the Verbrandeberg Formation (Horsthemke et al., 1990). Subsequent transgression and melioration of the climate is expressed by the landward-stepping marine facies of the Tsarabis and Huab formations. The resulting retrogradational stacking pattern is mirrored by contemporaneous, but much thicker Early Permian successions in South Namibia (Grill, 1997), South Africa (Visser, 1993) and South America (Rohn, 1994) and it is presumably related to a third-order rise in relative sea level (cf. Vail et al., 1991).

Fault related thickness and facies variations are related to a significant E-W extension, accommodated by a number of NNW-SSE trending faults. These fault trends do not coincide with the SW-NE trend of the Huab Basin axis defined by the Huab Fault (Fig. 2.1-4; chapter 1.2-4), but they parallel the axis of the early southern South Atlantic rift (Stollhofen et al., 2000). In the south-western and eastern Huab area SW-NE and E-W trending faults (e.g. the Bloukrans Graben, fault sets north-east of the Bergsig Fault and south of the Uniab Fault; Fig. 2.1-4) are also structural relevant. Some of them might relate to inherited Pan-African structures. In this context, the Bloukrans-Graben represents an important, roughly E-W trending structure that influenced significantly the Lower Permian sedimentary sequence.
A considerable amount of crustal stretching affecting the area is manifested by the volcanics of the Verbrandeberg Formation, which presumably erupted along extensional faults.

An early uplift of the eastern hinterland might be indicated by coarse, westwardly prograding deltaic deposits of the upper Huab Formation in the eastern Huab area. A westwards propagation of this uplift lowered the base level of deposition at the basin margins, that ultimately resulted in a depositional break. This depositional break and possibly associated erosion are documented by the para-conformity, which separates the marine facies of the Huab Formation from the lacustrine Gai-As Formation. Furthermore, the implied palaeo-topography is confirmed by the successive onlap of Lower Gai-As Formation stratigraphy towards the basin margin in the east (Fig. 2.3-3).
On a regional scale this local tectonic event might be related to an orogenic pulse in the Cape Fold Belt, dated at 258±2 Ma (Hälbich et al., 1983).
In the more central part of the Paraná-Huab rift depression and in the Karoo foreland basin the depositional sequence is fully developed and records a marine-nonmarine transition (Teresina Formation and equivalents), which is missing in the Huab area. Porada et al. (1996) relate the establishment of the huge Gai-As freshwater lake (Gai-As Formation and equivalents) to the cut-off from the marine realm

towards the south, which was probably caused by large-scale uplift of the Argentinean Puna Highlands during the Cape-Ventana and San Rafael orogenies (cf. Veevers et al., 1994b).

The Gai-As lake deposits express a final phase of aggradational overall basin infill, as indicated by their overall upward-shallowing trend. Playa deposits record a warm, semi-arid climate. During aggradation the onset of the Doros Formation announces a new phase of block tectonics, which is best shown at places where a slightly angular unconformity developed (Rhino Section, Fig. 2.4-4). A fan delta formed at the south-eastern basin margin (Doros Crater, Mt. Brandberg), which was probably related to initial uplift of the SW-NE trending tectonostratigraphic "Northern Central Zone" (Miller, 1983) of the Damara orogen (Fig. 3.4-1).

In summary, widespread extension prevailed during the Permian with fault trends corresponding to the major Permian rift depressions (Fig. 1.2-5) and Pan-African basement anisotropies (Fig. 1.2-2). Block tilting intensified during the deposition of the Doros Formation, but less pronounced at this stage than during the Cretaceous megasequence (Etendeka Group; chapter 5.3).

Chapter 3: The Triassic-Jurassic Megasequence and associated Unconformities

In north-western Namibia the Triassic-Jurassic megasequence accumulated in the SW-NE trending elongated Waterberg-Erongo depository (Fig. 3.4-1) (Reuning & Von Huene, 1925; Gevers, 1936; Porada et al., 1996) which is northwardly confined by the Waterberg-Omaruru Fault zone (Holzförster et al., 1999). Farther north, the sequence comprises only Triassic strata, which are reported from the Mt. Brandberg, Goboboseb and Otjongundu areas. According to Porada et al. (1996) the corresponding depocentre has been a south-westerly trending halfgraben (Goboboseb-Otjongundu Basin; Fig. 3.4-1).

In all these depositories the Triassic is represented by the entirely continental red-bed sequence of the fluvial Omingonde Formation (and equivalent strata), whereas the Jurassic is represented by the fluvio-aeolian desert deposits of the Etjo Formation (and equivalent strata) (Fig. 3.1-1).

<u>3.1 STRATIGRAPHY AND FACIES OF THE OMINGONDE FORMATION</u>

3.1.1 INTRODUCTION

The Omingonde Formation splits naturally into four units, each comprising several upward-fining cycles reflecting distinct architectural characteristics (Holzförster et al., 1999) (Fig. 3.1-1). The lower two units coincide with the Lower and Middle Omingonde Formation, the upper two units make up the Upper Omingonde Formation. Thicknesses and architectural style of the units vary laterally and are influenced by synsedimentary tectonic activity of the Waterberg-Omaruru Lineament. The full range of facies is best developed in the Mt. Waterberg area, whereas farther west in the Erongo region more proximal facies are favoured. This proximal facies association has been termed the Krantzberg Formation in the western, northern and eastern Erongo area, and interfingers with the Lions Head Formation towards the south-eastern Erongo region (Hegenberger, 1988).

North of the Ameib Line, which is part of the Waterberg-Omaruru Fault zone, no rocks of Karoo age are preserved until the Otjongundu Basin, which hosts a 350 m thick conglomeratic red-bed sequence that probably correlates with the Omingonde Formation (Miller, 1980). The Otjongundu Basin itself is confined by the Otjohorongo Thrust to the south. This thrust extends and splits to the south-west into two slightly diverging faults systems, of which the southern one is termed the Autseib Fault (cf. Milner, 1997). Between Mt. Brandberg and Messum Crater the Goboboseb Basin is located, which forms probably the south-western extension of the Otjongundu Basin. There an up to 120 m thick succession of immature gravely arkose was deposited, which presumably correlates with the Middle Omingonde Formation.

3.1.2 LITHOSTRATIGRAPHY AND FACIES

The Omingonde Formation comprises red and white coloured clastics including conglomerates, sandstones and siltstones. Many units are channelised and show palaeosol development.

Fig. 3.1-1

Overview section of the Waterberg-Erongo lithostratigraphy (modified from Holzförster et al., 1999). Stratigraphic positions of vertebrate fossil findings and detailed measured sections (Figures 3.1-2, 3.1-3, 3.1-6, 3.2-1, 3.2-3 and 3.2-3) are indicated. Legend

3.1.2.1 LOWER OMINGONDE FORMATION

The Lower Omingonde Formation (unit 1 *sensu* Holzförster et al., 1999) attains 250 m cumulative thickness in the south-eastern Erongo area and pinches out towards the north-east and south-east of the Mt. Waterberg area.

Three major facies are developed, including channelised, trough cross-bedded conglomerates, followed by sandy mudstones, which are interbedded with more massive, pebbly mudstones (Fig. 3.1-2).

The conglomerates are matrix-supported and grade vertically and laterally into sandstones. Their clasts derived dominantly from metasedimentary and granitic rocks. The channels are usually well confined and amalgamation occurs rarely. The sandy mudstones are reddish in colour, they contain lenticular interbeds of fine grained sandstones. They are characterised by a climbing-ripple lamination. Furthermore, the mudstones contain thin interlayers of blocky calcrete and carbonate nodules with some rhizoliths.

Massive pebbly mudstones are restricted to positions very proximal to the Waterberg-Omaruru Fault. They are characterised by a massive matrix-supported texture and a low structural and compositional maturity of both, matrix and clasts.

Fig. 3.1-2

Detailed section of the Lower Omingonde Formation at Mt. Etjo (modified from Holzförster et al., 1999). Legend

3.1.2.2 MIDDLE OMINGONDE FORMATION

The Middle Omingonde Formation (unit 2 *sensu* Holzförster et al., 1999) occurs in both basins, the Waterberg-Erongo and the Goboboseb-Otjongundu Basin (Fig. 3.4-1). The two depositories are characterised by a cyclic alternating facies association of amalgamated conglomeratic channel deposits and a reddish concretionary mudstone facies. However, the lithologies and architecture of the deposits in both basins exhibit distinctive differences.

<u>Waterberg-Erongo Basin</u>

The channel facies of the Mt. Waterberg area is made up of sheet-like conglomerate bodies, which are partly embedded in muddy deposits (Fig. 3.1-3). Individual channels are 100-350 m wide and display vertically and laterally amalgamation and pronounced trough cross-bedding. The channel fill conglomerates comprise sub to well rounded pebbles, cobbles and boulders of vein quartz, granitoids, as well as flattened mud clasts, which are interpreted as rip-up clasts derived from finer grained intercalations.

The channel facies is usually overlain by reddish-purple mudstones, that contain fine grained sandstone units, some 10-50 cm thick with erosively scoured bases. The facies contains about 30% calcareous detritic material and is characterised by abundant carbonate concretions. A spherical type of concretions, a few cm in diameter, can be distinguished from a vertically orientated elongated type, which is up to half a meter long (Holzförster et al., 1999). These structures are typical for pedogenic modification of fine grained sediments (cf. Freytet & Plaziat, 1982).

The alternating conglomerate - mudstone units are cyclically arranged. In the Mt. Waterberg area Holzförster et al. (1999) identified 16 individual depositional cycles, each 4-17 m thick.

The proximal facies in the Erongo area reveals channelised matrix-supported breccia and massive clast bearing mudstones. Channel bases are commonly non-erosive, with individual units up to 3.5 m thick. The channel fills conglomerates reveal a wide spectrum of clast sizes ranging from pebbles to angular boulders of up to 1 m in diameter. This ultra-proximal facies exhibits pronounced thickness variations from a few metres north of the Waterberg-Omaruru Fault zone to several hundred metres immediately south of it.

<u>Goboboseb-Otjongundu Basin</u>

The Omingonde Formation attains 350 m thickness in the Otjongundu Plateau and 120 m in the Brandberg/Goboboseb area. There it is preserved at Mt. Brandberg (Tafel Section, Fig. 3.1-4) and in the south-eastern Goboboseb Mts. only. The succession is characterised by thick units of amalgamated, white gritstones and conglomeratic channel fillings, which grade into purple, fine grained sandstones and siltstones, arranged in individual upward-fining units each up to 18 m thick. Usually the channels developed erosively scoured basal surfaces with the conglomerates forming channel lag deposits, which mainly consist of well rounded quartz pebbles and cobbles, mica and less abundant reworked intraclasts. The conglomerates grade into trough cross-bedded quartz-pebble gritstones, current-ripple cross-laminated, white sandstones and finally into thinly bedded, mica-rich, purple fine sandstones and siltstones. Occasionally they contain clay nodules and small rhizoliths indicating pedogenic modification. Palaeo-currents, deduced from trough cross-beds, reflect transport directions towards the NW, with a second mode towards the NE (Fig. 3.1-5).

Fig. 3.1-3
Detailed section of the Middle Omingonde Formation at Mt. Große Waterberg (modified from Holzförster et al., 1999).
Legend

Fig. 3.1-4
Detailed section of the Middle Omingonde Formation at the south-western slope of Mt. Brandberg (compiled together with Holzförster). The unconformity to the overlying Etendeka lavas is marked by a thin sandsheet of the aeolian Twyfelfontein Formation. Legend

Fig. 3.1-5
Palaeo-current directions of the Middle Omingonde Formation at Mt. Brandberg (Tafel Section, Fig. 3.1-4). The two prominent modes are NW and NE directed palaeo-current vectors.

3.1.2.3 UPPER OMINGONDE FORMATION

The Upper Omingonde Formation is best developed in the Mt. Waterberg area, but no occurrences have been found in the Goboboseb-Otjongundu region. The succession begins with a laterally amalgamated channel facies alternating with a sandy mudstone facies. In the lower units (unit 3 *sensu* Holzförster et al., 1999) the channel bodies are 80-150 m wide and the degree of lateral amalgamation is similar to the Middle Omingonde Formation, but vertical stacking is less pronounced and decreases systematically upwards. In the upper part of the formation (unit 4 *sensu* Holzförster et al., 1999) (Fig. 3.1-6) channels are isolated and only a few metres wide. Up-section, the maturity of channel fills successively increases until well rounded medium to coarse grained sandstones dominate. The architecture remains cyclic with the channel facies alternating with laminated fine grained sandstones and bioturbated mudstones.

Fig. 3.1-6

Detailed section of the Upper Omingonde Formation at Mt. Große Waterberg (compiled together with Holzförster). Legend

3.1.3 FACIES DEVELOPMENT

The Omingonde Formation comprises chiefly braidplain deposits, which record semi-arid climatic conditions prevailing until the Upper Omingonde Formation. The latter marks a temporary shift to a wetter climate until semi-arid conditions re-established with the succeeding Etjo Formation. According to Rust (1978) and Friend et al. (1979) the degree of channel amalgamation and palaeosol development corresponds with sediment discharge rates: More confined meandering channels and extensive palaeosol developments are associated with lower discharge rates rather than amalgamated braided channel systems.

The channel geometries and the low degree of amalgamation in the Lower Omingonde Formation indicate an overall meandering river system. Sandy mudstones were generated during crevasse splays and associated flood stages. Intervals of non-deposition are indicated by immature palaeosols, which are represented by scarce rhizoliths and blocky calcretes. In proximal positions to the Waterberg-Omaruru-Fault pebbly mudstone units reveal mud flow deposits related to an enhanced fault generated relief.

Holzförster et al. (1999) interpret the Middle Omingonde Formation as a mobile braided stream system due to the well developed trough cross-bedding and upward-fining character of poorly confined channel bodies, together with their sheet-like appearance and ubiquitous vertical and lateral amalgamation. In addition, significant vertical accretion is indicated by the preservation of overbank fines, that abundantly show pedogenic modification. The proximal deposits of the Erongo area show textural and grain size characteristics of debris flows that amalgamated to form fault-bounded alluvial fan aprons (Holzförster et al., 1999). These fans give evidence for persistent synsedimentary activity of the Waterberg Fault system.

The Upper Omingonde Formation reflects a progressive change from a braided river system to a more meandering river system with decreasing discharge rates. Temporarily wetter climates are indicated by bioturbated mudstones, that formed in seasonal lakes. Although a hiatus to the overlying Etjo Formation is proposed (chapter 3.3.2.2), the facies of the Upper Omingonde Formation represents a transitional stage to the facies of the Lower Etjo Formation.

3.1.4 BIOSTRATIGRAPHIC CONSTRAINTS AND CORRELATION

Several biostratigraphic relevant vertebrate fossils have been reported from the Omingonde Formation of the Waterberg-Erongo Basin, but such stratigraphic fossil constraints are lacking for the Goboboseb-Otjongundu Basin. However, due to striking lithological similarities Miller (1980) and Porada et al. (1996) assign the coarse clastic deposits of the Goboboseb-Otjongundu area strictly to the Omingonde Formation. Furthermore, these deposits can be more intimately attributed to the Middle Omingonde Formation, because of the similar architecture of amalgamated channel facies, arranged in upward-fining cycles (as described in the previous chapter).

Pickford (1995) describes a partial skeleton of a proterosuchian cf. *Erytrosuchus* in the Lower Omingonde Formation near Mt. Etjo. This fossil establishes a correlation with the Burgersdorp Formation (Beaufort Group) of the Eastern Cape region (Fig. 3.3-1), which is of an Olenekian/Anisian age (Kitching, 1995).

The Middle and Upper Omingonde Formation contain a variety of vertebrate fossil fragments, which have been collected in the behalf of this study on the south-western slope of Mt. Klein Waterberg. The fragments most probably belong to *Cynognathus* sp., *Kannemeyeria cristarhynchus*, *Dolichuranus* sp. and an erytrosuchid archosaur (Rubidge, pers. comm. 1999). This assemblage has a strong affinity to the Cynognathus Assemblage Zone in the main Karoo Basin, which is assigned to an Olenekian/Anisian age (Kitching, 1995). More vertebrate fossils that are related to the Cynognathus Assemblage Zone are reported by Keyser (1973, 1978), who describes a moderately diverse fauna from Mt. Etjo including eripoid amphibian and several reptiles including *Proterosuchia*, *Dicynodontia*, *Bauriamorpha* and *Cynodontia*.

In summary, the Omingonde Formation reveals exclusively vertebrate fossils which refer to an Early Triassic to early Middle Triassic age, implying a relatively short time span for the up to 500 m thick sequence.

3.2 STRATIGRAPHY AND FACIES OF THE ETJO FORMATION

3.2.1 INTRODUCTION

The Lower Jurassic Etjo Formation follows with an erosional unconformity above the previously described Omingonde Formation. It is the uppermost preserved unit of Karoo sediments in northern Namibia and forms usually the top-parts of the exposures. Only north-east of the Mt. Waterberg area the Etjo Formation is overlain by the probably Jurassic Rundu Formation (basalts) and the Cenozoic Kalahari Group (sediments).

The Etjo Formation breaks into three genetically related units displaying a transition from semi-arid to fully arid climatic conditions (Fig. 3.1-1). The facies spectrum of the three units is best developed immediately south of the Waterberg-Omaruru Fault trace, where the entire sequence attains its maximum thickness of at least 140 m. Particularly the lower two units vary considerably in thickness with a general decrease towards the south and the north of the Waterberg-Omaruru Fault. Good outcrops provide the cliffs in the Waterberg Plateau Park and Mt. Etjo area.

3.2.2 LITHOSTRATIGRAPHY

3.2.2.1 LOWER UNIT

Immature pebbly sandstones are cyclically interbedded with laminated to massive, well sorted sandstones (Fig. 3.2-1). In the vicinity of the Waterberg Plateau the cycles are approximately 1 m thick and the entire unit measures 25 m. The pebbles are commonly concentrated in layers which results in a faint internal bedding of the pebbly sandstone units. The clast spectrum embraces granitoids and fragments of pedogenic carbonate nodules and silcretes originating from the underlying Omingonde Formation. The interlayered laminated sandstones are fine grained and isolated interleaved channel bodies occur. They contain thin mudstone layers which exhibit an ichnofauna assemblage including *Planolites* isp. and *Palaeophycus* isp. and often mud curls and desiccation cracks. Minor palaeosols are indicated by rhizocretions.

Fig. 3.2-1
Detailed section of the Lower Unit of the Etjo Formation at Mt. Große Waterberg (compiled together with Holzförster). Legend

3.2.2.2 MIDDLE UNIT

The up to 15 m thick Middle Unit forms the prominent reddish cliff beneath the Waterberg Plateau summit. Homogeneous, well sorted sandstones with a few laminated mudstone interlayers make up the sequence (Fig. 3.2-2). Sandstone composition, grain and sorting characteristics are similar to those of the overlying aeolian unit. The only pronounced difference is the occurrence of reworked intraformational mud clasts and the scarcity of internal structures.

Trough cross-bedding and normal graded bedding has been observed rarely. Trace fossils and rhicocretions occur sporadically and are much fewer than in the Lower Unit.

Fig. 3.2-2
Detailed section of the Middle Unit of the Etjo Formation at the plateau of Mt. Große Waterberg (slightly modified from Holzförster et al., 1999). Legend

3.2.2.3 UPPER UNIT

The mainly aeolian Upper Unit (Fig. 3.2-3) achieves a maximum exposed thickness of at least 100 m in the northern Mt. Waterberg area, but a well located 125 km north-east of the Waterberg Plateau reveals thicknesses exceeding 300 m under Kalahari sediment cover (Osborne 1985). The top of numerous mountains, e.g. Mt. Große Waterberg, Mt. Etjo, the Omatako Mts. and Mt. Gamsberg is marked by this unit, documenting its widespread occurrence. Almost the entire succession is made up of homogeneous sandstones consisting of well sorted and rounded medium sized quartz grains. Tabular large scale cross-beds display reverse grading and thin laminated layers. Dip-orientations of the cross-beds reveal palaeo-winds from southerly and westerly directions (Fig. 3.2-3). Sporadically the cross-beds are truncated by thin mudstone or coarse to gritty sandstone layers. The thickness of those layers is roughly proportional to the mean grain size. In several of those horizons dinosaur trackways are preserved. They are good exposed at several localities, e.g. on the Waterberg Plateau and at Dinos Farm (a portion of Otjihaenamaperero 92 Farm).

Fig. 3.2-3
Detailed section of the Upper Unit of the Etjo Formation at Mt. Große Waterberg (slightly modified from Holzförster et al., 1999). Dip directions of dune foresets reveal palaeo-winds from southerly and westerly directions. Legend

3.2.3 FACIES DEVELOPMENT

Stanistreet & Stollhofen (1999) compare the Lower and Middle Etjo Formation with the "semi-wet" desert environment of the modern Kalahari. Facies architecture and bedding features of the Lower Unit suggest that the abundant sandstones have been deposited by shallow, rapid flow of sheet floods. Interbedded sandstones are interpreted as aeolian sand sheets deposits by Holzförster et al. (1999). However, thin mud layers record intermittent periods of aquatic deposition. They either deposited during waning stages of sheet floods or in shallow temporary lakes.

The Middle Unit represents a prograding desertification to the fully arid environment of the Upper Unit. Mudstone layers accumulated in interdune areas, which provided the mudstone intraclasts found in the coarse grained horizons.

The Upper Unit represents dominantly dune deposits, which reveal palaeo-winds from south-westerly directions. The single grain layers, truncating aeolian cross-beds, represent palaeo-deflation surfaces in interdune areas, which often preserve dinosaur tracks. Imprinting of the tracks occurred in a wet substrate, as indicated by track morphology and soft sediment deformation (Löffler & Porada, 1998).

3.2.4 BIOSTRATIGRAPHIC CONSTRAINTS AND CORRELATION

Traditionally the Etjo Sandstone Formation of the Mt. Waterberg/Mt. Etjo region has been correlated with the aeolianites of the Huab area, due to their comparable lithological characteristics (e.g. SACS, 1980; Miller & Schalk, 1980). However, the latter interfinger with lavas of the Lower Cretaceous Etendeka Group (chapters 4.1.9.2 & 4.1.10) which is inconsistent with the biostratigraphically based correlation of the Etjo Formation from its type localities at Mt. Waterberg and Mt. Etjo with the Jurassic Clarens Formation of South Africa (Geevers, 1936). Due to this problem Martin (1982) suggested that the Etjo Formation in the Mt. Waterberg/Mt. Etjo area might be intermediate in age between the Clarens Formation of the main Karoo Basin and the aeolian units of the Huab area, implying a pronounced diachroneity of the sub-Etjo surface.

New vertebrate finds confirmed a Jurassic age for the Etjo Formation: In the Middle Unit a well preserved mould of a largely articulated skeleton of a prosauropod dinosaur, identified as *Massospondylus* sp. (Holzförster et al., 1999) was found just below the Waterberg Plateau summit.

Within the Upper Unit dinosaur trackways on palaeo-deflation surfaces have been found at numerous localities (Gürich 1926a, 1926b; Von Huene, 1925; Wiechmann, 1938), but not all of them are biostratigraphically classified yet. Small saurischian tracks found on Dinos Farm (a portion of Otjihaenamaperero 92 Farm) probably relate to the theropod dinosaur *Syntarsus* sp. (cf. Raath, 1980), which is known from the Jurassic Forest Sandstone of Zimbabwe. Kitching & Raath (1984) place the latter in a stratigraphic level transitional to the Middle and Upper Elliot Formation in the main Karoo Basin. This overlaps with the occurrence of *Massospondylus* sp., which is confined to the Middle and Upper Elliot Formation (Upper Triassic) and the Lower Jurassic Lower Clarens Formation (Kitching & Raath, 1984). Other dinosaur trackways found at Dinos Farm and on the Waterberg Plateau have been compared by Löffler & Porada (1998) and Stanistreet & Stollhofen (1999) with the South African uppermost Triassic/Lower Jurassic Stormberg footprint assemblage of Ellenberger (1970) including *Quemetrisauropus princeps* and *Prototrisauropus crassidigitus*.

According to the find of *Massospondylus* sp. and the ichnostratigraphic constraints the maximum age for the Etjo Formation is Upper Triassic. However, Holzförster et al. (1999) conclude a Lower Jurassic age for the Etjo Formation, which implies its correlation with the Hettangian to Sinemurian Clarens Formation (Olsen & Galton, 1984) of the main Karoo Basin (Fig. 3.3-1).

This Jurassic age is favoured because of a proposed hiatus to the underlying Upper Triassic Omingonde Formation (chapter 3.3.2.2).

A Jurassic minimum age of the Upper Etjo Formation is indicated by the Rundu basalts which overly the north-easterly inclined Etjo beds subsurface north-east of the Waterberg Plateau (Martin, 1961b). The

Rundu basalts have been traced into the south-eastern Ovambo Basin of northern Namibia (Miller, 1997) and they most likely correlate with the southern Namibian Kalkrand flood basalts (Fig. 1.2-4) that are time equivalent to the Jurassic Drakensberg basalts of Lesotho (Fig. 1-2) being dated at 186-183 Ma (Duncan et al., 1997).

Holzförster et al. (1999) introduced the threefold subdivision of the Etjo Formation as described in chapter 3.2.2. Their subdivision into (1) a fluvial dominated lower unit, (2) a fluvio-aeolian middle unit and (3) an aeolian dominated top unit is fully developed in the Waterberg National Park region. Etjo equivalents in South Africa (cf. Eriksson, 1979; Kitching & Raath, 1984) show a comparable architecture implying that this organisation of Etjo subunits is not a local, but a regional development.

3.3 DISCONFORMITIES IN THE TRIASSIC/JURASSIC SEQUENCE

The Triassic Omingonde Formation is confined by disconformities at its base and top. The proposed magnitudes of stratal gaps vary between the Goboboseb-Otjongundu and the Waterberg-Erongo Basin. They are deduced either from tetrapod assemblage zones or by correlation with stratigraphically classified strata.

Fig. 3.3-1

Triassic and Jurassic vertebrate assemblage zones of the main Karoo Basin in South Africa (after Rubidge et al., 1995 and Kitching & Raath, 1984) correlated with the Omingonde and Etjo formations in the Waterberg-Erongo Basin.

3.3.1 THE PERMIAN/TRIASSIC DISCONFORMITY

3.3.1.1 GOBOBOSEB-OTJONGUNDU BASIN

The Tafel Section at the south-western slope of Mt. Brandberg exposes the Permian Gai-As Formation, succeeded by a coarse grained maroon gritstone unit (Fig. 2.2-9) followed by white quartz-pebble gritstones and conglomerates (Fig. 3.1-4). The sedimentary succession is topped by the Lower Cretaceous Etendeka Group. According to Porada at al. (1996) the bounding surface between the two gritstone/conglomerate units is defined by an erosive unconformity (unconformity between Sedimentary Unit III and IV). Its erosive character is expressed by rip-up clasts of the lower maroon gritstone unit occurring in the lowermost white conglomerates of the upper unit. At places white conglomeratic channel bodies incise erosively into the underlying maroon gritstones. However, no apparent angular unconformity is developed and at most places the contact between the two units is more concordant than discordant.

Proposing a hiatus between the two gritstone units is only possible after dating both units. However, stratigraphic relevant fossils have not been found and hence ages are constrained by lithostratigraphic correlation with comparable deposits in the Huab area and the Waterberg-Erongo Basin, respectively. It should be noted at this place, that direct correlation is impossible due to the lack of outcrops connecting either basins.

The maroon gritstone unit correlates well with the sandsheet deposits of the Doros Formation in the Huab area (Rhino Section, Fig. 2.2-7), which reveals an Upper Permian radiometric age (265.5±2.2 Ma; chapter 2.2.3.4). The white sandstones and conglomerates are comparable with the Middle Omingonde Formation, which are most likely of an Olenekian/Anisian age (chapter 3.1.4). After these correlations a stratal gap of at least 15 Ma is inferred.

3.3.1.2 WATERBERG-ERONGO BASIN

The Waterberg-Erongo Basin hosts the Permian Tevrede Formation succeeded by the Early Triassic Lower Omingonde Formation (Fig. 3.1-1). Since the disconformity between the two sequences is only known from poor borehole data it cannot be characterised in detail. But, the pedogenically modified top of the Tevrede Formation seems to be a well constrained feature, that characterises the disconformity. Holzförster et al. (1999) and Stollhofen (1999) propose a hiatus between the two formations in the order of 40 Ma. The assumption of such a stratal gap implies that the minimum age of the Tevrede Formation is Lower Permian. Although there is no biostratigraphic evidence for such an age, correlation based on lithostratigraphy and facies associations justifies such a Lower Permian age: The upper part of the up to 40 m thick Tevrede Formation comprises glauconite bearing mudstones and coal seams that point to a shallow marine, lagoonal environment, which contained large amounts of plant material on swampy coastal plains (Gunthorpe, 1987). This facies corresponds to the glauconitic and coal bearing Lower Permian Ecca Group in southern Namibia (Kingsley, 1985) and South Africa (Smith et al., 1993), as well as to the Verbrandeberg and Tsarabis formations of the Huab-Goboboseb region.

Despite of this correlation, Porada et al. (1996) presume that the shales and mudstones (Upper Tevrede Formation) in the Waterberg-Erongo Basin (Otjiwarongo Basin *sensu* Martin, 1982) partly correlate with the uppermost Gai-As Formation (Upper Permian), which would imply a much smaller time-stratigraphic gap. But, this correlation seems to be rather unlikely as the Upper Permian Gai-As Formation is entirely continental, whereas the glauconitic horizons of the Tevrede Formation indicate marine influence.

An early Triassic age for the overlying Lower Omingonde Formation is supported by the find of a proterosuchian cf. *Erytrosuchus* (Pickford, 1995).

3.3.2 THE TRIASSIC/JURASSIC LOWER CRETACEOUS DISCONFORMITY

The Triassic Omingonde Formation is succeeded either by the Etendeka Group (Goboboseb Basin) or by the Lower Jurassic Etjo Sandstone Formation (Waterberg-Erongo Basin). The anatomy of the corresponding disconformities shows pronounced differences due to the tectonic environment of the depositories.

3.3.2.1 GOBOBOSEB-OTJONGUNDU BASIN

At the south-western slope of Mt. Brandberg (Tafel Section, Fig. 3.1-4) and in the south-eastern Goboboseb Mts. conglomeratic Middle Omingonde Formation equivalents (Early to Middle Triassic; chapter 3.1.4) are disconformably capped by basaltic lavas of the Etendeka Group (Lower Cretaceous; chapter 4.2.6). The contact to the basal lava is plane and rather distinct. In places sedimentary deformation structures, similar to load casts were observed. Furthermore, a SW-NE trending striation, that corresponds to the final lava movement, occurs on the top-surface of the sedimentary succession. These structures point to soft sediment deformation caused by the emplacement of the lavas. However, it appears unlikely that presumably Triassic sediments remain unlithified until the onset of flood volcanism in the Lower Cretaceous, nevertheless this possibility can not be excluded. It is therefore proposed, that either the top of the sedimentary succession is part of the Etendeka Group (equivalent to the Twyfelfontein Formation; chapter 4.1), or it had been temporarily mellowed by weathering-off the cement-matrix during a phase of exposition. The first possibility is confirmed, where almost plane, a few cm thick aeolian sandsheets appear at the disconformity (Fig. 3.1-4). These thin beds most likely belong to the Lower Cretaceous Twyfelfontein aeolianites (chapters 4.1.8 & 4.1.9) that mainly bypassed the formerly exposed Omingonde surface contemporaneous to flood basalt volcanism. The situation becomes more problematic, where the lava caused soft sediment deformation in conglomerates. They could either belong to the conglomeratic Krone Member of the Twyfelfontein Formation or to the lithological essentially similar, but formerly mellowed or never lithified Omingonde Formation top. In the first case a partly angular unconformity between the conglomerates of the Triassic Omingonde Formation and those of the Early Cretaceous Krone Member should be expected, as the Krone Member of the adjacent Huab area acquaints a tectonically rather active phase. However, the identification of such an unconformity is rather difficult as the lithologies of the Krone Member and most parts of the Middle Omingonde Formation are rather similar.

In either case, the presumed time-stratigraphic gap between the top of the Omingonde Formation and the Etendeka Group embraces the time span from the (Middle ?) Triassic to the Lower Cretaceous. The possible presence of the Krone Member would not significantly affect the extent of this hiatus.

3.3.2.2 WATERBERG-ERONGO BASIN

In the Waterberg-Erongo Basin the Triassic Omingonde Formation is succeeded by the Early Jurassic Etjo Sandstone Formation in the Mt. Waterberg and Mt. Etjo region or in the Erongo region by the stratigraphically equivalent White Quartzite Unit of Hegenberger (1988) (Holzförster et al., 1999), now referred to as the Nieuwoudt Member (Milner, 1997). Keyser (1973), Löffler & Porada (1998) and Holzförster et al. (1999) describe an erosive unconformity on top of the Omingonde Formation associated with a significant hiatus.

Evidence for this gap is well constrained by outcrop geology at Mt. Etjo and Otjihaenamaperero 92 Farm. At Otjihaenamaperero 92 Farm the gap is associated with a well defined angular unconformity, which is exposed at the western flank of the river bed, approximately 600 m north-west of the farm housings (S21°02´52´´/E14°23´36´´, Fig. 3.4-3). Intensively folded coarse fluvial clastics of the Upper Omingonde Formation are overlain by slightly deformed to undeformed aeolianites of the Upper Etjo Formation with the entire lower and middle part of the Etjo Formation missing. This equates to at least 40-50 m of missing strata compared to the more completely developed sections at the Waterberg Plateau Park (chapters 3.2.2.1 & 3.2.2.2).

The Triassic and Jurassic sediments in the Waterberg-Erongo Basin were deposited in the tectonically active environment of the Waterberg-Omaruru transfer fault zone (Porada et al., 1996; Holzförster et al., 1999; Stollhofen, 1999; chapter 3.4.2), suggesting that the hiatus at the Triassic/Jurassic boundary could be locally exaggerated due to this structural setting. However, the similar facies architecture of the Etjo Formation in the Mt. Waterberg region and of equivalents in the main Karoo Basin suggests, that the described hiatus is of regional importance and probably correlates with a major 2nd order sequence boundary between the Molteno and Elliot Formation (Turner, 1999). It may also coincide with the major pre-Piramboia Group unconformity in the Paraná Basin if a latest Triassic/earliest Jurassic age (cf. Porada et al. 1996) for the latter is assumed. If these correlations are correct, the Triassic/Jurassic hiatus of the

Waterberg-Erongo Basin would embrace the missing stratigraphic equivalents of the entire up to 500 m thick South African Elliot Formation (cf. Smith et al., 1993). However, such an extent of the hiatus could only be confirmed with a much higher resolution of the biostratigraphic framework for the Upper Omingonde Formation in Namibia.

Another disconformity within the Waterberg-Erongo Basin is restricted to the area north and north-east of the Waterberg Plateau Park. It defines the boundary between the Etjo Formation and the Rundu Formation basalts. This disconformity reflects the onset of extensive flood basalt volcanism, which occurred probably contemporaneous to the Kalkrand Formation in southern Namibia (see chapter 3.2.4). Although only poor bore hole data exists (viewed at the Department of Water Affairs, Windhoek) a lack of a hiatal gap is well constrained: As shown in Figure 3.3-2 medium size grained sandstones beside minor mudstones are intercalated between basalt units. The sandstone horizons are often discontinuous and they probably form lens-shaped bodies between individual basalt units. According to the log notes (viewed at the Department of Water Affairs, Windhoek) grain size and sorting characteristics of many intercalated sands suggest an aeolian origin. The mudstones seem to be regularly associated with sandstones as no isolated mudstone layers are reported within the basalts. At basalt-sandstone contacts the sandstones are usually indurated due to contact-metamorphism caused by the hot basalts pouring onto the sediments. The amygdaloidal character of the basalts suggest that the majority of them formed as lavas rather than intrusions.

The lack of data on sedimentary structures makes the construction of a facies model difficult. It can be speculated that the majority of the sandstones represent aeolian deposits, that have been partly reworked by fluvial processes. The mudstones possibly record intermittent periods of aquatic deposition, that either occurred during waning stages of sheet floods or in shallow temporary lakes. The intense fringing of lavas with sediments suggest that intense lava-sediment interactions occurred. Therefore it is assumed that the basaltic lavas poured out onto an active fluvio-aeolian system. The aeolian deposits probably developed sandstone-lava interrelationships, which are similar to those reported from the Huab area (chapter 4.1.9.2). The aquatic mudstones might have been influenced by the lavas in a similar way as the lacustrine interlayers known from the Kalkrand Formation at Hardap (cf. Gerschütz et al. 1995).

In summary, the basalt-sediment interactions favour a depositional model in which basaltic lavas poured into an active aquatic-aeolian system. Therefore a depositional break between the sedimentary deposits (probably Upper Etjo Formation) and the mixed basalt-sediment deposits (Rundu Formation) seems to be rather unlikely.

Fig. 3.3-2

Well log comprising the uppermost Etjo Formation and the Rundu Formation. The Rundu Formation displays intense basalt-sediment interactions. Log from well 6452 Kanovlei (district Tsumkwe). Map on the right shows the location of boreholes with the depth of the first basalts. Well 6452 is highlighted with an arrow. Data obtained at the Department of Water Affairs, Windhoek. Legend

3.4 TECTONIC CONTROL ON THE TRIASSIC/JURASSIC SEQUENCE

Tectonic control on sedimentation is registered by fault related thickness variations and palaeo-current patterns in both depocentres, the Waterberg-Erongo Basin and the Goboboseb-Otjongundu Basin. The "Northern Central Zone" of the Damara orogen *sensu* Miller (1983) is located between the two basin systems. Since this zone was the formerly exposed mountainous hinterland that acted as a sediment source for either basins, Porada et al. (1996) introduced the term "Damaraland Uplift" (Fig. 3.4-1). The Waterberg-Omaruru Fault zone including the Ameib Line confines the Damaraland Uplift to the south. Its northern margin is marked by the Otjohorongo Thrust, the Autseib Fault and other parallel and conjugated faults. Stollhofen (1999) regards the SW-NE trending faults on either side of the "Northern Central Zone" as a transfer fault zone related to pre-South Atlantic rifting between Africa and South America. An inferred sinistral strike-slip component of these fault systems points to the presence of associated transtensional and pull-apart basins in which the Triassic-Jurassic sequence attained their maximum thickness. Despite to the presumed transtensional stress regime, the Waterberg Fault on Otjihaenamaperero 92 Farm and the Otjohorongo Thrust indicate strong compressive movements on both sides of the Damaraland Uplift. In the vicinity of both fault systems neo-proterozoic basement has been thrust onto Karoo rocks. Thrusting of Damara basement towards the south at the southern margin of the Damaraland uplift associated with northwards directed thrusting at the northern margin corresponds with transpressive stress and not with transtension. However, it could be argued that the observed compressive features are only locally developed and do not represent the prominent structural mode. Another possibility is tectonic inversion from a transtensive regime in the Triassic-Jurassic to transpression, that likely commenced with the breakup in the Lower Cretaceous. A detailed study of the structural patterns on Otjihaenamaperero 92 Farm showed, that the situation is even more complex. As described in chapter 2.4.2.1, the geometric relationships of the Waterberg Fault indicate alternating phases of extension and compression. A similar pattern can be concluded for the fault system delimiting the northern margin of the "Damaraland Uplift" as outlined in the next paragraph.

Fig. 3.4-1
Position of palaeo-highs in northern Namibia. The sedimentary influence of the "Damaraland Uplift" and the swell presumed by Miller (1997) already commenced in the higher Permian, whereas the "Skeleton Coast Uplift" became accentuated immediately prior to the deposition of the Etendeka Group.

3.4.1 GOBOBOSEB-OTJONGUNDU BASIN

Fluvial transports during deposition of the Triassic sequence favour north-westerly to north-easterly directions (Fig. 3.1-5), away from the Damaraland Uplift (Fig. 3.4-1). This coincides with a general southwardly directed thickening of the strata. In proximal positions to the basin confining faults (Otjongundu Plateau), also south-westwardly to southwardly directed palaeo-current vectors were observed (Porada et al., 1996). The south-direction refers to a depocentre near the basin confining faults, the south-west directions coincide with the depocentre-axis being parallel to the main tectonic trends.

An additional hint for an active fault delimiting the Goboboseb-Otjongundu Basin to the Damaraland Uplift is given by the facies architecture of the Middle Omingonde Formation at the Tafel Section (Mt. Brandberg, Fig. 3.1-4). This succession displays a superimposed upward-coarsening trend of stacked

upward-fining cycles. The superimposed upward-coarsening trend probably reflects a continuous rise of the source area, in which the uplift rate exceeded the contemporaneous denundation. It is here assumed, that this strong uplift has been facilitated by a predecessor structure of the Otjohorongo- and Autseib faults system. This predecessor structure acted as a synsedimentary fault of considerable magnitude.
Abundant sand filled fissures in the middle part of the preserved Triassic sequence are presumably related to strong seismic events, which support a general tectonic control on sedimentation.

3.4.2 WATERBERG-ERONGO BASIN

The synsedimentary tectonic control on the Triassic-Jurassic sequence is recorded by thickening of strata towards the Waterberg-Omaruru Fault, implying a persistent northern source area (Porada et al., 1996). The pronounced shift of the depocentre towards north-east (Fig. 3.4-2), discovered by Holzförster et al. (1998; 1999) suggests sinistral oblique-slip movements along the Waterberg-Omaruru Fault contemporaneous to deposition of the Triassic-Jurassic sequence.
A good example for a fault related thickness gradient is the Krantzberg Formation in the Erongo Mts. It reaches 700 m in thickness and displays an ultra-proximal facies (Hegenberger, 1988). The sequence abruptly terminates northwardly against the Ameib Line, but southwards it thins into the equivalent Lions Head Formation. Finally, at a distance of 40 km south of the Ameib Line, the thickness declines to about 30 m.
A continuous, north-east directed shift of the depocentre is indicated by north-east directed flow vectors and the fault-parallel thickening of individual depositional units in the same direction (Holzförster et al., 1999).
Other indications for syn-depositional tectonics are given by clastic dykes and convolute lamination structures which have been observed in the Erongo Mts. close to the Ameib Line. These structures are possibly related to earthquakes. In Addition, diamictites observed on Otjihaenamapereo 92 Farm are interpreted as mass flow deposits, which point to a tectonically rather active environment.

Fig. 3.4-2
Schematic cross sections of the Karoo Supergroup in the Waterberg-Erongo area illustrating thickness variations parallel and perpendicular to the Waterberg-Omaruru Fault (modified from Holzförster et al., 1999). The SW-NE section illustrates a north-east directed shift of the depocentre, which suggests a sinistral strike-slip movement of the Waterberg Fault. The N-S section shows that maximum accommodation space has been created immediately south to the fault.

3.4.2.1 THE ANATOMY OF THE WATERBERG-OMARURU FAULT ZONE

A portion of the Waterberg-Omaruru Lineament is well exposed in the vicinity of Mt. Klein Etjo, approximately 100 km south-west of Mt. Große Waterberg (Fig. 1-3). Due to good outcrops the anatomy of the fault was studied on Otjihaenamaperero 92 Farm. At this location the Waterberg Fault has the attitude of a thrust, juxtaposing neo-proterozoic Damara basement over Mesozoic rocks (Fig. 3.4-3).
At a closer look, at almost every locality, Damara limestone is juxtaposed over Middle, or in a few cases possible Upper Omingonde Formation. Damara rocks directly overly the Etjo Formation only at one locality. The corresponding shear zone is dipping towards the NW with 25-30°, suggesting that the SSW trend of the Karoo-Damara rock contact is more an erosional feature than structurally accommodated.

Thrust faulting is supported by the mainly down-dip of slip lineations from within the basal part of the Damara sequence.

An erosional half-window through the Damara Sequence exposes a zone of folded and faulted Omingonde Formation rocks. Widely varying azimuths and plunges of fold-axes (Fig. 3.4-3, Fig. 3.4-4), as well as small circle distributions of poles to bedding (Fig. 3.4-4), suggest a history of folding and refolding. The spatial distribution of fold-axes orientation offers a subdivision into two, or possibly three sub-zones in a NW-SE traverse (Fig. 3.4-5): (1) A zone of relatively uniformly steeply (50-75°) north-westwardly dipping beds of the Omingonde Formation, separated by a steeply dipping fault plane from (2) a narrow zone with roughly fault-parallel fold-axes grading to more south-westerly azimuths and highly variable dip-angles.

Intraformational slump folding has been interpreted as soft sediment deformation, possibly induced by seismic activity of the Waterberg Fault. This second zone is followed by (3) a wider zone of slightly deformed Omingonde Formation rocks.

The overlying Etjo Formation is largely unaffected by deformations. Only two isolated slices of Etjo Sandstone, each less than 10 m thick, have been fault-parallel tilted, with dipping 10-20° to the NW. In addition, slump-folding within the Etjo Formation has been observed south of the main study area at the south-western cliff of Mt. Etjo.

The combined stratigraphic data and structural observations permit the following conclusions (Fig. 3.4-6): (1) A phase of extensional movements along the Waterberg Fault and subsequent subsidence of the depository, which accommodates the Triassic Omingonde Formation. (2) A subsequent temporarily transpressive regime in the vicinity of the fault resulting in a zone of dominantly SW-NE trending fold-axes in the Omingonde Formation. (3) A tectonically relatively quiet phase with continuing or re-commenced subsidence during accumulation of at least 140 m Jurassic Etjo Sandstone Formation. (4) South-east directed thrusting of Damara basement onto Omingonde and Etjo Formation rocks. The loading of the Damara thrust sheet caused the underlying rocks to flex, resulting in an overall NW directed dip slope in the north-western portion of the deformed zone.

In summary, the Waterberg Fault system generated repeatedly subsiding depositories facilitating the accumulation of the Triassic Omingonde Formation. Probably these depositories classify as pull-apart and transtension basins, favouring generally transtensive stress patterns. A phase of tectonic inversion caused transpressive folding of the Omingonde Formation which exaggerated the hiatus to the overlying Etjo Formation (chapter 3.3.2.2.).

White et al. (2000) voiced the problem, that usually steeply dipping transtensional faults are difficult to transform in shallow thrusts like the present shallow (25-30°) north-westwardly dipping Waterberg Fault thrust. They therefore suggest that development of the sedimentary depository may initially have been controlled by another, more steeply dipping, precursor structure which potentially lies somewhere to the north-west underneath the thrust fault.

It should be noted at this place, that the thrust patterns observed at Otjihaenamaperero 92 Farm are only locally observed features and cannot confidently extrapolated to the whole Waterberg-Omaruru Fault system.

Fig. 3.4-3
Map of the Waterberg Fault on Otjihaenamapereo 92 Farm. The main fault strike is SSW-NNE. Thrusting of Damara rocks over Karoo rocks occurred in south-easterly direction.

Fig. 3.4-4
Orientation of fold axes and poles to bedding in the vicinity of the Waterberg Fault on Dinos Farm.

Fig. 3.4-5
Cross sections of the Waterberg-Omaruru Fault on Otjihaenamapereo 92 Farm. The NW-SE sections (A-B and C-D) reveal three tectonic zones of Karoo rocks: (1) Uniformly north-westwardly dipping Omingonde Formation, (2) intensely folded Omingonde Formation with a narrow zone of north-easterly oriented fold axes followed by a zone of more variable azimuths and (3) slightly deformed Omingonde Formation succeeded by almost undeformed Etjo Formation rocks. The fault parallel section (E-F) shows folding due to NW and SE plunging fold axes. See Figure 3.4-3 for section locations and Figure 3.4-4 for fold axes orientations. Sections compiled together with White.

Fig. 3.4-6
Cartoon illustrating the history of the Waterberg Fault at Otjihaenamapereo 92 Farm. For explanation see text in chapter 3.4.2.1.

3.5 TRIASSIC/JURASSIC TECTONIC EVOLUTION OF NORTH-WESTERN NAMIBIA

The Triassic/Jurassic tectonic evolution is characterised by the increasing control of SW-NE trending structures on sedimentation. Of particular importance is the SW trending Damaraland Uplift (Fig. 3.4-1) that already acted as a sediment source in the Uppermost Permian which is notified by the gritstone deposits of the Doros Formation in the Goboboseb Basin. A subsequent, more regional uplift was probably inducing the hiatus following above Doros Formation, or in respect to the area south of the Damaraland Uplift, the Tevrede Formation. Furthermore, this hiatus can be traced into the Ovambo Basin (Fig. 1.2-4), where according to Miller (1997) Prince Albert Formation equivalents are conformably followed by possible Etjo Formation equivalents. Miller (1997) attributes this hiatus to a phase of non-deposition on a long-lived smooth swell, running N-S through Namibia (Fig. 3.4-1). He favours this swell model, because of the lack of angular intra-Karoo unconformities and the absence of reworked Karoo rocks. This swell model conforms to a model of Stollhofen (1999) who considered a regional uplift of western Namibia related to early stages of southern South Atlantic rifting. Compensation of the variable timing and the variable amounts of extension along a northwardly prograding rift structure requires conjugated transfer faults (cf. Gibbs, 1990). In this context the Waterberg-Omaruru Fault zone acted as a sinistral strike-slip fault and facilitated accommodation space for the Triassic to Early Jurassic sequence in associated structural depressions. Almost contemporaneously the halfgraben-shaped Goboboseb-Otjongundu Basin system formed farther north. The associated faults (Autseib-, Otjohorongo Fault) might have been affected by an oblique-slip component similar to the Waterberg Fault, which is not yet confirmed so far. It even has not been fully understood, whether the fault systems on either side of the Damaraland Uplift were mainly compressive or extensional in character during the Triassic-Jurassic. Both cases could have resulted in the permanently relatively elevated position of the Damaraland Uplift: In a compressive regime the

Damaraland Uplift would be regarded as a large-scale positive flower structure (cf. Woodcock & Fischer, 1986) promoted by a lateral-slip component of the associated faults. A transtensive regime would favour pull-apart basins and negative flower structures (cf. Woodcock & Fischer, 1986) on either sides of the Damaraland Uplift. The concept of a large scale positive flower structure would explain the uplift of the tectonostratigraphic "Northern Central Zone" (sensu Miller, 1983) but it neglects the enhanced subsidence that occurred during the Triassic-Jurassic at either side of the uplift. Therefore the concept of a general transtension is more plausible, since it explains both, subsidence and the pronounced shift of the depocentre documented by Holzförster et al. (1999). Probably temporary tectonic inversion induced a transpressive phase, which is evident by the folding of the Omingonde Formation at Otjihaenamaperero 92 Farm.

The south Namibian counterpart to the northern Namibian Damaraland Uplift (Omaruru ridge *sensu* Stollhofen, 1999) is the Karasburg ridge, which is also flanked by adjacent SW-NE trending fault systems. Both ridges can be traced offshore Namibia by the sinistral offset of oceanic magnetic anomalies (Fig. 3.5-1), which supports a sinistral strike-slip component of the associated fault systems. Those sinistral offsets commenced at least with the oceanisation in the Lower Cretaceous, but definitely started much earlier along the Waterberg-Omaruru Fault zone where a pronounced shift of the Triassic-Jurassic depocentre has been recognised.

As outlined above, the fault zones flanking the Damaraland Uplift facilitated localised enhanced subsidence and associated enhanced sedimentary thicknesses. Holzförster et al. (1999) and Stollhofen (1999) emphasised the additional importance of the persistent rift-related uplift in western Namibia. According to this uplift proximal facies assemblages of the Early Triassic to Early Jurassic sediments developed as one moves from the Mt. Waterberg area towards the south-west, closer to the present continental margin.

In summary, the tectonic setting of the Triassic-Jurassic megasequence in Namibia is characterised by pre-South Atlantic rift structures, reflecting a long-lived evolution of southern Gondwana rifting. The corresponding facies architecture of the Namibian megasequence to correlatives in the South African main Karoo Basin, that developed essentially in a foreland basin setting, indicates that rift and foreland basin evolution were influenced by the Cape Fold Belt, resulting in a similar sequence development in both tectonic settings. Trouw & De Wit (1999) reveal a strong relationship between collisional processes in the Gondwanide Orogen (Cape-Ventana-Fold Belt) and intracratonic deformation of a compressional, strike-slip and extensional nature (Fig. 3.5-2). From this perspective, the South African as well as the Namibian unconformity separating the Triassic from the Jurassic sequence might relate to an orogenic pulse in the Cape Fold Belt, dated at 215±5 Ma (Hälbich et al., 1983).

Fig. 3.5-1

Recent arrangement and denomination of ocean floor magnetic anomalies in the offshore area of southern Africa (compiled from Fouché et al., 1992 and Dingle, 1992/93). The Waterberg-Omaruru Lineament intersects anomalies M 2 and M 4 due to its sinistral strike-slip movement. Probably the Waterberg-Omaruru Lineament already commenced its tectonic control in the Mesozoic.

Fig. 3.5-2

Gondwana reconstruction showing the major Permo-Triassic structures of central Africa and adjacent Madagascar (modified from Trouw & De Wit, 1999). Orientation of large scale field stresses and tectonic movements are indicated with arrows.

Chapter 4: Etendeka Group Stratigraphy

The Lower Cretaceous Etendeka Group follows on Karoo respectively Damara rocks with a major unconformity (chapter 5.1). The Etendeka Group comprises mainly volcanic and subordinate sedimentary rocks. Milner et al. (1994) presented the first subdivision of the Etendeka Group with three major formations: The sedimentary Etjo Formation, (now named Twyfelfontein Formation, see below), the volcanic Awahab Formation and the Tafelberg Formation. Detailed studies have identified further sub-units within the Etendeka Group in both, the sedimentary and volcanic successions (e.g. Mountney et al., 1998; Jerram et al., 1999b). In the Huab area and at a few coastal locations the fluvial and aeolian units of the Lower Cretaceous Twyfelfontein Formation form the base for the volcanic succession.
These sediments have been deposited immediately prior and during early stages of Etendeka-Paraná flood basalt volcanism. Because of their close age relation to the overlying lavas (chapter 4.1.10) they are stratigraphically placed in the lower part of the Etendeka Group.

4.1 STRATIGRAPHY AND FACIES OF THE TWYFELFONTEIN FORMATION

4.1.1 TERMINOLOGY

Traditionally this formation has been ascribed Etjo Sandstone Formation according to lithostratigraphic correlation with aeolianites at Mt. Etjo (e.g. SACS, 1980; Miller & Schalk, 1980). There and in the Huab area the aeolianites form the top of the sedimentary succession and it has been believed that both were deposited synchronously. But as Early Cretaceous K/Ar ages were derived from lavas above the "Etjo-correlatives" in the Etendeka sequence (Siedner & Mitchell, 1976) Martin (1982), Porada et al. (1996) and Löffler & Porada (1998) speculated, that the aeolianites in the Huab area could be substantial younger than the Etjo Formation in its Mt. Etjo and Mt. Waterberg type locality. Following the separation of the Karoo Supergroup from the Etendeka Group as proposed by Milner et al. (1994), Stollhofen (1999) and Stanistreet & Stollhofen (1999) consequently introduced the term Twyfelfontein Formation for the Lower Cretaceous aeolianites in the Huab area. This is consistent with recent $^{40}Ar/^{39}Ar$ dates of 132.3±0.7 Ma for Etendeka Lavas that are interfingering with aeolianites of the Upper Twyfelfontein Formation (Renne, written comm. 1997). In contrast to these age constraints of the aeolianites in the Huab area, Holzförster et al. (1999) confirmed a Lower Jurassic Karoo age for the Etjo beds at Mt. Waterberg by the identification of the dinosaur *Massospondylus* sp. (chapter 3.2.4).

4.1.2 INTRODUCTION

The Twyfelfontein Formation consists of aeolian, fluvial and mass flow deposits representing an aridification succession. It is exposed over 5000 km^2 in the Huab Basin area and in a small extent 20 km north of Terrace Bay at the Skeleton Coast. On a basin wide scale the Twyfelfontein Formation forms a northwardly thickening wedge shaped sediment body tracing a half graben geometry.
The complex spatial and temporal variations in depositional style are controlled by tectonic faulting and a gradual transition from semi-arid to hyper-arid climatic conditions.
Mountney et al. (1998) divide this succession into four units, each is related to a distinct phase in sedimentary evolution. The succession starts with the basal alluvial-fluvial Krone Member, followed by a Mixed Aeolian-Fluvial Unit, then a Main Aeolian Unit and finally an Upper Aeolian Unit. At places the top

of the Main Aeolian Unit is marked by the basal Etendeka lavas, which divide the Main Aeolian Unit from the succeeding Upper Aeolian Unit (Fig. 4.1-1). The latter fringes with the lower Etendeka lavas and finally has been inundated by them.

Fig. 4.1-1
Stratigraphic overview section of the Huab area.

4.1.3 CORRELATIVES

Possible equivalents of the Twyfelfontein Formation are the Botucatu Formation in Brazil (Ledendecker, 1992) and Aeolianites of the Pirgua Group of north-western Argentina. The Botucatu Formation can be well correlated with the aeolianites of the Huab area as they both interfinger with the Lower Cretaceous Paraná-Etendeka flood basalts (Bigarella, 1970) (Fig. 2-1).
Ledendecker (1992) ascribed the Jurassic Clarens Sandstone Formation of the main Karoo Basin as an equivalent to the Twyfelfontein Formation, but due to its newly established Lower Cretaceous age this is no longer valid. The same applies to former correlations with the Ntane and Nkalatou formations in Botswana, and the Forest Sandstone Formation in Zimbabwe (Johnson et al. , 1996).

4.1.4 GEOMETRY OF THE TWYFELFONTEIN FORMATION IN THE HUAB AREA

The geometry of the Huab erg (fluvial-aeolian and aeolian units) can be described as a wedge with maximum thickness around the present day Huab River and pinching out to the west and to the south. North of the Huab River the pinching out occurs rapidly, but local deposits are found at many places as far as in the Kaokoveld (Hodgson, 1970). Therefore minor occurrences may exist under the Etendeka lavas north of the Huab area. Despite this, the southern margin is rather distinct and only minimal relicts of Cretaceous aeolianites have been observed south of the Ugab River. This distribution of the Huab-erg widely complies with the thickness-distribution of the Krone Member, although the latter is more restricted. Basically, distribution and thicknesses follow the pre-Etendeka Group/Twyfelfontein Formation basin morphology and the implied accommodation space. Basin morphology was mainly tectonically modified prior to the onset of Krone Member deposition (chapter 5.1.1.1), but nevertheless tectonic basin evolution continued after that and controlled further deposition.

4.1.5 KRONE MEMBER

The Krone Member forms the alluvial and fluvial base of the Twyfelfontein Formation. It erodes unconformably into the underlying Karoo succession. The spatial distribution of the Krone Member is rather restricted (Fig. 4.1-2). It appears to be strongly coupled with a fault induced palaeo-relief.

4.1.5.1 KRONE MEMBER DISTRIBUTION

The southernmost occurrences of the Krone Member are the outcrops north of Doros Crater, where the Krone Member consists of an up to 1.5 m thick alternation of conglomerates and gritty sandstones. Towards the north the thickness is continuously increasing. At the axis along the present day Huab River the maximum thickness of 10 m is attained, suggesting that the depocentre of the Krone Member coincides with it. Along a section from the Huab River to Klein Gai-As the thickness decreases gradually

until the Krone Member pinches out into a thin pebble layer. The pebbles of this layer are distinctly wind grooved and partly wind faceted indicating a palaeo-deflation surface. Some local rapid thickness variations have been observed east of the Bergsig Fault system, between the northern and eastern edge of the outcrop area (Fig. 4.1-3). Those thickness variations probably correspondent with an enhanced palaeo-relief. Particularly in the coastal region the sporadic extent of the Krone Member is restricted to small palaeo-depressions.
Planar cross-bedding and clast imbrication indicate current directions towards NNW and subordinate to NNE, while in the depocentre also westerly and easterly directions have also been observed (Fig. 4.1-2).
West of Doros Crater, at Gai-As and at Mt. Bruin, the Krone Member is missing. Thus in between these locations the Krone Member forms 5-10 km wide channels which become wider and thicker northwards and finally amalgamate laterally into a continuous sheet reaching as far as 3 km north of the present day Huab River. Farther west towards the coast, outcrops are only found north of the Huab River in at least 8 km distance from the coastline.

Potential equivalents of the Krone Member occur north of the Huab area and at the Albin Ridge (see Fig. 1-3 for location), as described in chapter 4.1.6.

Fig. 4.1-2
Distribution of the Krone Member in the Huab area and mean palaeo-current vectors.

4.1.5.2 KRONE MEMBER LITHOLOGY

The Krone Member comprises conglomerates, coarse sandstones and subordinate mudstones. The conglomerates are usually clast-supported with imbrication (Plate II-4) and planar cross-bedding. Matrix-supported conglomerates show transitions to pebbly sandstones. Trough cross-bedding and faint subhorizontal layering are rare (Plate II-2). The conglomerates consist of poorly to moderately sorted but well rounded pebbles and cobbles of predominantly Damara vein quartz, Damara quartzite, Damara granites, rarely Karoo sediments and very rarely volcanic rocks. Sorting and rounding of the pebbles becomes better towards the south-west. The sandstones are coarse-grained and often pebbly. Local interlayers of finer grained and better sorted horizons occur. Planar cross-bedding is common.
The sediment bodies are planar or form up to 30 m wide channels, that often have a deeply scoured base. Channel fill units are usually simple, although complex, multi-storey infills were observed. Where the conglomerates are overlain by fine clastics, desiccation cracks are sometimes preserved (Plate II-3).

4.1.5.3 KRONE MEMBER FACIES

The Krone Member comprises wadi, braidplain and mass flow deposits. The moderate sorting and clast imbrication together with the mainly clast supported nature of the conglomerates indicate bedload deposition. Traction transport within stream channels is indicated by well developed erosive channel bases and the presence of planar and trough cross-bedding. The relatively high abundance of horizontally laminated sandstone and pebbly sandstone horizons, interbedded with the conglomerates, is assigned to high energy deposition by flash floods and ephemeral streams. The finer grained deposits and the desiccation features developed with the decrease of flow energy and subsequent drying out.

Palaeo-current data yield a major fluvial system flowing to the south-west like the present day Huab River (Fig. 4.1-2). This interpretation is consistent with the increase in rounding and sorting of the sand and pebble component to the south-west. Tributary streams in the south flowed north-westwards into the main system. In the south these streams are relatively isolated, but northwards, towards the main system, they become wider and laterally amalgamated. The area north of the Huab River is dominated by proximal alluvial deposits with aeolian intercalations. In the southern area, braidplain deposits dominate. In the west local ephemeral streamchannel deposits are preserved.

Northern Krone Member facies

North of the Huab River the Krone Member is represented by coarse, poorly sorted, clast-supported conglomerates, which show a mixed alluvial-aeolian transition in the upper parts. The poorly sorted angular to subangular clasts are indicative of a proximal to medial alluvial fan setting. Palaeo-current data yield south directed flow directions (Hodgson, 1972). The depocentre was located only a few km south of the northernmost outcrop area of the Krone Member. There, up to 10 m thick beds of coarse clast-supported conglomerates accumulated. Around the depocentre the palaeo-flow directions turn from southwards to south-westwards (Fig. 4.1-2). Sources of the northern Krone Member were probably locally uplifted scarps that formed the northern basin margin.

Southern Krone Member facies

South of the Huab River pebbly sandstones, coarse sandstones and sandy conglomerates are predominant. Clast-supported conglomerates are restricted. Short distinct transitions to aeolianites are common. The widespread red colour of the Krone Member and the interfingering with aeolian deposits reflect an arid environment. Palaeo-current directions are uniform to the north-west. The moderate, continuous increase in sediment thickness towards the depocentre in the north suggests a gentle northwardly dipping palaeo-slope. The resulting sediment body has a sheet-like geometry with pinching out towards its southern margin. This sheet-like geometry together with the almost unidirectional current pattern are indicative for braidplain or alluvial plain deposits (Rust & Koster, 1984). A distal position is shown by the good rounding of the Damara pebbles, that derived from a source further in the south.

The southern area of the Krone braidplain has been rather exposed prior to the deposition of the succeeding aeolian units. This is indicated by wind-grooved and partly wind-faceted pebbles which reveal a palaeo-deflation plain.

Western Krone Member facies

In the Skeleton Coast Park and immediately east of it occurrences of the Krone Member are restricted to isolated depositories which were found north and immediately south of the Huab River. Palaeo-current vectors are highly variable, but transport to the west is still dominant (Fig. 4.1-2). Similar to the southern area, pebbly sandstones and matrix-supported conglomerates dominate. Planar and trough cross-bedding as well as a subhorizontal layering indicate sedimentation by flash-floods. The sediment thicknesses vary from a few centimetres to 1 metre. Usually a distinct transition to aeolian sandstones follows up-section. This together with the thickness variations is interpreted as ephemeral or wadi-like deposits in an uneven marginal depository. The good rounding of the pebbles together with the absence of large pebbles indicate a rather distal position. With respect to the source, the western Krone Member represents the sediment portion being transported westwardly across the main depocentre. The distal character of the western Krone Member, expressed by good rounding and sorting, excludes a source in the adjacent coastal area. Ledendecker (1992) assumed a coastal barrier which forced eastwardly directed flow patterns. A coastal high is reflected by the rather restricted distal facies occurring in the Skeleton Coast Park. Therefore it is assumed that drainage to the west was rather restricted.

Fig. 4.1-3

Tectonic interpretation of aerial photographs of the eastern Huab area and reconstruction of the syn-Krone Member relief. The correlation of the Krone Member is based on Huab Formation maximum flooding surface equivalents (not shown). This correlation roughly reflects the syn-Krone Member relief presuming that the main tectonically induced vertical offsets occurred immediately prior to and during Krone Member deposition. Major changes in palaeo-current directions occur across the faults indicated in the aerial photographs. Therefore it is concluded that these faults already influenced the topography and subsequently the current patterns during Krone Member deposition.

4.1.6 COASTAL EQUIVALENTS TO THE KRONE MEMBER

Lithological equivalents have been found at Albin Ridge, at Sanianab Locality, north of Terrace Bay and at Kharu-Gaiseb locality (see Fig. 1-3 for locations). They all belong to the Etendeka Group, but probably they were not all deposited contemporaneously.

4.1.6.1 ALBIN RIDGE CONGLOMERATE

The Albin Ridge is located 15 km east of the present Atlantic coast. It is running parallel to the present day coastline (NNW-SSE) and exposes sedimentary and volcanic strata. The strata is eastwardly dipping with dip-angles between 20-30°. This inclination is associated with a listric fault, which runs along the western flank of the ridge. The northern outcrop area (type section is located at S21°25´34´´/E13°53´44´´, Fig. 4.1-4) comprises the Karoo Sequence with the Permian Verbrandeberg, Tsarabis, Huab and Gai-As formations. A thin sequence of polymict conglomerates, informally termed the Albin Conglomerate (Swart, 1992/93) overlies these formations disconformably (Fig. 4.1-4, Plate II-1). A thin palaeosol is developed below the contact. In the southern part of the ridge the conglomerate rests directly on Damara basement rocks. Etendeka basalts and a quartz latite unit succeed the clastics and form the top of the ridge.

The Albin Conglomerate varies in thickness between 5 and 60 m. Bedding is poor, the individual beds exhibit normal grain size grading. Imbrication and a horizontal orientation of clasts are occasionally developed, but neither is common. Maximum clast size exceeds 0.5 m and the conglomerate is both matrix- and clast-supported (Swart, 1992/93). Near the top of the sequence sandstones and pebbly sandstones are interbedded. They comprise well-sorted and rounded sand, containing quartz and feldspar. The pebbles are up to 3 cm in diameter and consist of vein quartz.

The components of the conglomerate consist of poorly sorted but well rounded pebbles and cobbles of predominantly Damara quartzite, Damara granites, Damara vein quartz, basalts and seldom Karoo sediments. The presence of basaltic clasts is of special interest, because it might indicate a close age-relation to the overlying Etendeka basalts.

The Albin conglomerates are interpreted as debris flows with intercalations of braided stream deposits. They might be time-equivalents to the Krone Member in the Huab Basin. The sandstones near the top of the sequence could be the southern counterpart to the Mixed Aeolian-Fluvial Unit (chapter 4.1-7) of the Twyfelfontein Formation. However, time-equivalence is not fully constrained and the presence of basaltic clasts suggests a slightly younger age (chapter 4.1.12.1). The nearby presence of a prominent listric fault confirms the assumption that the source for the coarse clastics was a fault generated scarp in a rather proximal position.

Fig. 4.1-4
Section of sedimentary strata at the northern end of Albin Ridge (S20°25´34´´/E13°53´44´´). Legend

4.1.6.2 CONGLOMERATES AT SANIANAB LOCALITY

A small exposure of conglomeratic red-beds is located at the Sanianab River 20 km east of the Skeleton Coast (location at S20°01´58´´/E19°12´07´´, see Fig. 1-3). There, an up to 4 m thick sequence of monomict conglomerates and pebbly sandstones overlies Damara gneiss and is capped by aeolianites of the Twyfelfontein Formation. The top of the exposure is formed by flatly lying basaltic lavas which belong to the Awahab Formation (chapter 4.2.1).

Remarkable are the structural relationships: The gneiss footwall and the overlying red-beds are tilted slightly eastwards with a dip angle of up to 10°. The aeolianites drape this tilt-block relief and therefore form a wedge shaped body (Fig. 5.3-2).

A discussion of these structures is given in chapter 5.3.3.1.

The conglomerates are fully matrix-supported and are interbedded with pebbly sandstones. The angular to sub-angular clasts are up to 4 cm in diameter and consist uniformly of Damara gneiss. The matrix is sandy and in the lower parts predominantly reddish in colour. The matrix within the uppermost 30 cm is relatively pale in colour.

At least the upper 30 cm of the beds are interpreted as sheet-flood deposits with varying transport energies as reflected by the varying pebble abundance.

A distinct interpretation of the lower part of the succession is problematic due to poor exposure.

Stollhofen (1999) allocates the up to 3.5 m thick reddish lower part of the sequence to the Permian Doros Formation, the uppermost 30 cm to the Krone Member of the Twyfelfontein Formation. The inferred unconformity between the two units is marked by a marble textured, up to 10 cm thick horizon, which is interpreted as a palaeosol.

Synsedimentary micro-faults in the uppermost 30 cm extend into the aeolianites which proves a depositional age similar for both units. This fact strongly supports the assumption for the upper part of the pebbly sequence being the Krone Member.

Down-section no more synsedimentary faults have been observed. Instead, subvertical fissures occur within the uppermost 30 cm of the reddish unit. They are about 3 cm wide and filled with white sandstone. The sand probably derived from the Twyfelfontein Formation aeolianites, which follow up-section. The brittle fissures confirm that the lower pebbly sandstone unit is older than the Krone Member. This fact supports the presence of the Permian-Cretaceous unconformity postulated by Stollhofen (1999). But, it should be mentioned that the Permian age of the lower pebbly sandstone unit is not yet proven.

4.1.6.3 PEBBLY SANDSTONES NORTH OF TERRACE BAY

A poor exposure of immature, partly pebbly sandstones occurs 20 km north of Terrace Bay about 200 m east of the shoreline (location at S19°51´34´´/E12°57´53´´, see Fig. 1-3). The unit is up to 50 cm thick and directly rests on Damara gneiss. Up-section the pebbly sandstone unit is succeeded by the aeolianites of the Twyfelfontein Formation. Rounding and sorting of the sands are poor. The pebbles are up to 8 mm in diameter and consist of Damara gneiss. Small scale cross-bedding, trough cross-bedding, climbing-upwards ripples and normal grading are common sedimentary structures (Fig. 4.1-5). Most of these structures were observed on loose blocks, and therefore palaeo-current measurements were not obtained. The sedimentary structures indicate high current velocities with high sediment input. The immaturity of the sands confirms a purely fluvial origin and not a primary aeolian source. Therefore the sandstones have a stronger affinity to the Krone Member than to the Mixed Aeolian-Fluvial Unit of the Twyfelfontein Formation.

Fig. 4.1-5
Cross-bedding in pebbly sandstones of the Krone Member north of Terrace Bay. Outcrop is located 20 m west of the road, approximately 20 km north of Terrace Bay (S19°51´34´´/E12°57´53´´).

4.1.6.4 KHARU-GAISEB CONGLOMERATES

Ward & Martin (1987) describe a rudaceous conglomerate, which is exposed 20 km north-east of Terrace Bay. They discovered several small outcrops south and north of the Kharu-Gaiseb River (main deposit at S19°51´25´´/E13°37´10´´). The sequence is at least 500 m thick and has been informally termed Kharu-Gaiseb Conglomerate by Ward & Martin (1987). It is basically clast-supported with almost no sorting and rounding. Boulders and cobbles are rather abundant and no overall upward-fining trend can be recognised (Ward & Martin, 1987). Due to the rudaceous character sedimentary structures are indistinct, but cross-bedding and imbrication of larger clasts indicate transport from the west (Ward & Martin, 1987). The clasts and the matrix consists entirely of volcanic material which must have been derived from the surrounding Etendeka Group volcanics.

The Kharu-Gaiseb Conglomerate is interpreted as proximal debris flows fed from a scarp in the west. Ward & Martin (1987) explain the missing of a upward-fining trend with a fault created scarp-source of constant elevation during its erosion. This relatively constant elevation of the scarp implies a growth fault of considerable magnitude.

4.1.7 MIXED AEOLIAN-FLUVIAL UNIT

An up to 30 m thick unit, comprising fluvial and aeolian sandstones, conformably overlies the Krone Member. This unit records a transition from pebble dominated deposition to pure sand deposition (Fig. 4.1-6). The fluvio-aeolian sequence comprises 2 to 5 cycles, that are related to upward-wettening or upward-drying trends, respectively (Mountney et al., 1998).

4.1.7.1 MIXED AEOLIAN-FLUVIAL UNIT DISTRIBUTION

The Mixed Aeolian-Fluvial Unit is widespread in the central Huab area north and north-east of Gai-As until immediately north of the present day Huab River. At the southern basin margin, as well as in the coastal area it is restricted to small hangingwall traps. Usually the unit follows with a gradual or sharp transition onto the Krone Member. Only at Ambrosiusberg Locality the upper part of this unit rests directly on Karoo rocks, while the lower part is missing (Fig. 5.3-3).

4.1.7.2 MIXED AEOLIAN-FLUVIAL UNIT LITHOLOGY

The basal sediments of the Mixed Aeolian-Fluvial Unit consist of sheet like deposits of poorly to moderately sorted medium-grained sandstones, with rare to abundant small quartz pebbles. It is usually overlain by bimodally sorted, planar, thinly laminated, fine to medium grained, rippled sandstone that forms 5-20 mm thick laminae. These rippled sandstone beds are laterally continuous for up to several metres and they sometimes display slight inverse grading but have no preserved foresets (Mountney et al., 1998). The overlying deposits are usually cross-laminated and grade up into steep cross-stratified beds of about 2 m in thickness.

At places the basal sheet-like deposits are not overlain by the planar laminated sandstones, but by horizontal to low angle laminated sandstones that show a fine crinkly texture.

This pattern of basal sheet like deposits followed by laminated, mostly rippled sandstones and cross-stratified beds on top, is repeated 2 to 5 times (Fig. 4.1-6). Each of these cycles is between 5 and 10 m thick. The cyclicity is locally interrupted by erosive pebbly sandstone channel-fill units.

Both, within the cycles and as an overall trend, the abundance and size of pebbles decrease up-section.

Fig. 4.1-6

Section of the Twyfelfontein Formation at the south-western cliff of the Huab Outliers, approximately 20 km south of the Huab River (S20°45´03´´/E14°06´46´´). The Mixed Aeolian-Fluvial Unit comprises two upward-drying cycles.

Legend

4.1.7.3 MIXED AEOLIAN-FLUVIAL UNIT PALAEO-TRANSPORT ANALYSIS

The poor to moderately sorted sandstone and pebbly sandstone units show north-westwardly directed palaeo-currents that are roughly conform with the underlying Krone Member (Fig. 4.1-7). Near the coast, in the vicinity of the Ambrosiusberg Fault (see Fig. 2.1-4 for location) the mean palaeo-current direction is straight towards the east.

Aeolian transport is indicated by the well sorted, steeply cross-stratified sandstones. They reveal a bimodal distributed aeolian transport pattern: In the southern vicinity variable south-east directions are dominant, whereas farther north south-west transport directions, variable with 90°, predominate (Fig. 4.1-7).

4.1.7.4 MIXED AEOLIAN-FLUVIAL UNIT FACIES

The poorly to moderately sorted sandstones of the Mixed Aeolian-Fluvial Unit are interpreted as fluvial deposits. The non-erosive, laminated, sheet-like sediment-body geometries, the small pebble size and the low abundance of channels suggest transport by low energy sheetfloods. The rare crinkly textured

sandstones are interpreted as adhesion planes of saltating grains on a dump surface (Kocurek, 1981). They indicate the close proximity of the water table to the surface (Mountney et al., 1998).
The better sorted, cross-stratified deposits are of aeolian origin. The rapid upward steepening of cross-laminae within the basal part indicate dune bedforms with primary amplitudes of less than 10 m. Additionally the variability of foreset orientation together with their restricted lateral continuity indicate small isolated dunes with curved slipfaces. The relative immaturity of the sand reflects a proximal sand source, probably the fluvial deposits of the unit.

The initial stage of deposition was highly controlled by the palaeo-topography created prior and during the deposition of the Krone Member as expressed by the concentration of the aeolian facies along restricted palaeo-valleys.

Mountney et al. (1998) suggest a facies model in which each cycle of deposition within the Mixed Aeolian-Fluvial Unit represents a change in depositional environment from ephemeral fluvial channel or sheetflood at the base, through damp and/ or dry aeolian sandsheet and finally into small aeolian barchan dune deposits. Each cycle is interpreted to represent a drop in the local water table controlled by climatic and/ or tectonic factors.

Fig. 4.1-7

Palaeo-transport patterns of the Mixed Aeolian-Fluvial Unit. The fluvial deposits reveal north-westwardly directed palaeo-currents, which are similar to those of the Krone Member. Dip directions of dune foresets reveal bimodal palaeo-wind directions in the aeolian deposits: In the southern vicinity of the outcrop area palaeo-winds from north-westerly directions are favoured, whereas in the northern vicinity palaeo-winds are from north-easterly directions.

4.1.8 MAIN AEOLIAN UNIT

The Main Aeolian Unit is an up to 150 m thick purely aeolian large scale cross-bedded deposit comprising 10 to 50 m thick individual bed-sets. They are interpreted as large transverse and draa dunes. Palaeo-wind direction indicate an unidirectional wind from WSW (Fig. 4.1-8).

4.1.8.1 MAIN AEOLIAN UNIT DISTRIBUTION

The Main Aeolian Unit is rather extensive in the central and eastern Huab area and in the Bloukrans Graben. It covers both, Karoo rocks and Damara Basement. It is here assumed, that the Huab area east of the Uniab Fault system was formerly almost completely covered by the Main Aeolian Unit.
Exceptions were some palaeo-highs, e.g. Mt. Bruin, where the entire Twyfelfontein Formation seems to be primarily missing. In the coastal region the Main Aeolian Unit is restricted to small occurrences, that often reveal hangingwall traps, which are usually confined by N-S striking normal faults.
North of the Huab River and east of Twyfelfontein the Main Aeolian Unit rests on Damara basement. At Doros Crater and further south it is missing.

4.1.8.2 MAIN AEOLIAN UNIT LITHOLOGY

This deposit consists exclusively of fine to medium grained, yellow-white quartz sandstones that can be classified as litharenites or sublitharenites according to the scheme of Pettijohn et al. (1973). A high grade in structural maturity is documented by the well rounded and sorted grains. Cementation is highly variable, but preferentially poor resulting in high porosity and low induration.

4.1.8.3 MAIN AEOLIAN UNIT INTERNAL GEOMETRY

Individual bed-sets of the Main Aeolian Unit are in average 20 m thick. Maximum thicknesses of up to 50 m have been recorded in the basin centre, in contrast to a few metres at the basin margins. The planar-tabular foresets consist of grainflow and grainfall cross-strata with wind ripples preserved in the basal part of the beds. A reconstruction of the palaeo-wind direction was deduced from the dip-directions of the dune foresets. They give an uniform wind direction from SW to WSW (Fig. 4.1-8). Trough cross-beds are 200-500 m wide, when measured perpendicular to the palaeo-wind. In sections parallel to the palaeo-wind direction the cross-beds appear smaller and are crescentic in shape. Some beds are followed by superimposed smaller bed-sets. Bounding surfaces of larger extent subdivide this unit vertically. In the coastal region the deposits have reduced dimensions, also palaeo-wind directions are more variable.
At the top, the Main Aeolian Units is at many places capped by the basal Etendeka lavas (Fig. 4.1-1).

4.1.8.4 MAIN AEOLIAN UNIT FACIES

The large scale cross-stratification, grainflow and grainfall lamination and the well-sorted and well-rounded grains are typical for purely aeolian dune deposits.
The large dimensions of the stratification perpendicular to the palaeo-wind and the uniform palaeo-wind directions suggest large transverse draa dunes. Transverse dunes reflect a constant wind direction, a high sand input and a high growth rate in accommodation space (Wasson & Hyde, 1983).
Thus, the Main Aeolian Unit represents a dramatic increase in the provision of accommodation space in the central area and high extrabasinal sand supply.
In the coastal region the restricted occurrences of the aeolian deposits reflect no pronounced growth in accommodation space. The corresponding aeolianites show smaller bed-sets and a higher variety in foreset orientation, which are attributed to partly isolated barchan dune forms. An environment of high wind-energy and strong deflation with fast migrating barchans is therefore assumed for the coastal area.
The maturity of the sandstones indicates a distal source in the west. Probably the Huab erg was fed from the time-equivalent Botucatu erg in Brazil, which probably had been connected directly to the erg system in the Huab area.

4.1.9 UPPER AEOLIAN UNIT

The Upper Aeolian Unit represents a period of time when aeolian deposition was punctuated by the onset of flood volcanism that led to the creation of the Paraná-Etendeka Flood Basalt Province. The active aeolian sand sea (erg) system was progressively covered by the basalts of the Awahab Formation (chapter 4.2.1) resulting in the preservation of large parts of the active dune system (Mountney et al., 1998). Furthermore, a variety of aeolian sediment layers interleaving basaltic flow units are preserved. They document an intense interaction between lava flows and aeolian deposition (Jerram et al., 1999a).

4.1.9.1 UPPER AEOLIAN UNIT LITHOLOGY AND FACIES

The Upper Aeolian Unit is divided from the Main Aeolian Unit in two ways: Either the boundary between the two units is marked by the first intercalated lavas of the Awahab Formation (cf. Fig. 4.1.-1) or by a laterally extensive bounding surface.

The lithological composition is identical with that of the Main Aeolian Unit. The geometric dimension decrease due to the progressive shut down of sand supply by the successively covering of the erg system by the Etendeka flood basalts: The Upper Unit is up to 100 m thick and individual beds vary in thickness from 5 m to 100 m. In the basin centre completely preserved transverse draa bedforms with wavelengths of 1 km and amplitudes around 100 m were observed. Towards the basin margins the aeolian bedforms are smaller with amplitudes rarely exceeding 10 m.

Foreset dip orientations indicate palaeo-winds from north-westerly to south-westerly directions. They are more variable than in the Main Aeolian Unit, and also a switch to a dominantly north-westerly wind-direction is recognised (Fig. 4.1-8).

Fig. 4.1-8

Palaeo-wind analysis of the Main and Upper Aeolian Unit of the Twyfelfontein Formation in the central and eastern Huab area. The Main Aeolian Unit reveals palaeo-winds from south-westerly directions, whereas the Upper Aeolian Unit reveals more variable directions with a switch to dominantly north-westerly palaeo-winds.

4.1.9.2 BASALT-SEDIMENT INTERACTIONS OF THE UPPER AEOLIAN UNIT

Near the top of the Upper Aeolian Unit the lava sediment interactions caused 3 distinct types of geometries which document the successive shut down of the aeolian system (Jerram et al, 1999a; Fig. 4.1-9): (1) Laterally amalgamated multi-dune complexes, that are up to 20 m thick and 1 km wide. (2) Individual small barchans that are draped by lava flows, which have flooded the depositional surface. Further up (3) lens shaped sandfills of lava-topography characterised by a convex basis and planar top, (4) lava-sandstone breccias and (5) sand bypass-surfaces indicated by sand-filled fissures in the lavas.

These geometries are evidence for an active aeolian system becoming successively covered by a developing lava field. The successive sealing of the sand surfaces by lavas led to a decrease of the sand supply that consequently caused the disappearance of the aeolian system. Furthermore, the lavas consumed accommodation space by draping topography, and finally deflation prevailed instead of sedimentation.

Fig. 4.1-9

Schematic section of the stratigraphy of the lower Etendeka lavas and their interfingering with the aeolian sandstones of the Twyfelfontein Formation (slightly modified from Jerram et al., 1999a).

4.1.10 TIME CONSTRAINTS FOR THE TWYFELFONTEIN FORMATION

The pure aeolian units of the Twyfelfontein Formation interfinger with the Lower Etendeka lavas and therefore they can be dated by the lavas. The latter have been dated by measuring the $^{40}Ar/^{39}Ar$ isotopic composition of feldspar at 132±0.7 Ma (Renne, written comm. 1997; chapter 4.2.6). Volcanics overlying the aeolianites give similar ages from 132.3±0,7 to 131.7±0.7 Ma (Renne, 1996; chapter 4.2.6). Dating the base of the Twyfelfontein Formation including the Krone Member is a bit more problematic, because earlier authors voiced the probability of an erg system starting in the Triassic-Jurassic and extending into the Cretaceous (Martin, 1982; Ledendecker, 1992). But, Permian Karoo clasts and silicified wood and bones deriving from reworked Permian Karoo sediments indicate that the Krone Member is substantial younger than Permian. A few volcanic clasts found in the Krone Member confirm this age estimation: Their extreme scarcity points to a source different from the Etendeka lavas. An origin from metamorphosed basement rocks is excluded due to the lack of deformation features. Therefore it is suggested, that the volcanic clast found in the Krone Member originate from Karoo-aged volcanics such as the volcanic material found in the Permian Verbrandeberg Formation of the eastern Huab area (chapter 2.2.2.1).

More crucial is a recent find of a tetrapod bonebed in the Krone Member, that gives a Jurassic maximum age (Löffler & Porada, 1998). The minimum age of the Krone Member is given by the first Lower Cretaceous Etendeka Group lavas. As volcanic Etendeka Group clasts are probably not present, the minimum age of the Krone Member should be slightly higher than the first Etendeka Group lavas.

4.1.11 DURATION OF TWYFELFONTEIN FORMATION DEPOSITION

Because of the lack of fossils it is difficult to assess the duration of the development from the fluvial conditions of the Krone Member via a mixed fluvial-aeolian transition to the pure aeolianites, that are of the same age as the first Etendeka lavas. But, when considering the facies architecture of the Twyfelfontein Formation, an estimation of its duration is possible: The Twyfelfontein aridification succession begins with the alluvial-fluvial, semi-arid facies of the Krone Member. The rudaceous character of the conglomerate indicates short transports driven by high energy flashfloods with almost no multiple reworking. Therefore short episodes of deposition are implied. The following Mixed Aeolian-Fluvial Unit represents a similar facies and it records a rapid transition to the succeeding Main Aeolian Unit. The relative structural and compositional immaturity of the partly fluvial units as well as the consistent south-westerly wind-directions in the Main Aeolian Unit, suggest a rather quick deposition of the whole sedimentary sequence. Indicators for significant hiatal gaps like well developed palaeosols, were not observed. Such a short sedimentation interval is consistent with the negligible age variations of lavas interfingering with the Upper Aeolian Unit of the Twyfelfontein Formation (chapter 4.2.7).

As will be outlined below, the entire Twyfelfontein Formation has been affected by enhanced tectonism, that caused both, brittle and plastic deformation. The scale and the distribution of soft-sediment deformation features strongly affirm a quick sedimentation: At places soft sediment faulting can be traced from the basal Krone Member up into the Main Aeolian Unit. This means that no complete lithification took place from the deposition of the Krone Member via the transitional Mixed Aeolian-Fluvial Unit until the deposition of the basal Aeolian Unit. It appears unlikely, that theses sediments stayed for many million years unconsolidated and therefore a rather short time-span for the deposition of the Krone Member and the following units is assumed.

4.1.12 TIME CONSTRAINTS FOR THE COASTAL ETENDEKA GROUP SEDIMENTARY ROCKS

4.1.12.1 ALBIN CONGLOMERATE

The time relation of the Albin Conglomerate to the Krone Member of the Huab area is problematic: The facies architecture inclusive the transitional change to a mixed fluvial-aeolian unit is similar for both, but the relatively high abundance of basaltic clasts in the Albin Conglomerate points to a syn-volcanic deposition. The basaltic clasts are often amygdaloidal and vary in abundance from almost 0% to 10% of the clast population. Petrographically, they closely match basalt flows in the Tafelkop Member of the lower Awahab Formation (Swart, 1992/93). Therefore Swart (1992/93) assumed deposition during the early stages of Etendeka volcanism. More recent analysis of these basaltic clasts, however, resulted in a different composition than any other currently known basaltic rock type of the Etendeka (Marsh, pers. comm. 1998). Their trace element ratios display a strong affinity to OIB, similar to the Kudu basalts, which are located offshore southern Namibia (Marsh, pers. comm. 1998) (for location of the Kudu wells see Fig. 1.2-4). Consequently it is not clear whether the basaltic clasts represent an unpreserved phase of Etendeka volcanism or whether they are derived from an older, different volcanic unit. Therefore the remaining time constraints for the Albin Conglomerate are its facies architecture and its stratigraphic relationships. As both are rather comparable to those of the Krone Member, it is suggested that Albin Conglomerate and Krone Member are almost stratigraphically equivalent.

4.1.12.2 SANINAB CONGLOMERATES AND COASTAL PEBBLY SANDSTONES

The lithologies of the conglomerates and pebbly sandstones at the Sanianab outcrop and those of the coastal sandstones north of Terrace Bay are rather similar. Both are immature and contain the same type of gneiss clasts. Furthermore they both are succeeded by Twyfelfontein aeolianites. It is therefore an admissible assumption that they were deposited contemporaneously. As outlined in chapter 4.1.6.2 a Krone Member age is most probable.

4.1.12.3 KHARU-GAISEB CONGLOMERATE

The monomict Kharu-Gaiseb Conglomerate consists entirely of volcanic material from the Etendeka Group. It is deposited in a hangingwall position to an adjacent fault. The footwall rock type beneath the conglomerate is unknown (Ward & Martin, 1987). Furthermore it is not covered by any other rock type, which makes more distinct stratigraphical relations impossible. However, the monomict character of the conglomerate confirms its deposition after the extensive onset of flood basalt effusion. Furthermore the absence of any sand in the matrix supports a younger age than the whole Twyfelfontein Formation: The next exposures of the Twyfelfontein aeolianites are 10 km west of the conglomerate outcrop. If the Twyfelfontein erg had been active, sand should have infiltrated into the conglomerate.

The rudaceous character of the deposit deduces a high sedimentation rate and consequently a short sedimentation interval.

It should be noted that the conglomerate exposure has not been visited by the author. The interpretations given here are based on the observations reported in Ward & Martin (1987).

4.2 THE ETENDEKA GROUP IGNEOUS ROCKS

The Etendeka Group igneous rocks comprise all Lower Cretaceous extrusives found in north-western Namibia. They are well correlated with their Brazilian counterparts in the Paraná flood basalt province, as documented by the many stratigraphical, geochemical and geochronological conformities of the Etendeka volcanics to those of the Brazilian Serra Geral Formation (Hawkesworth et al., 1992; Milner et al., 1995a). They comprise a bimodal association of tholeiitic basalts/basaltic andesites (51-59 % SiO_2) and quartz latites (66-69 % SiO_2). The basalts form tabular lavas, dykes and sills. The quartz latites are interpreted as rheoignimbrites (Milner et al., 1992). Most of the intrusive bodies are presumably the subvolcanic roots of caldera structures.

For the southern Etendeka region Milner et al. (1994) subdivide the volcanics into the Awahab Formation and the unconformably overlying or onlapping Tafelberg Formation.

In the Huab Basin the Etendeka volcanics are only represented by the Awahab Formation and units of the Tafelberg Formation onlapping in the area north of the Huab River (Milner et al., 1992; 1995a).

4.2.1 THE AWAHAB FORMATION

The type locality for the Awahab Formation is Mt. Awahab (synonym for Mt. Mikberg) in the Huab Outliers.

The Awahab Formation comprises three main types of rocks: 1) olivine-phyric basalts with associated interleaved sediments of the Upper Aeolian Unit, 2) basalts/basaltic andesites, and 3) quartz latites - large-volume sheet-like silicic volcanics (termed rhyolites in the Paraná).

The Awahab Formation is subdivided into four Members (Milner et al., 1994) (Fig. 4.2-1):

1) The basal Tafelkop Member: Olivin-phyric basalts and aphyric basalts/basaltic andesites (Tafelberg-type).
2) The Goboboseb Quartz Latite Member: Quartz latites (units 1-3).
3) The Messum Mountain Member: basalts/basaltic andesites
4) The Springbok Member: Quartz latites

The above subdivision cannot be applied to the whole Etendeka region. For the Huab area, the Tafelkop Member only comprises olivin-phyric basalts, which are termed Tafelkop-Interdune Member by Jerram et al. (1999b) (see below). The succeeding aphyric basalts/basaltic andesites form the upper part of the Tafelkop Member in the Goboboseb area, whereas they are referred to the independent Tsuhasis Member in the Huab area (see below).

4.2.1.1 AWAHAB FORMATION OLIVIN-PHYRIC BASALTS:

The initial basalts poured out into the active erg system as passive, pahoehoe style low viscous lava flows. Evidence of the pahoehoe nature of these flow units is displayed by the compound, lobe and sheet, internal morphology and a variety of prints and moulds of ropy pahoehoe textures preserved in the thin aeolian sediment interlayers. Today, these lowermost olivine-phyric basalts are readily weathered and are generally poorly exposed. Good basalt samples can be found where the lava had ponded and degassed in inter-dune areas. Similar olivin-phyric basalts appear in the Goboboseb Mts. in a similar stratigraphic position (Ewart et al., 1998a; Milner et al., 1994). Geochemically the term "Tafelkop-type" was introduced by Milner & Le Roex (1996). Instead, Ewart et al. (1998a) use the term LTZ.H basalts due to their higher MgO, TiO_2, FeO* content, higher Ti/Zr ratios, and a wider range of Ti/Y ratios than the normal LTZ (low Ti & Zr) basalts of the Paraná-Etendeka Flood Basalt Province. In the Huab area these basalts are interbedded with the aeolian sandstone units and consequently Jerram et al. (1999b) introduced the term Tafelkop-Interdune Member basalts in order to ascribe olivin phyric basalts with Tafelkop-type geochemistry. The Tafelkop-Interdune Member basalts thicken towards the Huab Outliers area, with the units thinning to the north and to the west. At Mt. Tafelkop (northern Goboboseb Mts.) this member attains about 220 m in thickness.

4.2.1.2 AWAHAB FORMATION APHYRIC BASALTS AND BASALTIC ANDESITES:

Voluminous laterally extensive basalts/basaltic andesites flows onlap the olivin-phyric lavas (Tafelkop-Interdune Member) and smoothen the aeolian deposits (Twyfelfontein Formation) to the west. Due to this onlapping an unconformity on top of the Tafelkop-Interdune Member is implied (Unconformity I, Jerram et al., 1999b). Below the unconformity there appear a few individual flow units of basalts/basaltic andesites, which are similar to those above the unconformity. The basalts/basaltic andesites below the unconformity are interleaved between Tafelkop-Interdune Member basalts.

Most basalts/basaltic andesites form well exposed outcrops and consist of fine grained aphyric flows, with a massive lower and middle part (2/3) of the flow and an amygdaloidal top. Individual flow units are up to 60 m thick.

The basalts/basaltic andesites are geochemically equivalent to the common LTZ type basalts of the Paraná-Etendeka Flood Basalt Province (termed LTZ.L by Ewart et al., 1998a), viz. the Tafelberg-type basalts of the Etendeka (Milner et al., 1994) and the Gramado and Esmeralda basalts of the Paraná (Peate et al., 1992). Jerram et al. (1999b) introduced the term "Tsuhasis Member" for the basalts/basaltic andesites of the Mt. Awahab (Mt. Mikberg) area, because they do not stratigraphically correlate with the geochemical identical Tafelberg-type basalts of the Tafelberg locality.

The onset of these laterally extensive, more voluminous, basaltic andesite flows appears to have markedly cut off the supply of sand to the basin resulting in the termination of sediment interbeds (Jerram et al., 1999a).

4.2.1.3 AWAHAB FORMATION QUARTZ LATITES:

The upper part of the Awahab Formation contains the first occurrences of the large volume silicic volcanic units, the Goboboseb Quartz Latites (Unit I, II & III) and the Springbok Quartz Latite Member (Fig. 4.2-1). The Goboboseb quartz latites are of great interest as they have been geochemically linked with the Messum Complex to the south of the Huab Outliers (Milner & Ewart, 1989; Milner et al., 1992). At present these remain the only large volume silicic flow units in the Paraná-Etendeka Flood Basalt Province which have been linked to their eruptive centre. The Goboboseb quartz latites have been correlated into the Paraná Basin indicating the dramatic lateral extent (approximately 350 km from the Messum Complex) and aerial coverage (33 000 km^2) of such flows (Milner et al., 1995a). Within the Huab Basin the Goboboseb Unit I Quartz Latite overlies the Tsuhasis Member basalts in the Huab River area and north-western Huab Outliers, whereas in the very south-eastern part of the Huab Outliers the Goboboseb Unit I

Quartz Latite lies directly on Tafelkop-Interdune Member basalts. The Springbok Quartz Latite caps the succession and is 150 m thick at Mt. Awahab with the top of the flow unit missing.

In summary, the Awahab Formation in the Huab Basin consists of the Tafelkop-Interdune Member basalts at its base and interleaved aeolian sediments of the Twyfelfontein Formation (Upper Aeolian Unit). These are followed up-section by the Tsuhasis Member basalts/basaltic andesites which are mainly concentrated in the Huab River area. Two major quartz latite units, the Goboboseb (I & II) and the Springbok Quartz Latite Member (Milner & Ewart, 1989) form the upper parts of the succession interleaved with the Tsuhasis Member basalts/basaltic andesites.

In the Goboboseb area all the volcanic rocks are part of the Awahab Formation. The basal, up to 300 m thick Tafelkop Member consists of both, olivin-phyric basalts (Tafelkop-Interdune Member) and aphyric basalts/basaltic andesites. The Tafelkop Member is overlain by three quartz latite units (Units I, II & III *sensu* Milner & Ewart, 1989), which constitute the Goboboseb Quartz Latite Member.

4.2.2 THE TAFELBERG FORMATION

Type locality for the Tafelberg Formation is Mt. Tafelberg, the highest table mountain located at the eastern margin of the Etendeka Plateau north of the Huab River (for location see Fig. 1-3). At this locality the volcanics attain their maximum thickness of 900 m.

The Tafelberg Formation follows disconformably above the Awahab Formation (Milner, 1988; Milner et al., 1994). At Tafelberg type locality an angular unconformity separates Tafelberg basalts (chapter 5.2.3.2) of the Tafelberg Formation from Tsuhasis Member basalts of the Awahab Formation (Unconformity II, Jerram et al., 1999b). It is believed, that this unconformity is part of a more extensive regional disconformity (Milner, 1988; Milner et al., 1994) that can be traced into the Paraná Basin (Milner et al., 1995a). However, it should be noted that the degree of down-cutting on this disconformity in the Tafelberg type locality and its full regional context is not fully constrained.
The Tafelberg Formation consists of two main volcanic rock types: (1) basalts/basaltic andesites, and (2) quartz latites.

4.2.2.1 TAFELBERG FORMATION BASALTS AND BASALTIC ANDESITES:

The Tafelberg Formation basalts and basaltic andesites display the same geochemistry as the Tsuhasis Member basalts of the Awahab Formation (Tafelberg-type or the LTZ.L of Ewart et al., 1998a). But due to their similar field appearance they can only be separated from Tsuhasis Member basalts on the basis of the stratigraphic context. Under this definition no Tafelkop-type basalts (*sensu* Milner et al., 1994) appear in the Tafelberg Formation.

4.2.2.2 TAFELBERG FORMATION QUARTZ LATITES:

The Tafelberg Formation contains the Wêreldsend, Bergsig, Grootberg and Beacon Quartz Latite members (Fig. 4.2-1), with the Bergsig Member missing at the type section. A latite unit, whose lateral extent is limited against the disconformity, occurs near the base of the formation and is informally termed the Tafelberg Latite (Jerram et al., 1999b).

4.2.3 THE CURRENT STRATIGRAPHIC PICTURE OF THE HUAB AREA

The basal Awahab Formation comprises Tafelkop Member basalts interleaved with the aeolian units of the Cretaceous Twyfelfontein Formation. These units are onlapped by the basalts/basaltic andesites indicated by disconformity I. The highest preserved unit of the Awahab Formation is the Springbok Quartz Latite Member. A major disconformity then occurs, disconformity II, which separates the Awahab Formation from the Tafelberg Formation. The highest preserved unit in the Tafelberg Formation is the Beacon Quartz Latite Member.

How this stratigraphic picture presented for the Huab Basin relates to other stratigraphic sections within the Etendeka is problematic. Within the Goboboseb Mts., 50 km to the south, Tafelkop-type basalts are interleaved with Tafelberg-type basalts. It is possible that the interleaved Tafelberg-type basalts erupted locally around the Goboboseb Mts. and are not present in the Huab Basin. Also it is possible that the upper Tafelkop-type basalts, that lie above the Tafelberg-type in the Tafelkop Section, represent locally erupted flows that are not present in the Huab Basin. However, the lack of detailed lithostratigraphic correlation of the Tafelkop basalts within the Goboboseb Mts. and the missing stratigraphic link between the Goboboseb Mts. and the Huab Outliers, complicates the solution.

4.2.4 THE MAIN LAVA FIELD

The main lava field is located north of the Huab River and covers most of the Etendeka region. It is structurally divided by the Ambrosiusberg Fault zone into an eastern and a western zone. The western (coastal) region is characterised by a westwardly increasing easterly dip and down-faulting of block-segments. East of the Ambrosiusberg Fault zone the volcanics are essentially flat lying with only minor block-rotations and down-faulting.

The overwhelming amount of volcanics covering the area north of the Huab River relate to the Tafelberg Formation. Only the southern part, exclusive the faulted coastal region west of the Ambrosiusberg Fault zone, is dominated by the Springbok Quartz Latite Member of the Awahab Formation. The western (coastal) facies of the Tafelberg Formation (west of the Ambrosiusberg Fault zone) comprises an approximately 1000 m thick sequence of up to 20 quartz latite units, which correspond geochemically to the Wêreldsend, Grootberg and Beacon Quartz Latite members of the Tafelberg type section (Milner & Duncan, 1987; Milner et al., 1994).

North of Terrace Bay a new type of basalts occurs, which are termed Khumib-type basalts. It becomes the dominant basalt compositional type in the northern Etendeka region. The Khumib-type basalt belongs to the HTZ-type (high-Ti and high Zr) of Marsh (1987).

4.2.5 THE ALBIN RIDGE VOLCANICS

The Albin Ridge is located 15 km east of the Atlantic coast (Fig. 1-3). It is running parallel to the present day coastline (NNW-SSE). It comprises Etendeka basalts and a quartz latite unit, overlying a thin sequence of conglomeratic sediments (chapter 4.1.6.1).

The basaltic units at the base of the volcanic succession are characterised by their high CO_2 content and their plagioclase-phyric nature. Within the Etendeka Group they represent an unique type of basalts, which have been termed Albin-type basalts by Erlank et al. (1984). After Milner et al. (1994) they are part of the Tafelkop Member within the Awahab Formation of the Etendeka Group. The Albin-type basalts appear to be restricted to the coastal sections of the Etendeka, but stratigraphic relations are often complicated by faulting and poor exposure. Although these basalts occur invariably at the base of the succession (Erlank et al., 1984), and are stratigraphically beneath the Goboboseb quartz latites, their relation within the Etendeka context is complex with flows of the Albin-type occurring as far north as Terrace Bay. Yet no Albin-type basalts have been found in the Huab area.

At Albin Ridge the Albin-type basalts are overlain by a series of aphyric Tafelberg-type basalts (both Tafelkop Member of the Awahab Formation), the latter are capped by a quartz latite unit. This unit is compositionally correlated with the Goboboseb Quartz Latite Member of the Awahab Formation.

4.2.6 AGE CONSTRAINTS FOR THE PARANÁ-ETENDEKA LAVAS

The first radiometric dates of volcanics from the Etendeka province have been presented by Siedner & Mitchell (1975) who calculated K/Ar ages between 110 and 140 Ma. Later, Blümel et al. (1979) obtained K/Ar dates from basalts of the Erongo area, that might be related to Etendeka volcanism. Their calculated ages varied between 161 and 137 Ma. Other K/Ar dates of the Paraná-Etendeka lavas show a similar wide range from 166 to 100 Ma (Erlank et al., 1984; Rocha-Campos et al., 1988). Recent datings provide a narrow age range with dates focusing on the interval from 130 to 135 Ma (Milner et al., 1992, 1995b; Peate et al., 1992, Renne et al., 1996).

$^{40}Ar/^{39}Ar$ dates of the Etendeka province vary between 131.7±0.7 Ma and 132.3±0.7 Ma. Most of these dates are related to equivalents of the stratigraphic level above the olivin-phyric basalts of the Awahab Formation. Only one sample derived from a lower stratigraphic level. This sample is dated at 132±1Ma (Renne, written comm. 1997) and was collected in the Huab area.

4.2.7 DURATION OF VOLCANISM

The narrow range of the $^{40}Ar/^{39}Ar$ dates for the Etendeka basalts and quartz latites suggest a rather short time-span for their effusion.

Magnetostratigraphic studies (Glen et al., 1997; Renne et al., 1996) recognise a single reversed magnetic polarity interval in both, the Awahab Formation and the Tafelberg Section, which may or may not be equivalent. This confirms a short time-span, especially when considering the high frequency of polarity changes during the Early Cretaceous (cf. Gradstein et al., 1994).

Fig. 4.2-1
Stratigraphy of the Etendeka Group (slightly modified from Milner et al., 1994).

4.2.8 MESOZOIC DYKES AND SILLS

4.2.8.1 THE HORINGBAAI DOLERITES

Geometry

Erlank et al. (1984) report a swarm of thin (< 1 m) dolerite dykes and sills in the area north of Cape Cross (Fig. 1-3). They termed this dykes and sills the Horingbaai dolerites. Due to a map in Erlank et. al (1984) the predominant strike of the Horingbaai dykes is NW-SE. Subordinate strike maxima are N-S and NE-SW. The NW-SE striking dykes are the thickest within the dyke population. Their NW-SE trend conforms the dominant fault trend in the area. A sinistral oblique slip component associated with the NE-SW striking dykes is expressed by the sinistral offset of some NW-SE striking dyke segments.

Field observations in behalf of this study document, that the dykes and sills intrude the Albin- and Tafelberg-type basalts at the Albin Ridge as well as the underlying Karoo sediments and Damara rocks.

Composition

In terms of normative classification of basaltic rocks, the Horingbaai dolerites are transitional in composition between alkali basalts (ne-normative) and olivin tholeiites (ol- and hy-normative) (Erlank et al., 1984).

Their chemical and isotopic composition shows pronounced differences to all other mafic Etendeka rocks: The TiO_2 *vs*. Zr trends and Ti/Th ratios are clearly different from Albin- and Tafelberg-type basalts (Erlank et al., 1984). Also the εNd values are significant higher and the εSr values are lower than in the Albin- and Tafelberg-type basalts (Erlank et al., 1984).

Le Roex et al., (1983) explain the high εNd and low εSr values, the initial $^{87}Sr/^{86}Sr$ ratios (0,703-0,704) and the REE patterns of the Horingbaai dolerites with a depleted magma source. Furthermore, they emphasise that this explanation is consistent with the geochemically strong affinity to T-MORB from the South Atlantic (Le Roex et al., 1983).

Age constraints and age relations

Erlank et al. (1984) dated two Horingbaai dolerites at 125 Ma and 125-130 Ma by using the $^{40}Ar/^{39}Ar$ incremental step heating technique on plagioclase separates. These ages are substantially younger than those of the Etendeka volcanic rocks, dated by the same technique at 132 Ma (Renne et al.,1996). This age relation is consistent with the observation, that the Horingbaai dolerites intrude the basalts at Albin Ridge.

Milner (1992) emphasises that the Horingbaai dolerites have essentially the same Barrêmian age as estimates for the initial spreading of the South Atlantic at the same latitude (Austin & Uchupi, 1982) (chapter 6). Therefore the intrusion of the Horingbaai dolerites post-dates the Etendeka event and is directly linked to the initial formation of oceanic crust.

4.2.8.2 DYKES IN THE HUAB AREA

Geometry

Dolerite dykes can be found in the whole Huab area. The thickness of the dykes varies between 50 cm and 10 m. Many dykes swell and pinch out along strike and have therefore a feather-like appearance on a regional scale.

Several dykes can be followed over long distances of more than 30 km. East of the Wêreldsend Fault system most dykes have a consistent N-S trend (Plate VI-2 and VI-4). There the dykes usually intrude faults, which have a longer lateral extent than the dykes themselves. Intrusion took place either syn- or post-kinematic, because no clear fault induced offsets of the dykes has been observed. West of the Wêreldsend Fault system strike maximum is NW-SE, beside N-S and SW-NE dyke trends (Fig. 5.3-11, Fig. 5.1-4). There the majority of the dykes do not seem to have intruded older faults.

Most dykes on either side of the Wêreldsend Fault have a vertical or steep attitude, only a few are shallow dipping, or are part of sill-dyke-complexes (Plate VI-3).

Fig. 4.2-2
N-S striking dyke comprising amydaloid basalt. The dyke can be traced as far south as Doros Crater, which is at a distance of more than 10 km. The dyke acted as a feederdyke for the lowermost Awahab Formation which is exposed 50 m north of the photograph location. Location of the outcrop is 10 km north of Doros Crater at the south-western end of a SW-NE trending incised valley (Rhino Section; S20°40´39´´/E14°11´20´´). Compare with Plate VI-2.

Timing of dyking

The dykes display the same composition as the Etendeka lavas and therefore Duncan & Armstrong (1990a) assumed that they acted as their feeder dykes. Field relations constrain this assumption as many dykes are undoubtedly the eroded conduits of the basalts that inundated the area. North of Doros Crater (Rhino Section, Fig. 2.4-4; Plate VI-2), for example, a 10 m wide dyke refers as a feeder dyke for the lowermost olivin-phyric lava flow (Tafelkop-Interdune Member) covering the Twyfelfontein Formation aeolianites (Main Aeolian Unit) (Fig. 4.2-2). This dyke can be traced from Doros Crater 9 km northwards to the Karoo escarpment, clearly cutting through Damara basement rocks, Karoo rocks and the Main Aeolian Unit of the Twyfelfontein Formation.
Most dykes, however, terminate in an undetermined position somewhere in the volcanic pile. It is not clear, whether they fed individual lava flows or post-date them.
Some dykes cut clearly through the Etendeka volcanics and clearly post-date the preserved Etendeka lavas/rheoignimbrites.
Towards the coast, late dyking becomes more conspicuous: At the Uniab Fault and west of it a few NW-SE trending dykes cut clearly without offset through post Etendeka faults (Fig. 5.3-11). These faults have clearly down-faulted and tilted the Etendeka Group lavas (Poiki-locality, Fig. 5.3-6), which confirms the late- or post-volcanic timing of dyke implacement.

4.2.8.3 THE HUAB SILLS

Geometry

The occurrence of a major dolerite sill in the Huab area was noted by Hodgson & Botha (1975). Botha & Hodgson (1976) suggested that the dolerite in the Huab valley was a large sill with an outcrop area of at least 400 km^2 and a thickness in the order of up to 100 m. A detailed study by Duncan et al. (1989) has shown a number of sills intruded approximately in the same stratigraphic level, rather than a single sill. The sills have an outcrop area in excess of 600 km^2 (Fig. 2.1-4) and individual sills are up to 130 m thick. They intrude the Damara basement, the overlying Karoo sediments and the sandstones and conglomerates of the Twyfelfontein Formation. The major intrusion levels are between the Damara basement and the overlying Karoo sediments and within the pelites of the Verbrandeberg Formation *sensu* Horsthemke (1992).

Composition

All the dolerite types are tholeiitic, containing both normative hypersthene and olivine (Duncan et al., 1989). Due to the chemical analysis of Duncan et al. (1989) the magnesium content and the Ti/Zr ratios of the sills are equivalent to the Tafelkop-type basalts *sensu* Milner & Le Roex (1996) or LTZ.H basalts *sensu* Ewart et al. (1998a).
Lava equivalents in the Huab valley area could be the olivin-phyric highly weathered basalts at the base of the Awahab Formation, which display the same geochemistry (Duncan and Milner, pers. comm., 1997).

Intrusion levels

An extremely shallow level of intrusion would be implied, if the Huab sills and the olivine-phyric lavas were coeval and if they formed almost contemporaneous. In the central Huab River valley, for example, a thick (> 70m) sill intruded directly beneath the Krone Member of the Twyfelfontein Formation. The strata that formerly covered the sill during its emplacement likely comprised the entire Twyfelfontein Formation and the olivin-phyric lavas of the Awahab Formation. Presumably no stratigraphically higher lavas had been deposited at that time. The constructed thickness of these strata measures around 100-200 m, suggesting a rather shallow intrusion level of the sill.

Age relations

The sills are most probably very similar in age to the basal Etendeka lavas, and their close age relation to the Etendeka Group sediments (Twyfelfontein Formation) is confirmed from field relations:
The Krone Member suffered contact metamorphism near the sill's surface. This proves that the sill intruded below the basal Krone Member, rather than deposition of the Krone Member onto an exhumed sill surface. At places the bounding surface of the Krone Member towards the sill displays plastic deformation features, which indicates that the Krone Member was not completely consolidated during sill implacement. Therefore a narrow time span from deposition of the Krone Member until the intrusion of the sill is suggested. Furthermore, this model implies a rather short sedimentation period of the whole Twyfelfontein Formation, because the olivin-phyric lavas on top of the sedimentary succession are considered to be approximately of the same age as the dolerite intrusions (see above).

4.2.9 DAMARALAND INTRUSIVE COMPLEXES

A prominent north-east trending line of Mesozoic, anorogenic, subvolcanic complexes extends across Damaraland in Namibia from the coast at Cape Cross to Okorusu, 350 km inland. Most conspicuous are the Spitzkoppe-, Erongo-, Messum- and Brandberg complexes (Fig. 1-3) as they form well defined mountainous complexes in a relatively plane area of low relief. All complexes are high-level subvolcanic intrusions. Many of them have preserved extrusive remnants and are inferred to be caldera-collapse structures. The various intrusive components of the Damaraland Complexes range in age from 124 Ma to 137 Ma (Milner et al., 1995b; Renne et al., 1996, Armstrong et al., 1998). This narrow range confirms the essential synchronicity of the subvolcanic complexes, contemporaneous, or at least temporarily associated with the Etendeka flood basalts.

A short description of the Brandberg, Messum and Doros Intrusive Complexes is given here, because they are prominent geological features of the study area. Furthermore their formation is temporarily and genetically linked with the volcano-tectonic evolution of north-western Namibia (chapter 6).

4.2.9.1 BRANDBERG

Mt. Brandberg is a huge intrusion of alkali granite 30 km across and 2573 m high. Remnants of Karoo sediments overlain by Etendeka basalt and quartz latite form a dark-coloured rim around the granite. The complex exhibits a number of subvolcanic magmatic centres which are entirely granitic in composition. Studies of the Brandberg Complex indicate that it represents an anorogenic ring complex of intra plate emplacement at high level in the crust during a period between 125 and 135 Ma (Watkins et al., 1994). U/Pb SHRIMP dates were obtained from a alkali ring dyke at Mt. Brandberg which yield ages at 135±1 Ma (Armstrong et al., 1998). However, the main Brandberg granites must be younger, because they intrude the Etendeka lavas, which display an age range between 131.7±0.7 Ma and 132.3±0.7 Ma ($^{40}Ar/^{39}Ar$ dates) (chapter 4.2.6).

4.2.9.2 MESSUM CRATER

The Messum Crater is a very complex concentric structure. It is roughly circular and plane and is approximately 18 km in diameter. Its rock spectrum includes gabbroids, granitoids, syenites and mafic and acid volcanics (Milner & Ewart, 1989). The contact to the adjacent Etendeka volcanic rocks is clearly faulted and the Etendeka Group rocks are contact metamorphosed by the complex. These contact relationships indicate that the final structural movements adjacent to the complex and its final thermal imprint clearly postdate the eruption of the Etendeka Group basalts and quartz latites (Milner & Ewart, 1989). Most rocks found at Messum Crater cannot be directly correlated with the adjacent volcanic rocks of the Goboboseb Mts. Milner & Ewart (1989), however, found strong evidence for the quartz latites of the Goboboseb Mts. (Goboboseb Member of the Awahab Formation) being erupted from Messum Crater itself or a nearby predecessor structure.
This link is of special interest, because the Goboboseb Member has been correlated with the Palmas Acid volcanics of the south-eastern Paraná region (Milner et al., 1995a), which gives an estimation of lateral extent (> 350 km), areal coverage (> 33 000 km²) and volume of these quartz latites.

4.2.9.3 DOROS CRATER

Doros Crater is a gabbro intrusive complex located between Mt. Brandberg and the Huab Outliers. According to the field observations during this study, the intrusive geometry of Doros Crater is best described by a lakkolith-ring-dyke complex: The centre consists mainly of gabbro, the rim of steep dolerite inclined towards the centre. The dolerite could be a finer grained, coeval equivalent to the gabbro in the centre, or it also might represent an independent intrusion.
Several dolerite dykes seem to radiate from the structure. But it has not yet been solved, whether the gabbro/dolerite was a magma chamber feeding the dykes.
Reuning & Martin (1957) and Jerram et al. (1999b) speculated that the Doros Complex was part of the conduit system of a volcano. The geometry of the lower Awahab Formation (olivin-phyric basalts) fits well with that of a shield volcano with its centre at Doros Crater (chapter 5.2.3.1).

At the northern rim of Doros Crater remnants of Karoo sediments are preserved, due to their induration by contact metamorphism. These are the only outcrops of Karoo sediments between the Huab Outliers and Mt. Brandberg.

Chapter 5: Etendeka Group Disconformities and associated tectonic Features

5.1 THE KAROO/ETENDEKA UNCONFORMITY IN THE HUAB-ETENDEKA REGION

The Karoo/Etendeka unconformity is well developed particularly in the Huab area. Outside the Huab area it is exposed at Albin Ridge (Section 31, Fig. 4.1-4), Sanianab Locality (Fig. 5.3-2) and in the Goboboseb Mts. The Etendeka Group follows with an angular unconformity usually on Karoo strata, only north of Krone 721 Farm it rests directly on Damara basement. In a few places north of the Huab River and at Mt. Bruin the Twyfelfontein Formation is primarily missing and Etendeka lavas rest directly on Karoo rocks. The same situation occurs in most places of the Goboboseb Mts. and in the Mt. Brandberg area. Only at the south-western slope of Mt. Brandberg, relictic deposits of the Twyfelfontein Formation have been observed (chapter 3.3.2.1; Fig. 3.1-4).

A reconstruction of the present topography of this pre-Cretaceous unconformity reflects the tectonic zonation by several N-S trending fault-systems as shown in Figure 5.1-1. Along a W-E traverse four north-south trending fault zones cause a stepwise relief of the unconformity-surface: In the coastal area the unconformity lies close to the present see level, whereas at Twyfelfontein it is approximately 700 m above. This increase in altitude along a section from the coast to the east roughly coincides with the present day surface-topography. It is here assumed that the pre-Twyfelfontein Formation relief was roughly inverse to the recent unconformity-topography (see later this chapter), but with the same four fault systems causing a pronounced tectonic and palaeo-topographic zonation.

The anatomy of the pre-Etendeka Group unconformity shows considerable variations and can therefore be subdivided into four groups:

1) The Krone Member conglomerate lies sheet-like on Karoo rocks.
2) Krone Member conglomerates erode in several meters wide channels into the underlying Karoo strata. The erosional boundary is uneven or wavy, pedogenic features are missing (Fig. 5.1-2, Plate III-2).
3) A plane palaeo-deflationsurface consisting of a single layer of polished quartz pebbles defines the unconformity. Dune deposits follow usually without fluvial intercalations.
4) Aeolianites or basal Etendeka lavas rest directly on Karoo or Damara basement rocks. The bounding surface is plane.

Type 1 and 2 are typical in the area extending from the outcrops north of Doros Crater to those immediately north of the Huab River. A local but thick deposit of these types is also present in the Bloukrans Graben. Type 3 is restricted to Klein Gai-As, Gai-As and around the boundary to the Skeleton Coast Park. This single pebble layer (type 3) is interpreted as the southern- and westernmost extension of the Krone Member. Type 4 is most abundant. Places where aeolianites rest directly on Damara basement are north-east of Krone 721 Farm and north of Terrace Bay.

Pedogenic features associated with the pre-Etendeka unconformity have been observed north of the Huab River on Krone 721 Farm and at Sanianab Locality only. There, fine grained sandstones of the underlying

Karoo Sequence developed autobrecciation, alveolar-texture and typical soil colours (marble texture) in a 30 cm thick zone (Fig. 5.1-2, Plate III-3). Field relationships suggest that the presence of such well-drained palaeosols depends mainly on their preservation potential which is essentially controlled by the structural position of the outcrop area during and after soil development as well as the degree of incision of the overlying Krone Member.

Fig. 5.1-1
Interpolated recent topography of the Karoo-Etendeka unconformity. Three major faults reflect the tectonic zonation of the area. Most minor fault have not been considered in this model. X,Y and Z values for the position of the unconformity have been obtained during field work.

Fig. 5.1-2
Outcrop photographs showing the Krone Member at the Karoo-Etendeka unconformity. In Figure A Krone Member conglomerates erode in several metres wide channels into the underlying Gai-As Formation. The erosional boundary is uneven and pedogenic features are missing. In Figure B fine grained sandstones of the Tsarabis Formation developed autobrecciation, alveolar-texture and typical soil colours (marble texture) in a 30 cm thick zone below the unconformity. Erosional features are missing. Exposure A is located at a cliff in the vicinity of the north-eastern Huab Outliers (S20°35´50´´/E14°09´54´´). Exposure B is at the southern slope of an isolated ridge east of the housings on Krone 721 Farm (S20°30´20/E14°01´32´´). Compare with Plates III-2 and III-3.

5.1.1 EVIDENCE FOR TECTONIC ACTIVITY

Dramatic tectonic activity deformed the sediments to either side of the unconformity. Below the unconformity brittle fracturing is common (extensional fissures, see below chapter 5.1.1.2) whereas soft sediment deformation is abundant above (e.g. at Krone Farm, chapter 5.3.5.1). All deformation features reveal extensional block tectonics accompanied by a regional uplift. The unequivocal evidence for this regional uplift is the pre-Etendeka Group unconformity itself. The importance of it is best emphasised by the complete absence of Triassic and Jurassic strata in the Huab area. The uplift prevailed during Etendeka Group deposition, which is reflected by the spatial- and thickness distribution of Etendeka Group sedimentary and volcanic successions.

5.1.1.1 PALAEOTOPOGRAPHY

As mentioned above, the age of the strata underlying the pre-Cretaceous unconformity is highly variable. North-east of Krone 721 Farm, nearby Sanianab Locality and 20 km north of Terrace Bay the Twyfelfontein Formation rests with pronounced angular unconformity on top of metamorphosed Pan-African basement. In the central Huab area it rests on Karoo rocks. Immediately north and south of the Huab River the Lower Permian Verbrandeberg and Tsarabis formations form the footwall underneath the unconformity, whereas the Upper Permian Gai-As Formation forms the footwall elsewhere in the Huab area. This diachroneity of the sub-Twyfelfontein Formation surface reflects the tectonic architecture of a pre-Etendeka Group uplift: Within the Huab Basin the Karoo strata is towards the north successively more completely preserved below the pre-Cretaceous unconformity. Therefore the amount of eroded Karoo strata was higher in the northern vicinity than in the southern vicinity of the Huab area, which indicates that the pre-Etendeka Group uplift was more accentuated in the northern Huab area than farther south.

The corresponding palaeo-relief was essentially fault generated, as reflected by the spatial distribution and facies variations of the Krone Member (see chapter 4.1.5), that immediately follows onto the pre-Etendeka Group unconformity: Palaeo-current vectors (Fig. 4.1-2) correspond well with thickness distribution, both indicating a depocentre immediately north of the present day Huab River. There, the palaeo-current directions are highly variable, but favouring westerly and minor easterly orientations. South of the depocentre palaeo-current vectors are consistently towards the north directed, suggesting a southern elevated source area for the braidplain deposits found in the southern Krone Member area. The northernmost Krone Member area, instead, comprises alluvial fans and debris flows, which probably derived from proximal scarps located at the northern boundary of the depository. In the area west of the Uniab Fault zone the small average pebble size, reduced thickness and limited extent of the Krone Member indicate restricted drainage towards the west.

On a basin wide scale, the syn-Krone Member relief reveals a WSW-ENE trending halfgraben structure with a master fault running immediately north of the present day Huab River axis. The relief from the basin margins to the basin floor exceeds 100 m in a palaeo-topographic reconstruction of the eastern Huab area (Fig. 4.1-3). The basin floor was slightly inclined towards the northern, relatively steep boundary, which is indicated by the northwardly thickening Krone Member. This slightly northwardly directed dip of the basin floor indicates local tectonic inversion, as it contrasts with the above described accentuated uplift at the northern Huab Basin prior to the deposition of the Krone Member.

The restricted occurrences of the Krone Member in the western Huab area point to a relatively elevated basin floor. This coastal swell coincides with the "Skeleton Coast Uplift" discussed below.

The Skeleton Coast Uplift

At numerous localities in the coastal area of north-western Namibia the Etendeka Group follows directly on Pan-African basement (e.g. 6 km east of the Toscanini Well; north of Terrace Bay; at both locations no Karoo is preserved). These locations underwent enhanced uplift and erosion prior to the deposition of the Etendeka Group. Furthermore, they remained in an relatively uplifted position during the deposition of the basal Etendeka Group, which is mirrored by the low thicknesses or complete absence of the Twyfelfontein Formation. Porada et al. (1996) proposed the term "Skeleton Coast Uplift" for the uplifted zone running parallel to the present coast (Fig. 3.4-1). They assume that the major portion of this structure has subsided during post-rift thermal subsidence and is thus hidden underneath the continental shelf and slope. In contrast to this, the Toscanini Well at north-western Huabmond documents a thick sequence of Permian Karoo strata preserved nearby the coastline thus west of the missing Karoo localities. Possibly the well was located west of the "Skeleton Coast Uplift" implying that at this latitude the uplift did not extend into the present day shelf and slope.

Farther south, the "Skeleton Coast Uplift" might extend into the Albin Ridge, which is reflected by the successively more complete Karoo strata along a traverse from the Albin Ridge via the Goboboseb Mts. to Mt. Brandberg. At Albin Ridge the youngest preserved Karoo strata is the Permian Gai-As Formation (section 31, Fig. 4.1-4), whereas at Mt. Brandberg the Triassic Omingonde Formation is largely preserved under the pre-Cretaceous unconformity (section B-P1, Fig. 3.1-4). However, this W-E traverse follows the axis of an approximately 40 km wide halfgraben (Goboboseb Basin *sensu* Porada et al., 1996; chapter 3), which complicates the geometries of the uplifted zone in this part of the study area.

5.1.1.2 BRITTLE DEFORMATION

The above described palaeo-topography and regional uplift reveals regional tectonic processes that, on a smaller scale, are mirrored by brittle deformation features. Sand filled fissures are the most obvious features, which can be observed at the pre-Etendeka Group unconformity (Fig. 5.1-3). Those fissures start at the pre-Cretaceous unconformity and reach up to 25 m deep into the underlying Karoo strata (usually Gai-As or Doros Formation). The gap-width at the fissure tops varies from less than 1 cm to more than 20 cm and it corresponds with the fissure depth. The fissures are commonly subvertical and their orientation is sub-parallel to the general fault trends: N-S, NW-SE, SW-NE and E-W trends are favoured (Fig. 5.1-4). N-S and NW-SE orientations predominate in the vicinity of the Ambrosiusberg and Uniab fault systems. E-W and SW-NE trends become more abundant towards the central and eastern Huab area, however N-S and NW-SE trends are still frequent there.

Fissures were found almost everywhere where the unconformity is preserved. Maximum density and size of the fissures are observed in the central Huab area around Gai-As and Klein Gai-As and in the Skeleton Coast Park.

The fissures are filled with aeolian sand and sometimes they also contain wind-grooved pebbles of the Krone Member. At some places the fissures do not end directly at the top of the Karoo rocks, but they can be traced up-section across the pre-Cretaceous unconformity into the lowermost Twyfelfontein Formation. There, the fissure delimiting faults are indistinct, which reveals a formation of the fissures prior to the complete solidification of the basal Twyfelfontein Formation. The ubiquitous presence of basal Twyfelfontein Formation fills within the fissures indicates, that the main peak of fissure formation coincides with the onset of Etendeka Group (Krone Member) deposition.

Fig. 5.1-3

Sand filled fissures occurring in the vicinity of the Bergsig Fault system. The fissures reach from the Lower Cretaceous Twyfelfontein Formation down into the Permian Gai-As Formation. Outcrop is located in a small gully 50 m east of a vehicle track towards south (S20°45´52´´/E14°06´58´´).

Fig. 5.1-4

Rose diagrams showing the strike orientations of sand filled fissures and dykes in the Huab area.

5.1.1.3 TECTONIC INTERPRETATION

The specific anatomy of the pre-Cretaceous unconformity is a result of regional uplift with block tectonics and associated extensional fracturing. Estimates of the primary thicknesses of the missing Triassic and Jurassic strata are difficult, as preserved corresponding strata at other places accommodated in different tectonic environments (e.g. Erongo Mts., Mt. Waterberg). But, the time and spatial extent of the hiatus give an impression of the amount and spatial extent of relative uplift. The dominance of N-S, NE-SW and NW-SE trending extensional fractures (e.g. faults of Fig. 5.3-11; sand filled fissures and dykes of Fig. 5.1-4) indicates that the favoured direction of extension has been approximately ENE-WSW.

Following common rift models (e.g. Keen, 1985), the most likely driving force for this extensional uplift was accelerated mantle convection with high heat flow, preceding flood basalt volcanism. This interpretation is consistent with the observed close age relation of the Krone Member including the sand filled fissures to the first Etendeka lavas (chapter 4.1-10). Furthermore, this close age relation highlights the genetic coupling of extensional, domal uplift with the onset of flood basalt volcanism (chapter 5.4).

5.2 ETENDEKA GROUP DISCONFORMITIES

Within the Etendeka Group at least four disconformities/unconformities have been observed, which do not mark significant hiatal gaps as they embrace short periods of time only.

The lowermost disconformity is placed within the Twyfelfontein Formation between the Mixed Aeolian-Fluvial Unit and the Main Aeolian Unit. A second disconformity is found between the latter and the succeeding Upper Aeolian Unit. Within the Awahab Formation another disconformity divides the olivin-phyric basalts from the basalts/basaltic andesites (disconformity I; Jerram et al., 1999b). Finally, a more pronounced disconformity divides the Awahab Formation from the Tafelberg Formation (disconformity II; Jerram et al., 1999b).

5.2.1 MIXED AEOLIAN-FLUVIAL-/MAIN AEOLIAN UNIT DISCONFORMITY

An abrupt transition from small dune forms to large-scale draa and transverse dune forms occurs above the Mixed Aeolian-Fluvial Unit. The bounding surface separating the Mixed Aeolian-Fluvial Unit from the Main Aeolian Unit represents a sharp distinctive horizon that can be traced for over 30 km from the Huab River to immediately south of the Huab Outliers. Therefore it has the characteristic of a major bounding surface rather than a simple bedform-set confining surface. A notable hiatus separating the two facies types is indicated by two types of structures which exclusively occur at the bounding surface: (1) Vertical tubular burrows of *Skolithos*-type (Fig. 5.2-1, Plate III-1) and (2) rootlets, both indicating surface stabilisation on a partly stabilised substrate. Additionally steep faults, that displace deposits of the Mixed Aeolian-Fluvial Unit are often terminated at the bounding surface. However, at some places faults are crossing the bounding surface exhibiting soft sediment faulting characteristics on either side.

The duration of the hiatus was too short for a complete consolidation of the units below the disconformity, as indicated by soft sediment faulting on either side of the disconformity. Furthermore, time was too short for the development of a proper palaeosol with abundant rootlets and surface crusts. Therefore a period of non-deposition lasting thousands of years is favoured rather than a period lasting several hundred thousands of years.

Fig. 5.2-1

Vertical burrow closely below the Mixed Aeolian-Fluvial-/ Main Aeolian Unit disconformity at Verbrandeberg locality (upper cliff at the end of road D-3254; S20°37´26´´/E14°25´01´´). Compare with Plate III-1.

5.2.2 MAIN AEOLIAN-/ UPPER AEOLIAN UNIT DISCONFORMITY

The presence of the Main Aeolian-/ Upper Aeolian Unit disconformity has first been reported by Jerram et al. (1999a+b). The aeolian sandstones of the Upper Aeolian Unit are distinct from deposits of the Main Aeolian Unit as they occur after the initial eruption of the basaltic lavas in the region (Tafelkop Member basalts of the Awahab Formation). These initial lavas are restricted in spatial extent and consequently the

Upper Aeolian Unit lies either directly on the lowermost basalts or on sedimentary deposits of the Main Aeolian Unit. In the latter case, the two successions are separated by a laterally extensive bounding surface.

This disconformity reflects the onset of extensive flood basalt volcanism with the initiation of the Awahab Formation, rather than a hiatal gap. The abrupt decrease of sand supply by the first lavas covering the erg system caused a rapid change in erg dynamics which is reflected by the bounding surface between the two aeolian units.

5.2.3 VOLCANIC DISCONFORMITIES

5.2.3.1 DISCONFORMITY I

Around the Huab Outliers, a conspicuous disconformity within the Awahab Formation has been observed. There, Tafelkop-Interdune Member basalts (olivin-phyric basalts) thicken towards the Huab Outliers with the units thinning to the north and to the west. This geometry fits to a shallow shield volcano, as suggested by Jerram et al. (1999b). Aphyric flows of Tafelberg-type basalts onlapped this shield volcano and smoothened its relief.

The disconformity shows no evidence for erosion, soil development and other signs of longer phases of non-deposition. It documents an abrupt change in volcanism rather than a stratigraphic gap.

The shield volcano, consisting of Tafelkop-Interdune Member basalts, represents a phase of pahoehoe style volcanism. The onlapping basalt/basaltic andesites document a change to more voluminous, laterally extensive flows.

5.2.3.2 DISCONFORMITY II

The Awahab and Tafelberg formations of the Etendeka Group are largely juxtaposed and separated by a major disconformity (Milner, 1988; Milner et al., 1994; disconformity II, Jerram et al., 1999b): Tafelberg-type basalts and higher units (all Tafelberg Formation) onlap over an erosional surface onto the Awahab Formation. As shown in Figure 5.2-2, at Fontaine 717 Farm an angular unconformity is well defined between the Springbok Quartz Latite Member (upper Awahab Formation) and the Wêreldsend Quartz Latite Member and underlying basalts (all Tafelberg Formation). The estimated palaeo-relief in this section exceeds 120 m over a distance of 1.5 km (section not measured, estimation based on panorama view). At the Tafelberg type locality the units that overly the disconformity onlap from the south and the amount of palaeo-relief is approximately 200 m (Milner, 1988). The genesis of this palaeo-relief is still a matter of debate: River channel erosion, erosion along a fault scarp and steep flow front topography due to high flow viscosity are discussed (Duncan & Milner, pers. comm. 1997). Erosion is the most likely explanation, because no flow top facies of the Springbok Quartz Latite has been observed under the disconformity. Additionally, sedimentary material has been found at the contact.

Despite the strong evidence for an erosional phase, precise $^{40}Ar/^{39}Ar$ datings of units on either side of the disconformity give the same age within the error range (Milner et al., 1994). This infers rapid erosion, probably facilitated by quick alteration. Despite the negligible time-stratigraphic gap, this unconformity is believed to be part of a more extensive regional disconformity (Milner, 1988; Milner et al., 1994) that is traceable into the Paraná Basin (Milner et al., 1995a).

Such a regional disconformity implies a major volcanic-tectonic event, important for unravelling the evolutionary steps of the Paraná-Etendeka Flood Basalt Province.

However, it should be noted that at this stage of investigations the degree of down-cutting on this disconformity in the Tafelberg type locality and its full regional context is not fully constrained.

Fig. 5.2-2

Schematic cross-section of the disconformable relationships between the Awahab and Tafelberg formations. The Wêreldsend Quartz Latite and a sequence of underlying basalts (Tafelberg Formation) onlap the Springbok Quartz Latite (Awahab Formation) from the north. View is to the east from the road 2620 to Kamanjab at approximately 20°18´40´´ South in latitude.

5.3 FAULT ACTIVITY DURING ETENDEKA GROUP DEPOSITION

The entire Twyfelfontein Formation shows brittle and plastic deformation features. Soft sediment deformation structures indicate that faults have been active during the deposition of the Krone Member and the following units of the Twyfelfontein Formation. The deformations are mainly attributed to synsedimentary block faulting, but also to gravitational collapse of unconsolidated sediments. The latter can be partly explained with seismic events and the formation of a tectonically triggered relief.

Deformation continued after deposition of the Twyfelfontein Formation and also affected the volcanic succession, preferentially in the coastal area.

5.3.1 SPATIAL DIVISION OF ETENDEKA GROUP DEFORMATIONS

The spatial and temporal variations in depositional and tectonic style are rather complex. Block faulting is a widespread phenomenon in the Huab area. At least five prominent N-S striking fault zones cause a tectonic zonation (Fig. 5.3-11). Total offsets along these faults range from a few metres in the eastern Huab area to more than 100 m in the coastal region. E-W trending faults appearing in the eastern Huab area show offsets of about 25 m.

Two N-S striking fault zones separate the Huab and adjacent areas into three distinct tectonic domains: The Ambrosiusberg Fault zone separates the Coastal Zone from the Intermediate Zone. Farther east the Uniab Fault system defines the boundary to the Eastern Zone.

5.3.1.1 COASTAL ZONE:

The coastal area is characterised by listric faulting and subsequent tilt-block tectonics. In this context the Purros Fault zone is of major importance as it separates the Namibian continental margin into two domains. The Purros Fault zone extends from the northern Kaokoveld into the Etendeka Plateau. Farther south it extends into the Huab area, where it is termed the Ambrosiusberg Fault zone. The latter might extend into the NNW-SSE striking fault system, which delimits the Albin Ridge along its eastern flank.

The domain west of the Purros Fault zone (and its extensions) is characterised by an "en echelon" arrangement of NNW-SSE and N-S trending listric faults and subsequent tilting. Commonly tilted blocks dip eastwards with dip-angles of up to 30°.

The influence of pre- and syn-Twyfelfontein rotational block faulting is particularly well documented in several outcrops close to the present day coast line such as Sanianab and Ambrosiusberg localities. There the Twyfelfontein Formation forms wedge shaped sediment bodies developed on top of the Gai-As Formation (Ambrosiusberg Locality; Fig. 5.3-3). Post-volcanic rotational block faulting is well documented at Albin Ridge and Kharu-Gaiseb localities. At these places lava flows were encountered in the block-rotations.

5.3.1.2 INTERMEDIATE ZONE:

This zone extends from the Ambrosiusberg Fault system until the Uniab Fault zone approximately 20 km farther east. N-S and NNW-SSE trending faults are still prominent, but in addition a SSW-NNE fault trend becomes important. Block tilting is less pronounced and less common. Block-rotations with less than 20°, but also roll-over structures and vertical block tectonics are abundant.
The Poiki Locality at the eastern margin of this zone provides another key locality for a post-volcanic (at least post Awahab Formation) block-rotation. Such pronounced rotations of Etendeka Group extrusives occur only exclusively west of the Uniab Fault zone.

5.3.1.3 EASTERN ZONE:

East of the Uniab Fault zone block-rotations play a minor role in block-tectonics, but vertical offsets are still conspicuous and sometimes exceed 70 m. The Wêreldsend Fault 30 km east of the Uniab Fault might mark the easternmost occurrence of clear block tilting. Farther east only a few minor, rather low angle rotations have been observed.
Beside the N-S trend SW-NE and E-W fault trends play an important role and can be well studied at the southern Huab Outliers and around Twyfelfontein. At Krone 721 Farm deformation features are related to the elsewhere obscured Huab Faults system. This fault system traces roughly the axis of the present day Huab River, and it gives the Huab Basin a halfgraben-shaped geometry.

5.3.2 TEMPORAL DIVISION

Faulting is a phenomenon that accompanies continuously the whole Etendeka Group. Peaks in deformation and spatial migration of deformation maxima are distinguishable, when supposing that the stratigraphic units of the Etendeka Group define a time-stratigraphic succession free of pronounced diachroneities.

5.3.2.1 PRE- AND SYN-TWYFELFONTEIN FAULTING

Prior to the deposition of the Krone Member a pronounced palaeo-relief existed, which was almost completely fault generated. Block faulting, uplift in the coastal area (Coastal and Intermediate Zone) and possibly strike-slip movement along the Huab Fault are documented by the Krone Member conglomerate and its equivalents. With the succeeding aeolian units block-rotations became more pronounced in the coastal area, as reflected by the wedge-geometries of Ambrosiusberg and Sanianab localities (Coastal Zone). In the Intermediate and Eastern Zone the tectonic activity seems to increase with the deposition of the Main Aeolian Unit, which indicates another tectonic peak. With this tectonic peak accommodation space was created as expressed by the local occurrences of the Main Aeolian Unit in hangingwall traps along the Wêreldsend and Uniab Fault systems.

Faulting continued during deposition of the Upper Aeolian Unit, but with reduced intensity. After the onset of flood volcanism (Main-/Upper Aeolian Unit disconformity) vertical offsets became rather inferior in the Eastern Zone.

These above spatial features suggest several deformation peaks. On a closer look, brittle and soft sediment deformation features reveal a clear subdivision into three phases of syn-Twyfelfontein Formation faulting:

1) An early phase during the basal Krone Member. This is documented by the extension of sand filled fissures from the Karoo rocks over the pre-Cretaceous unconformity into the lowermost Krone Member. This feature has been abundantly observed south of the Huab Outliers.
2) A pre-Main Aeolian Unit phase. Evidence for this phase is given by extensional synsedimentary faults, which terminate at the bounding surface to the Main Aeolian Unit. This feature is best exposed along the Huab River and immediately south of the Huab Outliers
3) A syn-Main Aeolian Unit phase: Soft sediment faulting dominates in the Main Aeolian Unit. Rather common is this phenomenon in the Twyfelfontein area. An obvious example is a prominent NW-SE trending fault west of Twyfelfontein, which is down-faulting the Karoo and Damara rocks to the east for several tens of metres, but the offset diminishes in the upper portion of the Main Aeolian Unit (Fig. 5.3-1).

Fig. 5.3-1

Outcrop photograph and interpretation of a NE-SW striking normal fault west of Twyfelfontein (location is 500 m north of the track to Twyfelfontein at S20°35´20´´/E14°20´05´´). The footwall comprises north-east tilted Permian strata overlain by Etendeka Group sedimentary rocks (Twyfelfontein Fm.). On the hangingwall the lower units of the Twyfelfontein Formation are flexurally deformed due to syn-depositional fault activity, whereas the upper units are almost unaffected from fault movements. The apparent curvature of the fault plane is a result of outcrop geometries rather than a true bending.

5.3.2.2 SYN-VOLCANIC FAULTING

Early syn-volcanic faulting is documented by the sand filled fissures in the lower Awahab Formation of the Eastern Zone. They only reflect tension without any vertical displacement. In the Coastal Zone syn-volcanic faulting with pronounced block-rotations is expressed by decreasing dip angles of the volcanic units from 15° at the base of the succession to 10-12° at the top. Furthermore, prominent vertical offsets are indicated by the absence of the Awahab Formation quartz latites between the Springbok and Wêreldsend Quartz Latite.

Farther east (Intermediate and Eastern Zone) syn-volcanic faulting is less conspicuous, but a tectonic peak is indicated by the disconformity between the Awahab- and Tafelberg formation (chapter 5.2.3.2).

5.3.2.3 POST-VOLCANIC FAULTING

Post-volcanic faulting can be observed all over the study area. Many faults cut through the whole succession and therefore their final movement was clearly post-volcanic. Pronounced post-volcanic tectonics occurred only in the Coastal and Intermediate Zone as documented by the final block-rotations at the Albin Ridge, Poiki Locality and the Kharu-Gaiseb Section. However, it should be noted that the term "post-volcanic" only refers to the preserved volcanic succession. It is likely, that the extrusives associated with the late stages of Etendeka volcanism are not preserved. Therefore it is possible, that tectonic movements appear to be post-volcanic although they occurred during a late volcanic stage.

5.3.3 EXAMPLES FOR THE COASTAL ZONE

5.3.3.1 SANIANAB LOCALITY

At Sanianab Locality (Fig. 5.3-2, Plate V-2) (location at S20°01´58´´/E19°12´07´´) the Krone Member is rotated eastwardly causing a dip angle between 6° and 10°. The succeeding aeolianites drape this tilt-block relief and form a wedge shaped body. The sedimentary succession is followed by a flat lying lava flow, which forms the top of the outcrop.

The Krone Member is repeatedly faulted by sets of small normal faults causing offsets of a few tens of centimetres. Up-section most faults diminish in the overlying aeolian unit, but a few terminate at the base of the lava. NNW-SSE fault trends dominate with a steep inclination towards WSW. A few conjugated normal faults dip steeply to the north-east. Synsedimentary fault movements are expressed by thickness-changes or pinching out of single sedimentary horizons when tracing them across the faults (Plate IV-1).

A few steeply westwardly dipping normal faults extend from the basement up-section into the aeolianites where they diminish. They displace the base of the Twyfelfontein Formation for up to 20 cm and their synsedimentary activity has been confirmed by soft-sediment deformation features within the aeolian unit.

<u>Interpretation</u>

The orientation of the small scale normal faults within the Krone Member and the following aeolian unit coincides with the more regional faults in the area. As the course of the Ambrosiusberg Fault zone runs through the outcrop area all observed deformation features within the Krone Member are attributed to the reactivation of the Ambrosiusberg Fault zone.

Fig. 5.3-2

Outcrop photograph and interpretation of a NE-SW section at Sanianab Locality. A wedge shaped body of the Twyfelfontein Formation developed on an eastwardly tilted footwall. Small scale faults exhibit syn-sedimentary soft sediment deformation as shown in the detailed Figure. The exposure is located at the southern bank of the Sanianab River, approximately 500 m east of the adjacent dune field (S20°01´58´´/E13°18´07´´). Compare with Plates IV-1 and V-2.

5.3.3.2 AMBROSIUSBERG LOCALITY

A to the Sanianab Locality similar geometry is developed farther south along the Ambrosiusberg Fault zone. In the Huab area, 9 km east of the Huab mouth (location at S20°54´12´´/E13°33´59´´), a wedge of sandstones of the Twyfelfontein Formation rests on an eastwardly rotated block of the Permian Doros Formation (Fig. 5.3-3, Plate V-1). Again, this sedimentary body is succeeded by flat-lying lavas.

The eastern side of the sediment body is bounded by a steep, westwardly dipping fault, which is slightly down-faulting the eastern footwall (except the lavas). Another steep fault cuts through the Twyfelfontein Formation causing a minor vertical offset of the unconformity between the Twyfelfontein Formation and Doros Formation. Immediately west of the sediment body, a steep fault causes a pronounced vertical offset within the basal lavas of the Etendeka Group. All faults are intruded by basaltic dykes which are cutting through both, the sedimentary and volcanic rocks.

At the top of the Doros Formation numerous roughly N-S striking fissures occur, which are filled with material of the overlying sandstone. Up-section, the fissures extend into this sandstone unit where they become indistinct faults that die out after a few centimetres. The overlying sandstones presumably belong to the Twyfelfontein Formation, as they comprise well rounded and well sorted medium size grained sands. Sedimentary structures are mainly aeolian, but small scale cross-bedding suggests some fluvial reworking. Probably the sedimentary unit represents a distal type of the Mixed Aeolian-Fluvial Unit. The Krone Member and other units of the Twyfelfontein Formation are missing.

<u>Interpretation</u>

The geometrical relations enable to distinguish between three tectonic phases:

1) Early phase of block-rotation and opening of fissures:

Block-rotation created a hangingwall trap during the Mixed Aeolian-Fluvial Unit. This is indicated by the sand filled fissures and the lack of syn-rotational erosion of the Doros block: If the Doros surface had been exposed during tilting, exaggerated erosion would have taken place at the relatively uplifted western part of the block. As no enhanced erosion of the western side and no corresponding Doros Formation clasts have been observed in the eastern side of the tilt-block, the Doros surface must have been sheltered by the sands of the Twyfelfontein Formation. Another hint for a rotation of the Doros block contemporaneous with deposition of the Mixed Aeolian-Fluvial Unit is given by the fissures. They are filled with the same sand that comprises the overlying Mixed Aeolian-Fluvial Unit. The soft sediment faulting immediately above the fissures (at the base of the Mixed Aeolian-Fluvial Unit) confirms opening of the fissures prior to the consolidation of the sands. The fissures opened due to extensional tectonic stresses, related to block-rotation. This is confirmed by their favoured N-S directions, that coincides with the N-S courses of the faults, which are bounding the rotated Doros footwall (Fig. 5.3-3).

2) Late phase of block faulting:

The steep faults at the western side of the outcrop vertically displace the lava flows. A later phase of fault activity is implied, as faulting occurred after the emplacement of the first two basaltic lava flows that are preserved at the outcrop.

3) Dyke intrusion into the faults:

The dykes intruded the faults and cut through the whole succession. The faults have not been reactivated after dyke emplacement. Therefore, the intrusion either post-dates the late phase of block faulting or it occurred contemporaneously with it.

Fig. 5.3-3

Cross section of the Ambrosiusberg Locality showing a wedge shaped body of the Twyfelfontein Formation resting on an eastwardly tilted footwall. The exposure is located at the southern bank of the Huab River, approximately 9 km east of the present coastline (S20°54´12´´/E13°33´59´´). Compare with Plate V-1.

5.3.3.3 ALBIN RIDGE

The Albin Ridge is located 15 km east of the present Atlantic coast. It is running parallel to the present day coastline (NNW-SSE) (Fig. 1-3) and exposes sedimentary and volcanic strata. The strata is eastwardly dipping with dip-angles between 20-30° (Fig. 5.3-4). Tilting of the strata was controlled by a prominent listric fault, which runs along the eastern flank of the ridge. The northern outcrop area comprises the Permian Karoo Sequence (chapter 2.2, Fig. 4.1-4), which is disconformably succeeded by a thin conglomeratic sequence referred to as the Albin Conglomerate (Swart, 1992/93; chapter 4.1.6.1). A thin palaeosol is developed below the contact. The conglomeratic sequence is succeeded by a sandstone unit which consists of fluvial pebbly sandstones interlayered with aeolian sandsheets. Up-section, Etendeka basalts and a quartz latite unit follow the clastics and form the top of the ridge.

A remarkable feature are sand filled fissures that start within the mixed fluvio-aeolian sandstone unit and penetrate down into the underlying conglomerate. The fissures are filled with sandy material that obviously derived from the sandstone unit.

Interpretation

The nearby presence of a prominent listric fault suggests that the Albin Conglomerate derived from fault generated scarps. Probably the age of the Albin Conglomerate is close to the age of the Krone Member or of the Mixed Aeolian-Fluvial Unit (chapter 4.1.12.1). The deposition of these sequences was temporarily associated with a tectonically rather active phase (chapter 5.3.2.1), which is confirmed by the reactivation of the Ambrosiusberg Fault at Ambrosiusberg and Sanianab localities (chapters 5.3.3.1 and 5.3.3.2). Enhanced fault activity is also reflected by the sand filled fissures within the Albin Conglomerate. Wittig (1976) investigated similar fissures in the Etjo Formation of Mt. Gamsberg and he relates their formation to earthquakes. Hence, it is suggested that the fissures at Albin Ridge also document such palaeo-earthquakes.

The main peak of block-rotation affecting the sequence at Albin Ridge commenced after lava emplacement, which is confirmed by the similar dip-angles of the sedimentary and volcanic units.

Fig. 5.3-4

Outcrop photograph and interpretation of a SE-NW section at Albin Ridge. Karoo and Etendeka rocks are uniformly tilted towards the east. The fluvio-aeolian sandstone and the Albin Conglomerate presumably belong to the Etendeka Group. The photograph shows the northern edge of the Albin Ridge (S21°25´34´´/E13°53´44´´).

5.3.3.4 TILT-BLOCKS NORTH OF TERRACE BAY

An at least 300 m thick sequence of aeolian sediments interleaved with up to 18 lava units is exposed 20 km north of Terrace Bay (exposures start at the road at S19°51´34´´/E12°57´53´´, see Fig. 1-3). The entire succession is dipping to the east, which is topographically expressed by a series of parallel NNE-

SSW trending tiny ridges and shallow depressions in between. The sequence starts with a thin pebbly sandstone unit, which is most likely the Krone Member (chapter 4.1.6.3). Aeolianites of the Twyfelfontein Formation follow. They are interbedded by 18 lava flows, each up to 14 m thick. The individual flow thicknesses are laterally highly variable due to the topography of the underlying aeolianites: At some places completely preserved barchan dune forms are preserved below lava flows.

The contacts between lavas and aeolianites are inclined towards the east with 16° to 25°. The measurements were obtained at places without dune topography. No systematic trend to higher or lower dip angles was recognised.

Interpretation

The outcrop exhibits the thickest succession of alternating lavas and sandstones that have been observed in the study area. The 18 lava-sandstone couplets may represent an undisturbed stratigraphy, but moreover it is likely that at places the contacts between lava and sandstone units are tectonic. In the latter case the observed 18 sets of sandstone-lava beds would correspond to an "en echelon" arrangement of NNE-SSW trending listric faults. The resulting group of synthetic rotated blocks would have caused a repeated cropping out of individual lava-sandstone couplets (Fig. 5.3-5).

It is difficult to observe the corresponding faults because of poor exposure. Here it is speculated, that the primary stratigraphic thickness was rather high, but less than the 300 m observed in the field. Furthermore, it is assumed that a few basalt-sandstone contacts are of tectonic origin, as outlined above. This assumption is based on the fact, that northerly trending faults are rather frequent in the area.

The presumed extraordinary high thickness of the sequence is explained with its position within the South Atlantic rift. According to Stollhofen (1999) and Glen et al. (1997) the South Atlantic rift already developed an accentuated rift depression (rift-valley) prior to the onset of Etendeka flood volcanism. This rift depression provided the accommodation space for a thick succession. Later, subsequent block-rotations resulted in the observed eastwards directed dip. These block-rotations possibly happened roughly simultaneous to the post-lava block-rotations at Albin Ridge and Poiki Locality. It should be noted that the determined age relation announces tilting after initial lava deposition only and it does not necessarily imply a complete post-Etendeka-Group movement (chapter 5.3.2.3).

Fig. 5.3-5

Measured W-E section of the area north of Terrace Bay combined with an aerial photograph. Repeated tilting of a lava-sandstone couplet along normal faults caused the observed geometries. The western side of the section starts immediately west of the coastal road, ca. 20 km north of Terrace Bay (S19°51´34´´/E12°57´53´´).

5.3.3.5 TILTING OF THE KHARU-GAISEB CONGLOMERATES

A sequence of conglomerates tilted towards the east is exposed 20 km north-east of Terrace Bay in several small outcrops located to the south and to the north of the Kharu-Gaiseb River (main deposit at S19°51´25´´/E13°37´10´´) (chapter 4.1.6.4). The conglomerates consist entirely of volcanic Etendeka Group material, which gives a maximum age for their deposition. Ward & Martin (1987) suggest that the Kharu-Gaiseb beds were laid down in a proximal position from a scarp source in the west. They assume that the scarp source was located along an active synsedimentary fault.

It should be noted that this exposure has not been visited by the author. The interpretations given here are based on the observations reported in Ward & Martin (1987).

Interpretation

Ward & Martin (1987) speculate that the conglomerates were shed eastwards off a high-lying rift shoulder that developed after the effusion of the Etendeka volcanics and was situated in the vicinity of the present day coastline. But despite the interpretation of Ward & Martin (1987) a pronounced rift shoulder topography already existed during Etendeka volcanism (chapter 5.4), which implies that the Kharu-Gaiseb Conglomerate could have formed not after, but during Etendeka volcanism. A deposition of the conglomerate during the main rift shoulder uplift appears unlikely, because of the low preservation potential for sediments that accumulate in the vicinity of a high-lying rift shoulder.

5.3.3.6 TILTING AND DOWN-FAULTING OF COASTAL VOLCANIC SUCCESSIONS

West of the Ambrosiusberg Fault zone differential rotation of volcanic units in successive fault blocks is indicated by an overall increase in eastwards directed dip towards the coast. Conjugated sets of SSW and SE trending normal faults intersect the tilted blocks and cause vertical offsets of up to 90 m.
For example, a coastal tilt-block north of the Huab River is truncated by a normal fault to the north (Fig. 5.3-11). The block itself dips with 15° to the east, but almost no tilting has been observed farther north.

Interpretation

The conjugated SSW and SE trending normal faults that are bounding rotated blocks to the north or to the south are older than the listric faults. Therefore it is assumed that the conjugated faults controlled the development and intersection of the N-S trending listric faults.

5.3.4 EXAMPLES FOR THE INTERMEDIATE ZONE

5.3.4.1 POIKI LOCALITY

A conspicuous ridge (Fig. 5.3-6, Plate V-3), being almost triangular in cross-section, is located ca. 2 km north of the present Huab River in the vicinity of the Uniab Fault (location at S20°46´35´´/E13°44´26´´). The ridge comprises aeolianites of the Twyfelfontein Formation succeeded by a 80 m thick lava sequence. The lavas itself and the contact between the lavas and the sandstone are inclined with 15° to 20° to the west. This implies that the lavas and the aeolianites have been tilted together and the lavas do not onlap a primary dune topography. The foresets of the underlying purely aeolian sandstone dip eastwardly with a dip angle of maximum of 15°. The primary dip angle of dune foresets should be approximately 30°. The reduction to 15° supports the recognised tilting of the whole sandstone-lava block.
The exposures immediately west and east of the ridge show Permian Karoo rocks (Gai-As Formation) with lavas on top. There, the Twyfelfontein Formation is entirely missing and the whole sequence is flat lying.

Interpretation

The Poiki Locality documents, that not only the Etendeka Group sedimentary rock, but also the lavas were affected by block tilting along major N-S striking faults. The absence of the Twyfelfontein Formation at the neighbouring outcrops gives evidence for down-faulting at the ridge locality prior to lava deposition. In this hangingwall position, the Twyfelfontein Formation aeolianites accommodated, before they were covered by the Etendeka lavas. Pronounced block tilting commenced after implacement of the lavas and caused the observed geometries.

Fig. 5.3-6
Outcrop photograph and interpretation showing a westwardly tilted block comprising Lower Etendeka Group sedimentary and volcanic rocks (Twyfelfontein and Awahab formation). The strata within the dotted rectangle appears to be horizontally bedded because of perspective distortion, but is in fact westwardly dipping. The exposure is located in the Skeleton Coast Park in the vicinity of the Uniab Fault (see Fig. 5.3-11). There, north of the Huab River an isolated SSW-NNE trending ridge provides the exposures of the Twyfelfontein and Awahab formation (Poiki Locality; S20°46´35´´/E13°44´26´´). Compare with Plate V-3.

5.3.5 EXAMPLES FOR THE EASTERN ZONE

5.3.5.1 SOFT SEDIMENT DEFORMATION AT KRONE FARM

At Krone 721 Farm (see Fig. 2.1-4 for location) soft sediment deformation within the Krone Member is rather conspicuous. In addition, this locality provides good examples of soft sediment deformation within the aeolian units.

The key locality for hydroplastic deformation features within the Krone Member is located at Krone Farm, a few hundred metres east of the former farm housings (location at S20°29´10´´/E14°01´55´´).

The outcrop shows a 30 m long body of the Krone Member, which dips with 20° towards the west (240°) (Fig. 5.3-7, Plate IV-2). Farther east the inclination smoothes out to horizontal. The inclined part shows abundant sheering bands and micro thrusts in a meter scale (Fig. 5.3-8, Plate IV-3) besides water escape structures. The shear and thrust sense coincides with the inclination of the whole body. Associated normal faults are almost vertical and show a maximum displacement of 35 cm. When tracing these faults up-section into higher stratigraphic levels the displacement becomes inconspicuous. Sets of “en echelon“ arranged faults commonly have 5 to 10 cm distance between one another. The orientations of the fault planes are rather variable, though N-S and E-W trends are preferred (Fig. 5.3-8). All deformation geometries show a ductile character due to pre-diagenetic deformation indicating that faulting, shearing and thrusting occurred during or shortly after deposition of the Krone Member.

<u>Interpretation</u>

The outcrop is located in the southern neighbourhood of the SW-NE trending Huab Fault. Therefore the sheering and thrusting within the Krone Member are directly ascribed to synsedimentary activity of the Huab Fault, which would infer a sinistral oblique-slip movement (Fig. 5.3-9). But, it should be mentioned that the Huab Fault is not directly exposed and that its oblique-slip character is not fully constrained.

The hydroplastic sheering and thrusting of the Krone Member could also be explained by debris flow or slumping from a proximal high being located east of the outcrop. The present Karoo and basement rock topography, however, does not supports this thesis, as basement highs are located north of the outcrop instead east of it.

Fig. 5.3-7
Outcrop photograph showing a westwardly dipping body of the Krone Member at Krone 721 Farm. The steep faults of the right figure exhibit soft sediment deformation characteristics. Outcrop location is at a sandstone cliff, 450 m east of the former farm housings (S20°29´10´´/E14°01´55´´, Fig. 5.3-11). Compare with Plate IV-2.

Fig. 5.3-8
Detail of a steep thrust fault exhibiting soft sediment faulting within the Krone Member. (same location as shown in Figure 5.3-7). The figure to the right shows the favoured orientations of this type of faults in the same outcrop area on Krone 721 Farm.

Fig. 5.3-9
Model for plastic sheering and tilting within the Krone Member in the vicinity of the Huab Fault. A sinistral transfer component might have caused the observed plastic deformation features as shown within the doted rectangle. Such geometries are developed at Krone 721 Farm as shown in Figure 5.3-7.

5.3.5.2 ISOLATED BODIES OF THE MAIN AEOLIAN UNIT

West of Gai-As a restricted deposit of the Main Aeolian Unit appears in the vicinity of the Wêreldsend Fault system (location at S20°46´53´´/E13°57´36´´). A N-S trending fault separates two blocks. The western block hosts a 30 m thick succession of the Main Aeolian Unit capped by the lavas of the Awahab Formation. In the eastern block the Main Aeolian Unit is missing and the lavas rest directly on the Mixed Aeolian-Fluvial Unit. The latter displays a similar thickness and facies architecture on both blocks.
A comparable situation has been observed north of the Poiki Locality along the Uniab Fault system (location at S20°44´17´´/E13°44´45´´). At this locality the Krone Member and the Mixed Aeolian-Fluvial Unit are missing.

Interpretation

The restricted, fault bounded occurrence of the Main Aeolian Unit indicates that accommodation space was created in hangingwall traps. The main faulting phase was after the deposition of the underlying units (Karoo rocks, Krone Member, Mixed Aeolian-Fluvial Unit), as indicated by their invariable thickness distribution across the faults. Therefore the peak in vertical displacement was shortly prior to, or during the deposition of the Main Aeolian Unit.

5.3.5.3 EXTENSION FISSURES WITHIN THE AWAHAB FORMATION

In the Huab area sediment filled fissures cut through the lavas of the Tafelkop Member (Awahab Formation). The most conspicuous fissures are around the Huab Outliers. There an up to 1.5 m wide NNW-SSE trending fissure reaches down into the Upper Aeolian Unit of the Twyfelfontein Formation (Fig. 5.3-10, Plate VI-1). In the Goboboseb Mts. sediment filled fissures were found in a similar stratigraphic position, although the Twyfelfontein Formation is missing there. In higher stratigraphic levels no fissures have been observed.

Interpretation

The huge dimensions of the fissures and the fact that they cross many lava flows at once proves that they are of tectonic origin rather than cooling joints. The sand material is completely aeolian which indicates aeolian sand transport over the lava flows. Open fissures provided the only accommodation space for the sands, as other depression were lacking in the area. The occurrence of the fissures in similar stratigraphic levels probably points to a peak in tectonic extension.

Fig. 5.3-10

Sand filled extension fissure within the Awahab Formation at the Huab Outliers. Exposure is on the south-eastern slope of a prominent mountain (S20°40´08´´/E14°08´ 56´´). Compare with Plate VI-1.

Fig. 5.3-11

Detailed structural map of the Huab area showing the faults and dykes that are constraint by field observation. The subdivision into three major tectonic zones is shown in the top bar. The location major faults and of Figures 2.4-4, 5.1-3, 5.3-1, 5.3-3, 5.3-6 and 5.3-7 is indicated. Map based on field observations and aerial photographs. For the coastal area additional data from Miller (1988) are used.

5.4 THE LOWER CRETACEOUS EVOLUTION OF THE HUAB-ETENDEKA REGION

A regional uplift accompanied by extensional block tectonics caused the erosion and/ or non-deposition of Triassic and Jurassic strata along the continental margin of north-western Namibia. The resulting pre-Cretaceous unconformity reveals a pronounced palaeo-relief that facilitated the sedimentation of the proximal deposits of the Krone Member and equivalent strata. Block faulting, partly controlled by inherited structures, continued and controlled deposition of the following units. Accommodation space for the aeolian units was created in hangingwall positions, whereas on uplifted blocks they are missing or reduced. In the coastal area, listric N-S striking faults were favoured, highlighting the special tectonic position of this area. This syn-Twyfelfontein Formation faulting phase documents a peak in tectonic activity that was followed by the onset of flood basalt generation. Block faulting with considerable vertical offsets continued but became spatially focused to the Skeleton Coast. There, a considerable uplift already commenced, which becomes apparent with the facies and thickness distribution of both, the basal sedimentary Etendeka Group and the following volcanic units.

All authors (Martin, 1975; Porada et al., 1996; Stollhofen, 1999) relate this "Skeleton Coast Uplift" to the intense continental rifting and final breakup of the South Atlantic in the Cretaceous. The genesis, however, is discussed controversial: Plume induced crustal doming, active up-warping of the asthenosphere due to an extraordinary hot mantle, extension induced mantel up-warping and a combinations of the three are considered. The close time relation of the uplift with the following extensive volcanism indicates a genetic coupling. Again, the volcanism might be directly coupled with the breakup and subsequent oceanisation. Hawkesworth et al. (1999) consider a major tectonic control on the Paraná-Etendeka volcanism and the involvement of a mantle plume. They deduce the tectonic control on magmatism from the consistently N-S trending dykes along the coast of Namibia and Brazil. Plume involvement might be reflected by the geochemical and isotopic similarity of the Damaraland Alkaline Complexes to the Tristan Mantle Plume

(Milner & Le Roex, 1996; Le Roex & Lanyon, 1998). In contrast to this Sheth (1999) generally denies plume involvement and emphasises the role of structurally controlled extension facilitating melt generation.

Evidence for a major structural control on Etendeka volcanism gives the long-lived structural evolution of the study area: As outlined in chapter 2.4, rift related north-south trending faults already caused a tectonic zonation during the Permian. During the Early Cretaceous this W-E directed tectonic zonation became more accentuated and tectonic zones became clearly confined by particular fault systems (chapter 5.3.1). This intensifying of the tectonic zonation immediately prior to the onset of Etendeka volcanism emphasises the prevalent structural control on rift evolution. As many dykes follow this rift related fault trends (N-S, NW-SE, W-E and SW-NE; chapter 4.2.8) evidence for further structural control on Etendeka volcanism is given. Furthermore, the rift topography comprising a rift shoulder (Skeleton Coast Uplift) and a rift valley became topographically well defined prior to the effusion of the Etendeka lavas (chapter 5.1.1.1), which confirms the influence of rifting on Etendeka volcanism.

Chapter 6: Conclusions in a Gondwana Context

Introduction

The Karoo megasequence in north-western Namibia mirrors both, the local tectonic settings of the corresponding depositories and the regional geodynamic evolution of southern Gondwana. Under this perspective the Huab-Goboboseb depository and the huge Paraná Basin represent the rift depression of the evolving South Atlantic rift. The Paraná Basin formed the main depocentre of this rift depression, whereas the Huab-Goboboseb depository is viewed as the marginal part of the latter. Therefore the term Paraná-Huab Basin is used here in order to ascribe both, marginal and central parts of the South Atlantic rift depression.

In contrast, the Waterberg-Erongo Basin is related to a transfer zone compartmentalising the South Atlantic rift. Both depositories (Huab-Goboboseb, Waterberg-Erongo depository) reveal a sequence development comparable with other Mesozoic depocentres in southern Gondwana (main Karoo Basin, Paraná Basin), strongly recommending a genetic coupling to Gondwana-wide climatic and tectonic signatures. Therefore, major regional gaps in the stratigraphic record potentially define correlative megasequences (cf. Hubbard, 1988) expressing successive extensional periods in the Gondwanan interior.

Factors controlling rifting in southern Gondwana

Mesozoic rifting in Gondwana is probably controlled by three factors: (1) Thermal anomalies, (2) impact tectonics and (3) inherited Pan-Gondwanan structures:

Following the model of Anderson (1982) regional rifting is essentially related to a high thermal gradient due to heat built-up under supercontinents. In this context the influence of mantle plumes seems to play an important role, but this is still a matter of debate (chapter 5.4). However the importance of thermal anomalies is documented by Burke (1976), who relates the location of hot spots to the junction of triple-rift systems. He identified over 100 grabens around the Atlantic ocean which are linked in a complex network to such triple-rift systems.

The second important factor is continental scale stresses induced by impact tectonics along the southern convergent margin of Gondwana (cf. de Wit and Ransome, 1992; Cobbold et al., 1992). The latter caused intracratonic E-W extension in southern Africa expressed by sinistral SW-NE and dextral NW-SE trending oblique-slip zones (Trouw & De Wit, 1999; Fig. 3.5-2).

The influence of Neoproterozoic structures is indicated by the distribution of major oblique-slip zones, which are preferentially located along Pan-African mobile belts around relative stable cratonic areas (Daly et al., 1989).

Evolution of the South Atlantic rift

Regional rifting events caused repeated northwardly directed marine incursions affecting the intracontinental rift valleys during sea level highstands associated with the Carboniferous Dwyka Group and the Permian Ecca Group (Martin, 1975; Visser, 1997; Bangert et al., in press). Contemporaneous, essentially E-W extensional activity is expressed by conjugate sets of NW-SE and NE-SW trending normal faults during Dwyka Group deposition (Stollhofen, 1999) followed by NNW-SSE and E-W trending fault sets during the Permian. The Paraná-Huab Basin essentially shows a corresponding tectonic zonation inflecting the orientation of facies belts corresponding to a step-wise deepening from the basin margins towards the basin centre. This zonation is emphasised by the structural subdivision of the Huab-Paraná Basin into numerous sub-basins (Chang et al., 1992) and spatially different subsidence patterns. The central Paraná Basin reveals a considerable amount of Carboniferous-Permian tectonic subsidence, followed by thermal cooling, which is clearly displayed by a backstripped subsidence curve from a representative well (cf. Zalán et al., 1990: Fig. 33-13). On the basin margins, however, rather condensed successions with an incomplete stratigraphic record developed. In the Huab area Early Permian Irati-Formation equivalents (Huab Formation) display intraformational erosion features and pedogenesis, while the majority of the Irati-Whitehill Sea extended deeply into the Gondwanan interior, reaching as far as central Brazil (Oelofsen, 1987; Williams, 1995). Following this marine incursion the sedimentary environment within the rift valley depression changed gradually into a freshwater lake. In the central Paraná Basin this gradual transition is documented by the up to 750 m thick Teresina Formation. The Teresina Formation reveals a sedimentation rate as high as 110 m/Ma, which is typical for rift basins (Da Cruz Cunha & França, 1994). Instead, the Huab area is lacking this transition due to a hiatal disconformity, which separates the marine Huab Formation from the lacustrine Gai-As Formation.

The change from a marine to a lacustrine environment probably corresponds to the successive cut-off from the marine realm towards the south, which was probably caused by large-scale uplift of the Argentinean Puna Highlands during the Cape-Ventana and San Rafael orogenies (cf. Veevers et al., 1994b; Porada et al., 1996). Associated continental scale stresses facilitated both, continuing subsidence in the rift basin centres and early rift shoulder uplift at the basin margins. In the Huab area the initial rift shoulder topography might be reflected by the eastwardly directed onlap of the Lower Gai-As Formation. Genetically, this early uplift seems to be coupled with an orogenic pulse in the Cape Fold Belt, dated at 258±2 Ma (Hälbich et al., 1983).

Evidence for contemporaneous, probably subduction related volcanism is derived from volcanic material found in numerous tuff-beds, which occur in the Dwyka and Ecca Group of South Africa and southern Namibia (Bangert et al., 1999) and in the Gai-As Formation of the Huab area (Stollhofen et al., 2000).

On the whole, some of the evolutionary steps of the Paraná Basin (cf. Klein, 1995) compare well with the development of African Karoo rifts involving sequential stages of extensional faulting, heating, and fault-controlled subsidence succeeded by subsidence related to thermal cooling and contraction.

Another period of uplift, but on a continental scale and much more pronounced, occurred during the Triassic and was succeeded by Early Jurassic volcanism and rifting, the latter culminating in the breakup of West (South America and Africa) from East Gondwana (India, Antarctica and Australia). This is well documented in southern Namibia where the Kalkrand Formation preserves a record of flood basalt volcanism and contemporaneous extensional tectonism (Stollhofen et al., 1998).

The activity of NNW-SSE and SW-NE trending faults continued with the SW-NE trending fault sets showing considerable involvement in basin development. In north-western Namibia the Goboboseb-Otjongundu and Waterberg-Erongo basins evolved along such SW-NE trending fault zones on either side of the Damaraland Uplift. A considerable sinistral strike-slip component of the Waterberg Fault system documents initial segmentation of the early southern South Atlantic rift zone.

Localised transpression is associated with the development of an angular unconformity between the Triassic and Jurassic, as described from the Waterberg-Omaruru Fault zone at Dinos Farm. This local feature seems to exaggerate more regional tectonostratigraphic signatures that are evident from similar facies architectures of the Triassic-Jurassic sequences in southern Africa. Again, the Triassic-Jurassic unconformity seems to be assigned to an orogenic pulse in the Cape Fold Belt, clearly underlining the far-field control of the Gondwanide orogenic belt.

A maximum of regional uplift preceded Early Cretaceous rifting and emplacement of the Etendeka volcanics in north-western Namibia, which is consistent with the predictive sequence of an active rift model (cf. Keen, 1985). This period of uplift is responsible for the development of the pronounced Karoo/Etendeka unconformity, defining the Karoo/Etendeka stratigraphic boundary. In the Huab area the anatomy of this unconformity demonstrates clearly strong tectonism appropriate to rift related doming including extensional fissures, vertical block faulting and block tilting along listric faults. The resulting palaeo-relief influenced facies and thickness distribution of the Lower Etendeka Group, as it is best viewed by the fluvial-alluvial deposits of the Krone Member. The Huab area provides a good insight in the spatial patterns of pre- and syn-Etendeka Group uplift. Distribution and palaeo-current vectors of the Krone Member together with the geometry of the Twyfelfontein erg indicate an uplift which is particular accentuated in the coastal area, referred to as the Skeleton Coast Uplift (Porada et al., 1996), accompanied by relative subsidence along the Huab Fault giving the Huab area a halfgraben shaped geometry. The NNW-SSE trending Skeleton Coast Uplift likely represents the South Atlantic rift shoulder, that became topographically well defined shortly prior to breakup. Contemporaneous activity of the SW-NE trending Huab Fault might be assigned to further segmentation of the evolving rift shoulder, as it already began in the Early Triassic along the Damaraland Uplift farther south.

During the Early Cretaceous extrusion of flood basalts the rift shoulder became successively more pronounced, which led to the erosion of the lavas at fault generated N-S trending scarps (Ward & Martin, 1987). Associated deposits have only been preserved in a few hangingwall traps nearby the Atlantic coast (Albin Ridge, Kharu-Gaiseb locality), suggesting that uplift continued after the main peak of flood volcanism. The corresponding N-S elongated rift valley depression is approved by the AMS (Anisotropy of Magnetic Susceptibility) inferred palaeo-flow patterns (Glen et al., 1997) and thickness distributions of lavas.

Migration of the main tectonic activity towards the coast is indicated by block tilting that progressively affected younger lavas on a section from the eastern Etendeka Plateau towards the west. Finally the onset of seafloor spreading prograded from the south to north. In the early Barrêmian the latitude of the Walvis Ridge was reached, which is verified by magnetic anomaly M4 (Rabinowitz, 1976) (Fig. 3.5-1). Segmentation of the evolving passive margins continued along the Waterberg-Omaruru and Autseib lineaments (Clemson et al., 1997) outlining the persistent control of early rift-related structures.

The control of basement fabrics on structural rift evolution is ambivalent as discordant and concordant structural relationships occur. For example, the Waterberg-Omaruru and Autseib lineaments trace the SW-NE fabric of the Damara inland branch concordantly. Similarly the N-S trending coastal faults follow concordantly the Kaoko belt. Discordant relationships with respect to Pan-African trends are reflected by the persistent Ponte Grossa High to the north and the Uruguay Shield-Windhoek Highlands to the south of the Huab Basin (Porada et al., 1996). It seems that distinct fault systems favour concordant trends in contrast to wide crustal depressions and uplifts that are not aligned to basement fabrics.

Summing up the whole Mesozoic rift evolution, it becomes apparent that a long phase of widespread diffuse rifting is eventually followed by a short phase of spatially focused rifting (cf. Praeg, 1999). The latter was associated with flood volcanism and ultimately succeeded by continental breakup with subsequent seafloor-spreading. In this sense the Mesozoic megasequences in north-western Namibia display a long-lived rift history of the early southern South Atlantic rift zone, that most of the time diffusely followed the axis of the Paraná Basin (Vidotti et al., 1995). With the Early Cretaceous flood basalt extrusion, rifting finally focused to the present day continental margin, whereas the rift branch penetrating the Paraná area became a failed rift (cf. Sibuet et al., 1984).

References

Anderson, D.L., 1982, Hotspots, polar wander, Mesozoic convection and the geoid: Nature, v. 297, p. 391-393.

Anderson, J.M., 1977, The biostratigraphy of the Permiam and the Triassic. Part 3. A review of Gondwana Permian palynology with particular reference to the northern Karoo Basin, South Africa: Memoir Botanical Survey South Africa, v. 41, p. 1-188.

Armstrong, R.A., Mc Dougall, I. and Watkins, R.T., 1998, Isotopic ages from Damaraland Complexes, Namibia, Abstract: Abstract Volume IAVCEI Symposium 1998.

Austin, J.A. and Uchupi, E., 1982, Continental-Oceanic crustal transition off Southwest Africa: Bulletin American Association Petroleum Geologists, v. 66, p. 1328-1347.

Bamford, M.K., 1998, Fossil woods of Karoo age deposits in South Africa and Namibia as an aid to biostratigraphical correlation (Abstract): Journal African Earth Sciences, v. 27, p. 16.

Bangert, B., Lorenz, V. and Armstrong, R.A., 1998, Bentonitic tuff horizons of the Permo-Carboniferous Dwyka Group in Southern Africa: volcanoclastic deposits as ideal time markers: Journal African Earth Sciences, v. 27, p. 18-19.

Bangert, B., Stollhofen, H., Geiger, M. and Lorenz, V., in press, High resolution tephrostratigraphy, fossil record and age of Carboniferous-Permian glaciomarine mudstones within the Dwyka Group of southern Namibia: Communications Geological Survey Namibia.

Bangert, B., Stollhofen, H., Lorenz, V. and Armstrong, R., 1999, The geochronolgy and significance of ash-fall tuffs in the glaciogenic Carboniferous-Permian Dwyka Group of Namibia and South Africa: Journal African Earth Sciences, Special Issue, v. 29/1.

Barberena, M.C., Araujo, D.C., Lavina, E.L. and Faccini, U.F., 1991, The evidence for close palaeofaunistic affinity between South America and Africa, as indicated by late Permian and Triassic tetrapods: 7th International Gondwana Symposium, São Paulo 1988, Proceedings, p. 455-467.

Bigarella, J.J., 1970, Continental drift and paleocurrent analysis, *in* Haughton, S.H., ed., I.U.G.S., 2nd Symposium on Gondwana Stratigraphy and Palaeontology: Pretoria, Council for Scientific and Industrial Research, p. 73-97.

Blümel, W.D., Emmermann, R. and Hüser, K., 1979, Der Erongo. Geowissenschaftliche Beschreibung und Deutung eines südwestafrikanischen Vulkankomplexes: Wissenschaftliche Forschung Südwestafrika, v. 16, p. 1-107.

Botha, B.J.V. and Hodgson, F.D.I., 1976, Karoo dolerites in northwestern Damaraland: Transactions Geological Society South Africa, v. 79, p. 186-190.

Burke, K., 1976, Development of graben associated with initial ruptures of the Atlantic Ocean, *in* Bott, M.H.P., ed., Sedimentary Basins of Continental Margins and Cratons: Tectonophysics, Volume 36 (1-3), p. 93-112.

Bystrow, A.P., 1938, *Dvinosaurus* als neotenische Form der Stegocephalen: Acta Zoologica Stockholm, v. 19, p. 1-87.

Cairncross, B., Groenewald, G.H., Rubidge, B.S. and Von Brunn, V., 1995, Guidebook Geocongress `95: Karoo sedimentology and palaeontology: Johannesburg, Geological Society South Africa, p. 1-48.

Chang, H.K., Kowsmann, R.O., Figueiredo, A.M.F. and Bender, A.A., 1992, Tectonics and stratigraphy of the East Brazil Rift system: an overview, *in* Ziegler, P.A., ed., Geodynamics of Rifting, Volume II. Case History Studies on Rifts: North and South America and Africa: Tectonophysics, Volume 213, p. 97-138.

Clemson, J., Cartwright, J. and Booth, J., 1997, Structural segmentation and the influence of basement structure on the Namibian passive margin: Journal Geological Society, London, v. 154, p. 477-482.

Cloos, H., 1911, Geologie des Erongo im Hererolande (Geologische Beobachtungen in Südafrika, Teil II): Königlich Preußische Geologische Landesanstalt zu Berlin, Beiträge zur Erforschung der Deutschen Schutzgebiete, v. 3, p. 1-84.

Cobbold, P.R., Gapais, D., Rossello, E.R., Milani, E.J. and Szatmari, P., 1992, Permo-Triassic intracontinental deformation in SW Gondwana, *in* De Wit, M.J. and Ransome, I.G.D., eds., Inversion tectonics of the Cape Fold Belt, Karoo and Cretaceous Basins of Southern Africa: Rotterdam, Balkema, p. 23-26.

Compston, W and Williams, L.S., 1984, U-Pb geochronology of zircons from lunar breccia 73217 using a sensitive high mass-resolution ion microprobe: Journal Geophysical Research (Supplement), v. 89, p. B525-534.

Compston, W., Williams, I.S., Kirschvink, J.L., Zhang, Z. and Zhang, M.G., 1992, Zircon U-Pb ages for the Early Cambrian time-scale: Journal Geological Society London, v. 149, p. 171-184.

Da Cruz Cunha, P.R. and França, A.B., 1994, Estudo das Taxas de Sedimentação das Formações Teresina e Rio do Rasto - Bacia do Paraná [Study of the Rates of Sedimentation of Teresina and Rio do Rasto Formations - Paraná Basin]: Boletim Geociências Petrobrás, v. 8 (2/4), p. 347-359.

Daly, M.C., Chorowicz, J. and Fairhead, J.D., 1989, Rift basin evolution in Africa: the influence of reactivated steep basement sheer zones, *in* Cooper, M.A. and Williams, G.D., eds., Inversion Tectonics, Special Publication Geological Society London, Volume 44, p. 309-334.

Davies, S.L. and Elliot, T., 1994, The gamma-ray response of high resolution key surfaces and system tracts; examples from the Upper Carbonifereous Clare Basin, western Ireland, *in* Johnson, S.D., ed., High resolution sequence stratigraphy: innovations and applications: Abstract volume of the Liverpool sequence stratigraphy conference, March, 1994, p. 77-81.

De Wit, M.J., Jeffery, M., Bergh, H. and Nicolaysen, L., 1988, Geological map of sectors of Gondwana: 2 sheets, 1:10 000 000: American Association Petroleum Geologists.

De Wit, M.J. and Ransome, I.G.D., 1992, Regional inversion tectonics along the southern margin of Gondwana, *in* De Wit, M.J. and Ransome, I.G.D., eds., Inversion tectonics of the Cape Fold Belt, Karoo and Cretaceous Basins of southern Africa: Rotterdam, Balkema, p. 15-21.

Dingle, R.V., 1992/1993, Structural and sedimentary development of the continental margin off Southwestern Africa: Communications Geological Survey Namibia, v. 8, p. 35-43.

Duke, W.L., Arnott, R.W.C. and Cheel, R.J., 1991, Shelf sandstones and hummocky cross-stratification: new insight on a stormy debate: Geology, v. 19, p. 625-628.

Duncan, A.R. and Armstrong, R.A., 1990a, MORB-related dolerites associated with the Etendeka volcanics, Northwestern Namibia: Abstract Geological Society South Africa Geocongress 1990, Cape Town, p. 143-146.

Duncan, A.R., Newton, S.R., Van den Berg, C. and Reid, D.L., 1989, Geochemistry and petrology of dolerite sills in the Huab River Valley, Damaraland, north-western Namibia: Communications Geological Survey Namibia, v. 5, p. 5-17.

Duncan, R.A., Hooper, P.R., Rehacek, J., Marsh, J.S. and Duncan, A.R., 1997, The timing and duration of the Karoo igneous event, southern Gondwana: Journal Geophysical Research, v. 102, p. 18127-18138.

Ellenberger, P., 1970, Les niveaux paleontologiques de premiere apparation des Mammiferes primordiaux en Afrique du Sud et leur ichnologie: Establissement de zones stratgraphiques détaillées dans le Stormberg du Lesotho (Afrique du Sud) (Trias superieur a Jurassique), *in* Haughton, S.H., ed., IUGS, 2nd Symposium on Gondwana Stratigraphy and Palaeontology, Pretoria, p. 343-370.

Eriksson, P.G., 1979, Mesozoic sheetflow and playa sediments of the Clarens Formation in the Kamberg area of the Natal Drakensberg: Transactions Geological Society South Africa, v. 82, p. 251-258.

Erlank, A.J., Marsh, J.S., Duncan, A.R., Miller, R.M., Hawkesworth, S.C., Betton, B.J. and Rex, D.C., 1984, Geochemistry and petrogenesis of the Etendeka volcanic rocks from South West Africa/Namibia, *in* Erlank, A.J., ed., Petrogenesis of the volcanic rocks of the Karoo Province: Special Publication Geological Society South Africa, Volume 13, p. 195-245.

Ewart, A., Milner, S.C., Armstrong, R.A. and Duncan, A.R., 1998a, Etendeka volcanism of the Goboboseb mountains and messum igneous complex, Namibia. Part II: Geochemical evidence of early Cretaceous Tristan Plume Melts and the role of crustal contamination in the Paraná-Etendeka CFB: Journal Petrology, v. 39 (2), p. 191-225.

Fauché, J., Bate, K.J. and Van der Merwe, R., 1992, Plate tectonic setting of the Mesozoic Basins, southern offshore, South Africa: A review, *in* De Wit, M.J. and Ransonme, I.G.D., eds., Inversion tectonics of the Cape Fold Belt, Karoo and Cretaceous Basins of southern Africa: Rotterdam, Balkema, p. 33-45.

França, A.B., Milani, E.J., Schneider, R.L., Lopez-Paulsen, M., Suãrez-Soruco, R., Santa-Ana, A., Wiens, F., Ferreiro, O., Rossello, E.A., Binaucci, H.H., Flores, R.E.A., Vistalli, M.C., Fernandez-Seveso, F., Fuenzalida, R.P. and Munoz, N., 1995, Phanerozoic correlation in southern South America, *in* Tankard, A.J., Suarez-Soruco, R. and Welsink, H.J., eds., Petroleum basins of South America: American Association Petroleum Geologists, Volume 62, p. 129-161.

França, A.B. and Potter, P.E., 1988, Estratigrafia, ambiente depositional e análise reservatório do Grupo Itararé (Permocarbonifero), Bacia do Paraná (Parte 1): Boletim Geociências Petrobrás, v. 2 (2/4), p. 147-191.

Freytet, P. and Plaziat, J.C., 1982, Continental Carbonate Sedimentation and Pedogenesis - Late Cretaceous and Early Tertiary of Southern France: Contributions Sedimentology, v. 12, p. 1-213.

Friend, P.F., Slater, M.J. and Williams, R.C., 1979, Vertical and lateral building of river sandstone bodies, Ebro Basin, Spain: Journal Geological Society London, v. 136, p. 39-46.

Fürsich, F.T. and Mayr H., 1981, Non-marine Rhizocorallium (trace fossil) from the Upper Freshwater Molasse (Upper Miocene) of southern Germany: Neues Jahrbuch Geologie Paläontologie, Monatshefte, v. 1981/6, p. 321-333.

Gerschütz, S., 1996, Geology, volcanology, and petrogenesis of the Kalkrand Basalt Formation and the Keetmanshop dolerite complex, southern Namibia [Ph.D. thesis]: Würzburg, Germany, Universität Würzburg.

Gerschütz, S., Stanisstreet, I., Stollhofen, H. and Lorenz, V., 1995, Karoo flood basalts and interleaved lake sediments of the Mariental area, Namibia: Extended Abstract, Centennial geocongress, Geological Society South Africa, v. 2, p. 769-772.

Gevers, T.W., 1936, The Etjo Beds of Northern Hereroland, South West Africa: Transactions Geological Society South Africa, v. 39, p. 317-329.

Gibbs, A.D., 1990, Linked fault families in basin formation: Journal Structural Geology, v. 12, p. 795-803.

Glen, J.M.E., Renne, P.R., Milner, S.C. and Coe, R.S., 1997, Magma flow inferred from anisotropy of magnetic susceptibility in the coastal Parana-Etendeka igneous province: Evidence for rifting before flood volcanism: Geology, v. 25, p. 1131-1134.

Gradstein, F.M., Agterberg, F.P., Ogg, J.O., Hardenbol, L., Van Veen, P., Thierry, J. and Huang, Z., 1994, A Mesozoic time scale: Journal Geophysical Research, v. 99, p. 24051-24074.

Gradstein, F.M. and Ogg, J., 1996, A Phanerozoic time scale: Episodes, v. 19 (1&2), p. 6-7.

Grill, H., 1997, The Permo-Carboniferous glacial to marine Karoo record in southern Namibia: sedimentary facies and sequence stratigraphy: Beringeria, Würzburger geowissenschaftliche Mitteilungen, v. 19, p. 3-98.

Gunthorpe, R.J., 1987, Final report, Samoa and Tevrede prospecting grants (M46/3/1240 and 1241), Otjiwarongo District, South West Africa/Namibia: Unpublished Report, Tsumeb Corp. Ltd., p. 1-14.

Gürich, G., 1926a, Über fossile Fährten im Etjo Sandstein S.W.A.: Jahrbuch Paläontologie.

Gürich, G., 1926b, Über Saurier Fährten aus dem Etjo-Sandstein von Südwestafrika: Paläontologische Zeitschrift, v. 8, p. 112-120.

Hälbich, I.W., Fitch, F.J. and Miller, J.A., 1983, Dating the Cape orogeny, *in* Söhnge, A.P.G. and Hälbich, I.W., eds., Geodynamics of the Cape Fold Belt: Geological Society South Africa, Special Publication, Volume 12, p. 149-164.

Harland, W.B., Armstrong, R.L., Cox, A.V., Craig, L.E., Smith, A.G. and Smith, D.G., 1990, A geological time scale: Camebridge, Camebridge University Press, 1-263 p.

Hawkesworth, C., Kelley, S., Turner, S., Le Roex, A. and Storey, B., 1999, Mantle processes during Gondwana break-up and dispersal: Journal African Earth Sciences, v. 28, p. 239-261.

Hawkesworth, C.J., Gallagher, K., Kelley, S., Mantovani, M., Peate, D.P., Regelous, M. and Rogers, N.W., 1992, Paraná magmatism and the opening of the South Atlantic, *in* Storey, B.C., Alabaster, T. and Pankhurst, R.J., eds., Magmatism and the Causes of Continental Break-up: Special Publication Geological Society London, Volume 68, p. 221-240.

Heath, D.C., 1972, Die Geologie van die Sisteem Karoo in die Gebied Mariental-Asab, Suidwes-Afrika: Memoir Geological Survey South Africa, v. 61, p. 1-44.

Hegenberger, W., 1988, Karoo sediments of the Erongo Mountains, their environmental setting and correlation: Communications Geological Survey S.W. Africa/Namibia, v. 4, p. 51-57.

Heward, A.P., 1981, A review of wave-dominated clastic shoreline deposits: Earth Science Reviews, v. 17, p. 223-276.

Hodgson, F.D.I., 1970, The geology of the Karroo System in the southern Kaokoveld, South West Africa, *in* Haughton, S.H., ed., I.U.G.S., 2nd Gondwana Symposium on Gondwana Stratigraphy and Palaeontology, Proceedings: Pretoria, p. 233-240.

Hodgson, F.D.I., 1972, The geology of the Brandberg-Aba Huab area, South West Africa [PhD thesis]: Stellenbosch, University Orange Free State.

Hodgson, F.D.I. and Botha, B.J.V., 1975, The Karroo sediments in the vicinity of Doros, South West Africa: Annotations Geological Survey South Africa, v. 10, p. 49-56.

Hoffmann, P.F., 1991, Did the breakout of Laurentia turn Gondwana inside-out?: Science, v. 252, p. 1409-1412.

Holzförster, F., 2000, Sedimentology, stratigraphy and synsedimentary tectonics of the Karoo Supergroup in the Huab and Waterberg-Erongo areas, N-Namibia [Ph.D. thesis]: Würzburg, Germany, Universität Würzburg.

Holzförster, F. and Stollhofen, H., in press, Early Permian Deposits of the Huab Area, NW Namibia: A Continental to Marine Transition: Communications Geological Survey Namibia.

Holzförster, F., Stollhofen, H., Lorenz, V. and Stanistreet, I.G., 1998, The Waterberg basin in Central Namibia: Transfer fault activity during early South Atlantic rift evolution: Journal African Earth Sciences, v. 27, p. 116-117.

Holzförster, F., Stollhofen, H. and Stanistreet, I.G., 1999, Lithostratigraphy and depositional environments in the Waterberg-Erongo area, central Namibia, and correlation with the main Karoo Basin, South Africa: Journal African Earth Sciences, v. 29, p. 105-123.

Hopson, J.A. and Reif, W.E., 1981, The status of *Archaeodon reuningi von Huene*, a supposed late Triassic mammal from southern Africa: Neues Jahrbuch Geologie Paläontologie, Monatshefte, v. 1981, p. 307-310.

Horsthemke, E., 1992, Fazies der Karoosedimente in der Huabregion, Damaraland, NW-Namibia: Göttinger Arbeiten Geologie Paläontologie, v. 55, p. 1-102.

Horsthemke, E., Ledendecker, S. and Porada, H., 1990, Depositional environments and stratigraphic correlation of the Karoo sequence in northwestern Damaraland: Communications Geological Survey South West Africa/Namibia, v. 6, p. 63-73.

Hubbard, R.J., 1988, Age and Significance of Sequence Boundaries on Jurassic and Early Cretaceous Rifted Continental Margins: Bulletin American Association Petroleum Geologists, v. 72, p. 49-72.

Huene Von, F., 1925, Ausgedehnte Karroo-Komplexe mit Fossilführung im nordöstlichen Südwestafrika: Centralblatt Mineralogie, Geologie Paläontologie, Abteilung B: Geologie Paläontologie, v. 25, p. 151-156.

Jerram, D., Mountney, N., Holzförster, F. and Stollhofen, H., 1999b, Internal stratigraphic relationships in the Etendeka Group in the Huab Basin, NW Namibia: understanding the onset of flood volcanism: Journal Geodynamics, v. 28, p. 393-418.

Jerram, D.A., Mountney, N.P. and Stollhofen, H., 1999a, Facies architecture of the Etjo Sandstone Formation and its interaction with the basal Etendeka flood basalts of NW Namibia: Implications for offshore analogues, *in* Cameron, N., Bate, R. and Clure, V., eds., The Oil and Gas habitats of the South Atlantic: Special Publication Geological Society London, Volume 153, p. 367-380.

Johnson, M.R., Van Vuuren, C.J., Hegenberger, W.F., Key, R. and Shoko, U., 1996, Stratigraphy of the Karoo Supergroup in southern Africa: an overview: Journal African Earth Sciences Middle East, v. 23, p. 3-15.

Keen, C.E., 1985, The dynamics of rifting: deformation of the lithosphere by active and passive driving forces: Geophysical Journal Royal Astronomical Society, v. 80, p. 95-120.

Keyser, A.W., 1973, A new Triassic vertebrate fauna from South West Africa: Palaeontologia Africana, v. 16, p. 1-15.

Keyser, A.W., 1978, A new Bauriamorph from the Omingonde Formation (Middle Triassic) of South West Africa: Palaeontologia Africana, v. 21, p. 177.

Kingsley, C.S., 1985, Sedimentological analysis of the Ecca Sequence in the Kalahari Basin, South West Africa/Namibia: Unpublished Open File Report, CDM Mineral Surveys, Namibia, p. 1-39.

Kitching, J.W., 1995, Biostratigraphy of the *Cynognathus* Assemblage Zone, *in* Rubidge, B.S., ed., Biostratigraphy of the Beaufort Group (Karoo Supergroup): South African Committee for Stratigraphy (SACS), Biostratigraphic Series 1, p. 40-45.

Kitching, J.W. and Raath, M.A., 1984, Fossils from the Elliot and Clarens Formations (Karoo Sequence) of the Northeastern Cape, Orange Free State and Lesotho, and a suggested biozonation based on tetrapods: Palaeontologia Africana, v. 25, p. 111-125.

Klein, G.D., 1995, Intracratonic basins, *in* Busby, C.J. and Ingersoll, R.V., eds., Tectonics of sedimentary basins: Cambridge, Blackwell, p. 459-478.

Kocurek, G., 1981, Significance of interdune deposits and bounding surfaces in aeolian dune sands: Sedimentology, v. 28, p. 753-780.

Krause, C., 1913, Über die Geologie des Kaokofeldes in Deutsch-Südwestafrika: Zeitschrift praktischer Geologen.

Le Roux, A.P., Dick, H.J.B., Erlank, A.J., Reid, A.M., Frey, F.A. and Hart, S.R., 1983, Geochemistry, mineralogy and petrogenesis of lava erupted along the Southwest Indian Bridge between Bouvet tripple junction and 11 degrees east: Journal Petrology, v. 11, p. 267-318.

Le Roux, A.P. and Lanyon, R., 1998, Isotope and trace element geochemistry of Cretaceous Damaraland lamprophyres and carbonatites, northwestern Namibia: evidence for plume lithosphere interactions: Journal Petrology, v. 39, p. 1117-1146.

Ledendecker, S., 1992, Stratigraphie der Karoosedimente der Huabregion (NW-Namibia) und deren Korrelation mit zeitäquivalenten Sedimenten des Paranabeckens (Südamerika) und des großen Karoobeckens (Südafrika) unter besonderer Berücksichtigung der überregionalen geodynamischen und klimatischen Entwicklung Westgondwanas: Göttinger Arbeiten Geologie Paläontologie, v. 54, p. 1-87.

Löffler, T. and Porada, H., 1998, Tontüten "Mud curls" aus der Etjo Formation am Gr. Waterberg (Namibia) und aus dem Germanischen Buntsandstein - Über die Erhaltungsbedingungen von aufgerollten Trockenriß-Segmenten [Mud curls ("Tontüten") from the Etjo Formation at the Gr. Waterberg (Namibia) and the German "Buntsandstein" - About the preservation of upcurled desiccation segments]: Freiberger Forschungshefte, v. C 475, p. 201-221.

Logan, B.W., Rezak, R. and Ginsburg, R.N., 1964, Classification and environmental significance of algal stromatolites: Journal Geology, v. 72, p. 68-83.

Ludwig, K.R., 1999, Isoplot/Ex version 2.00: A Geochronological Toolkit for Microsoft Excel: Berkeley Geochronology Center Special Publication, v. 1a, p. 1-46.

Marsh, J.S., 1987, Basalt geochemistry and tectonic discrimination within continental flood basalt provinces: Journal Volcanology Geothermal Research, v. 32, p. 35-49.

Martin, H., 1953, Notes on the Dwyka succession and on some pre-Dwyka valleys in South-West Africa: Transactions Geological Society South Africa, v. 56, p. 37-41.

Martin, H., 1961a, The hypothesis of continental drift in the light of recent advances of geological knowledge in Brazil and South West Africa: A.L. du Toit Memorial Lecture,7: Transactions Geological Society South Africa, Annex. to vol., v. 64, p. 1-47.

Martin, H., 1961b, Part III: The geology and distribution of the Karoo System in South West Africa, *in* Brandt, J.W., Martin, H. and Kirchner, J.G., eds., Interim report of the coal commission of South West Africa: Windhoek, unpublished, p. 17-86.

Martin, H., 1973, Palaeozoic, Mesozoic and Cenozoic deposits on the coast of South West Africa, *in* Blant, G., ed., Sedimentary basins of the African Coast 2nd Part - South and East Coast. Symposium Association African Geological Surveys, Volume 2: IUGS Commission, Marine Geology: Montreal, p. 7-15.

Martin, H., 1975, Structural and palaeogeographical evidence for an Upper Palaeozoic sea between Southern Africa and South America, *in* Campbell, K.W.S., ed., I.U.G.S., 3rd Gondwana Symposium on Gondwana Stratigraphy and Palaeontology, Proceedings: Canberra, p. 37-51.

Martin, H., 1981, The Late Palaeozoic Dwyka Group of the Southern Kalahari Basin in Namibia and Botswana and the subglacial valleys of the Kaokofeld in Namibia, *in* Hambrey, M.J. and Harland W.B., eds., Earths Pre-Pleistocene Glacial Record: Cambridge, Cambridge University Press, p. 61-66.

Martin, H., 1982, Die Trias im südlichen Afrika: Geologische Rundschau, v. 71, p. 937-947.

Milani, E.J., França, A.B. and Medeiros, R.Á., 1998, Geology of the southeastern Paraná Basin: With Emphasis on Reservoirs and Source Rocks; Field Trip 11-14 November 1998, The 1998 AAPG International Conference & Exhibition: Petroleum Geology in a Changing World: Rio de Janeiro, p. 1-20.

Milani, E.J., França, A.B. and Schneider, R.L., 1994, Bacio do Paraná: Boletim Geociências Petrobrás.

Milani, E.J. and Zalan, P.V., 1999, An outline of the geology and petroleum systems of the Paleozoic interior basins of South America: Episodes, v. 22(3), p. 199-205.

Miller, R.M., 1980, Geology of a portion of central Damaraland, South West Africa/Namibia: Memoir Geological Survey South Africa, S.W. African Series, v. 6, p. 1-78.

Miller, R.M., 1983, The Pan-African orogen in South West Africa/Namibia, *in* Miller, R.M., ed., Evolution of the Damara Orogen of South West Africa/Namibia: Special Publication Geological Society South Africa, Volume 11, p. 431-512.

Miller, R.M., 1988, Geological map of Namibia, Sheet 2013 Cape Cross, Scale 1:250 000: Geological Survey South West Africa/Namibia.

Miller, R.M., 1997, The Ovambo basin of northern Namibia, *in* Selley, R.C., ed., African Basins, Sedimentary Basins of the World, 3: Amsterdam, Elsevier, p. 237-268.

Miller, R.M. and Schalk, K.E.L., 1980, South West Africa/Namibia geological map 1:1000 000: Geological Survey Namibia.

Milner, S.C., 1988, The geology and geochemistry of the Etendeka Formation quartz latites, Namibia [PhD thesis thesis]: Cape Town, South Africa, University of Cape Town.

Milner, S.C., 1997, Geological map of Namibia, Sheet 2014 Omaruru, Scale 1:200 000: Geological Survey Namibia.

Milner, S.C. and Duncan, A.R., 1987, Geochemical characterisation of quartz latite units in the Etendeka Formation: Communications Geological Survey S.W.A./Namibia, v. 3, p. 83-90.

Milner, S.C., Duncan, A.R., Ewart, A. and Marsh, J.S., 1994, Promotion of the Etendeka Formation to Group status: A new integrated stratigraphy: Communications Geological Survey Namibia, v. 9, p. 5-11.

Milner, S.C., Duncan, A.R. and Ewart, A.R., 1992, Quartz latite rheoignimbrite flows of the Etendeka Formation, north-western Namibia: Bulletin Volcanology, v. 54, p. 200-219.

Milner, S.C., Duncan, A.R., Whittingham, A.M. and Ewart, A., 1995a, Trans-Atlantic correlation of eruptive sequences and individual silicic volcanic units within the Paraná-Etendeka igneous province: Journal Volcanology Geothermal Research, v. 69, p. 137-157.

Milner, S.C. and Ewart, A., 1989, The geology of the Goboboseb Mountain volcanics and their relationship to the Messum Complex: Communications Geological Survey Namibia, v. 5, p. 31-40.

Milner, S.C. and Le Roex, A.P., 1996, Isotope characteristics of the Okenyenya igneous complex, northwestern Namibia: constraints on the composition of the early Tristan plume and the origin of EM 1 mantle component: Earth Planetary Science Letters, v. 141, p. 277-291.

Milner, S.C., Le Roex, A.P. and O´Connor, J.M., 1995b, Age of Mesozoic igneous rocks in northwestern Namibia, and their relationship to continental breakup: Journal Geological Society London, v. 152, p. 97-104.

Mountney, N., Howell, J., Flint, S. and Jerram, D., 1998, Aeolian and alluvial deposition within the Mesozoic Etjo Sandstone Formation, northwest Namibia: Journal African Earth Sciences, v. 27, p. 175-192.

Mühlmann H. (coord.), 1983, Integração dos dados geológicos e geofisicos da Bacia do Paraná (mapas e seções estratigráficas): Petrobrás Internal report.

Oelofsen, B.W., 1981, An anatomical and systematical study of the family Mesosauridae (Reptilia; Proganosauria) with special reference to its associated fauna and palaeoecological environment in the Whitehill Sea [PhD thesis], University of Stellenbosch, South Africa.

Oelofsen, B.W., 1987, The Biostratigraphy and Fossils of the Whitehill and Irati Shale Formations of the Karoo and Paraná basins, *in* Mc Kenzie, G.D., ed., Gondwana six: Stratigraphy, sedimentology and palaeontology: Geophysical Monograph AGU, Volume 41, p. 131-138.

Oelofsen, B.W. and Araujo, D.C., 1983, Palaeoecological implications of the distribution of Mesosaurid Reptiles in the Permian Irati Sea (Paraná Basin), South America: Revista Brasileira Geociências, v. 13, p. 1-6.

Oelofsen, B.W. and Araujo, D.C., 1987, Mesosaurus tenuidens and Stereosternum tumidum from the Permian Gondwana of both Southern Africa and South America: South African Journal Science, v. 83, p. 370-372.

Olsen, P.E. and Galton, P.M., 1984, A review of the reptile and amphibia assemblage from the Stormberg of Southern Africa, with special emphasis on the footprints and the age of the Stormberg: Palaeontologia Africana, v. 25, p. 87-110.

Osborne, M.A., 1985, Otjiwarongo Karoo Basin coal project: Phase 1 (grants M46/3/1551, 1550, and 1549): Unpublished Report, Gold Fields Prospecting Company, p. 1-11.

Paces, J.B. and Miller, J.D., 1989, Precise U-Pb ages of Duluth Complex and related mafic intrusions, Northeastern Minnesota: Geochronological insights to physical, petrogenic, paleomagnetic and tectonomagmatic processes associated with the 1.1 Ga midcontinent rift system: Journal Geophysical Research, v. 98b, p. 13997-14013.

Peate, P.W., 1997, The Parana-Etendeka Province, *in* Mahoney, J.J. and Coofin, M.F., eds., Large Igneous Provinces: Continental, Oceanic and Planetary Flood Volcanism: American Geophysical Union, Geophysical Monograph, Volume 100, p. 217-245.

Pettijohn, F.P., Potter, P.E. and Siever, R., 1973, Sand and sandstone: Berlin, Springer-Verlag, p. 1-618 .

Pickford, M., 1995, Karoo Supergroup Palaeontology of Namibia and brief description of a Thecodont from Omingonde: Palaeontologia Africana, v. 32, p. 51-66.

Porada, H., 1979, The Damara-Ribeira orogen of the Pan-African-Brasiliano cycle in Namibia (Southwest Africa) and Brazil as interpreted in terms of continental collision: Tectonophysics, v. 57, p. 237-265.

Porada, H., 1989, Pan-African rifting and orogenesis in south to equatorial Africa and eastern Brazil: Precambrian Research, v. 44, p. 103-136.

Porada, H., Löffler, T., Horsthemke, E., Ledendecker, S. and Martin, H., 1996, Facies and palaeo-environmental trends of northern Namibian Karoo sediments in relation to West Gondwanaland palaeogeography, Gondwana nine (Proceedings 9th International Gondwana Symposium, Hyperabad/India), Volume 2, Oxford & IBH Publishing, p. 1101-1114.

Praeg, D., 1999, Non-Localised Modes of Pangean Extension: Lithospheric Controls on Permian and Triassic Basins NW Europe, *in* University of Edinburgh, ed., 38th Annual Meeting of the British Sedimentological Research Group: Edinburgh, Departement of Geology and Geophysics, University of Edinburgh, Scotland.

Raath, M.A., 1980, The theropod dinosaur Syntarsus (Saurischia; Podokesauridae) discovered in South Africa: South African Journal Science, v. 76 (8), p. 375-376.

Rabinowitz, P.D., 1976, Geophysical study of the continental margin of southern Africa: Bulletin Geological Society America, v. 87, p. 1643-1653.

Renne, P.R., Glen, J.M., Milner, S.C. and Duncan, A.R., 1996, Age of Etendeka flood volcanism and associated intrusions in southwestern Africa: Geology, v. 24, p. 659-662.

Reuning, E., 1923, Der Intrusionsverband der Granite des Mittleren Hererolandes und des angrenzenden Küstengebietes in Südwestafrika: Geologische Rundschau, v. 14, p. 232-239.

Reuning, E. and Martin, H., 1957, Die Prä-Karroo-Landschaft, die Karroo-Sedimente und Karroo-Eruptivgesteine des südlichen Kaokofeldes: Neues Jahrbuch, Mineralogie Geologie Paläontologie, v. 91, p. 193-212.

Reuning, E. and Von Huene, D., 1925, Fossilführende Karrooschichten im nördlichen S.W.A.: Neues Jahrbuch Mineralogie, Paläontologie Geologie, v. 52, Abt. B., Beilage Band, p. 94-122.

Richter, D.K. and Füchtbauer, H., 1981, Merkmale und Genese von Breccien und ihre Bedeutung im Mesozoikum von Hydra (Griechenland) [Properties and origin of breccias and their significance in the Mesozoic limestones of Hydra (Greece)]: Zeutschrift deutche geologische Gesellschaft, v. 132, p. 451-501.

Rocha-Campos, A.C., Cordani, U.G., Kawashita, K., Sonoki, H.M. and Sonoki, I.K., 1988, Age of the Paraná- Flood Volcanism, *in* Piccirillo, E.M. and Melfi, A.J., eds., The Mesozoic Flood Volcanism of the Paraná Basin: Petrogenetic and Geophysical Aspects: São Paulo, p. 22-25.

Roesener, H. and Schreuder, C.P., 1998, Uranium, The Mineral Resources of Namibia, Geological Survey Namibia: Windhoek, p. 7.1-1--7.1-55.

Rohn, R., 1994, Evolução ambiental da Bacia do Paraná durante o neopermiano no leste de Santa Catarina e do Paraná [PhD thesis]: São Paulo, Universidade de Sao Paulo, Brazil.

Rohn, R., Perinotto, J.A.J., Fulfaro, V.J., Saad, A.R. and Simões, M.G., 1995, On the significance of the *Pinzonella neotropica* assemblage (Upper Permian) for the Paraná Basin - Brazil and Paraguay: VI Simpósio Sul-Brasileiro de Geologia: Boletim Resumos Expandidos, p. 260-261.

Rubidge, B.S., Johnson, M.R., Kitching, J.W., Smith, R.M.H., Keyser, A.W. and Groenewald, B.S., 1995, An introduction to the biozonation of the Beaufort Group, *in* Rubidge, B.S., ed., Biostratigraphy of the Beaufort Group (Karoo Supergroup): South African Committee for Stratigraphy (SACS), Biostratigraphic Series 1, p. 1-45.

Rust, B.R., 1978, A classification of alluvial channel systems, *in* Miall, A.D., ed., Fluvial Sedimentology: Memoir Canadian Society Petroleum Geologists, Volume 5, p. 187-198.

Rust, B.R. and Koster, E.H., 1984, Coarse alluvial deposits, *in* Walker, R.G., ed., Facies Models: Geoscience Canada, Reprint Series 1, p. 53-69.

Santos, P.R.D., Rocha-Campos, A.C. and Canuto, J.R., 1996, Patterns of late Palaeozoic deglaciation in the Paraná Basin, Brazil: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 125, p. 165-184.

Sheth, H.C., 1999, Flood basalts and large igneous provinces from deep mantle plumes: fact, fiction and fallacy: Tectonophysics, v. 311, p. 1-29.

Sibuet, J.-C., Hay, W.W., Prunier, A., Montadert, L., Hinz, K. and Fritsch, J., 1984, Early evolution of the South Atlantic: Role of the rifting phase: Initial Reports Deep Sea Drilling Project, v. 75, p. 469-481.

Siedner, G. and Mitchell, J.A., 1976, Episodic Mesozoic volcanism in Namibia and Brazil: K-Ar-isochron study bearing on the opening of the South Atlantic: Earth Planetary Science Letters, v. 30, p. 292-302.

Smith, R.M.H., Eriksson, P.G. and Botha, W.J., 1993, A review of the stratigraphy and sedimentary environments of the Karoo-aged basins of Southern Africa: Journal African Earth Sciences, v. 16, p. 143-169.

Stahl, A., 1932, Die Verbreitung der Karru im nordwestlichen Kaokoveld (Südwest-Afrika): Zeitschrift deutsche geologische Gesellschaft, v. 84, p. 158-173.

Stanistreet, I.G. and Stollhofen, H., 1999, Onshore equivalents of the main Kudu gas reservoir in Namibia, *in* Cameron, N., Bate, R. and Clure, V., eds., The Oil and Gas habitats of the South Atlantic: Special Publication Geological Society London, Volume 153 , p. 345-365.

Stollhofen, H., 1999, Karoo Synrift-Sedimentation und ihre tektonische Kontrolle am entstehenden Kontinentalrand Namibias [Karoo synrift-deposition and its tectonic control at the evolving continental margin of Namibia]: Zeitschrift deutsche geologische Gesellschaft, v. 149, p. 519-632.

Stollhofen, H., Gerschütz, S., Stanistreet, I.G. and Lorenz, V., 1998, Tectonic and volcanic control on early Jurassic rift-valley lake deposition during emplacement of Karoo flood basalts, southern Namibia: Palaeogeography, Palaeoclimatology, Plalaeoecology, v. 140, p. 185-215.

Stollhofen, H., Stanistreet, I.G., Rohn, R., Holzförster, F. and Wanke, A., 2000, The Gai-As lake system, northern Namibia and Brazil, *in* Gierlowski-Kordesch, E.H & Kelts, K.R., eds., Lake Basins through space and time: Studies in Geology, American Association Petroleum Geologists, V. 46, p 87-108.

South African Committee for Stratigraphy (S.A.C.S.), 1980, Stratigraphy of South Africa. Part 1: Lithostratigraphy of the Republic of South Africa, South West Africa/Namibia and the Republics of Bothuthatswana, Transkei and Venda: Handbook Geological Survey South Africa, v. 8, p. 1-690.

Swart, R., 1992/93, Note: Cretaceous synvolcanic conglomerates on the coastal margin in Namibia related to the break-up of West Gondwana: Communications Geological Survey Namibia, v. 8, p. 137-141.

Tankard, A.J., Jackson, M.P.A., Eriksson, K.A., Hobday, D.K., Hunter, D.R. and Minter, D.E.L., 1982, Crustal evolution of Southern Africa: New York, Springer Verlag, p. 1-523.

Tankard, A.J., Uliana, M.A., Welsink, H.J., Ramos, V.A., Turic, M., França, A.B., Milani, E.J., De Brito Neves, B.B., Eyles, N., Skarmeta, J., Santa-Anna, H., Wiens, F., Cirbán, M., Lopez-Paulsen, O., Germs, G.J.B., De Wit, M.J., Machacha, T. and Miller, R.M., 1995, Structural and tectonic control of basin evolution in southwestern Gondwana during the Phanerozoic, *in* Tankard, A.J., Suárez-Soruco, R. and Welsink, H.J., eds., Petroleum basins of South America, American Association Petroleum Geologists, Volume 62: Tulsa, p. 5-52.

Tera, F. and Wasserburg, G.J., 1972, U-Th-Pb systematics in three Apollo 14 basalts and the problem of initial Pb in lunar rocks: Earth Planetary Science Letters, v. 14, p. 281-304.

Trouw, R.A. and De Wit, M., 1999, Relation between Gondwanide Orogen and contemporaneous intracratonic deformation: Journal African Earth Sciences, v. 28, p. 203-213.

Tucker, M.E. and Wright, V.P., 1990, Carbonate sedimentology: London, Blackwell, p. 1-482.

Turner, B.R., 1983, Braidplain deposition of the Upper Triassic Molteno Formation in the main Karoo (Gondwana) Basin, South Africa: Sedimentology, v. 30, p. 77-89.

Turner, B.R., 1999, Tectonostratigraphical development of the Upper Karoo foreland basin: orogenic unloading versus thermally-induced Gondwana rifting: Journal African Earth Sciences, v. 28, p. 215-238.

Vail, P.R., Audemard, F., Bowman, S.A., Eisner, P.N. and Perez-Cruz, C., 1991, The stratigraphic signatures of tectonics, eustasy and sedimentology - an overview, *in* Einsele, G., Ricken, W. and Seilacher, A., eds., Cycles and events in stratigraphy: Berlin, Springer, p. 617-659.

Veevers, J.J., Powell, M., Collinson, J.W. and López-Gamundi, 1994b, Synthesis, *in* Veevers, J.J. and Powell, M., eds., Permian-Triassic Pangean basins and foldbelts along the Panthalassan margin of Gondwanaland: Memoir Geological Society America, Volume 184, p. 331-353.

Vidotti, R.M., Ebinger, C.J. and Fairhead, J.D., 1995, Lithospheric structure beneath the Paraná Province from gravity studies, is there a buried rift system ?: EOS, Transactions American Geophysical Union, v. 76, p. 608.

Visser, J.N.J., 1986, Lateral lithofacies relationships in the glacigenic Dwyka Formation in the western and central parts of the Karoo Basin: Transactions Geological Society South Africa, v. 89, p. 373-383.

Visser, J.N.J., 1990, The age of late Palaeozoic glacigene deposits in southern Africa: South African Journal Geology, v. 93, p. 115-131.

Visser, J.N.J., 1993, See-level changes in a back-arc-foreland transition: the late Carboniferous-Permian Karoo Basin in South Africa: Sedimentary Geology, v. 83, p. 115-132.

Visser, J.N.J., 1996, Controls on Early Permian shelf deglaciation in the Karoo Basin of South Africa: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 152, p. 129-139.

Visser, J.N.J., 1997, Deglaciation sequences in the Permo-Carboniferous Karoo and Kalahari basins of southern Africa: a tool in the analysis of cyclic glaciomarine basin fills: Sedimentology, v. 44, p. 507-521.

Wanke, A. and Stollhofen, H., 2000, The Long-Lived Rift Evolution prior Break-Up of Gondwanaland: Evidence from the Karoo-Etendeka Depositories in NW-Namibia: Journal African Earth Sciences, v. 30, p. 86-87.

Wanke, A., Stollhofen, H., Lorenz, V. and Stanistreet, I.G., 1998, The pre-breakup evolution of a continental margin: Synsedimentary Karoo tectonics of the Huab basin in NW-Namibia: Journal African Earth Sciences, v. 27, p. 206.

Ward, J.D. and Martin, H., 1987, A terrestrial conglomerate of Cretaceous age - a new record from the Skeleton Coast, Namib Desert: Communications Geological Survey S.W. Africa/Namibia, v. 3, p. 57-58.

Warren, A.A. and Davey, L., 1992, Folded teeth in temnospondyls - a preliminary study: Alcheringa, v. 16, p. 107-132.

Warren, A.A., Rubidge, B.S., Stanistreet, I.G., Stollhofen, H., Wanke, A., Latimer, E.M. and Marsicano, C.A., in press, Oldest Known Stereospondylous Amphibian, Late Permian, Namibia: Vertebrate Palaeontology.

Wasson, R.J. and Hyde, R., 1983, Factors determing desert dune type: Nature, v. 304, p. 337-339.

Watkins, R.T., Mc Dougall, I. and Le Roux, A.P., 1994, K-Ar ages of the Brandberg and Okenyenya igneous complexes, north-western Namibia: Geologische Rundschau, v. 83, p. 348-356.

White, S., Holzförster, F. and Stollhofen, H., 2000, Syn- and post Karoo activity on the Waterberg Fault, Namibia, Geological Society London, Tectonic Studies Group, Research in Progress Meeting Manchester 10th-12th January, 2000, p. 37.

Wiechmann, P., 1938, Praehistorische Spuren am Omuramba Omambonde: Roan News, v. 7, p. 1-2.

Williams, I.S. and Claesson, S., 1987, Isotopic evidence for the Precambrian provenance and Caledonian metamorphism of high grade paragneisses from the Seve Nappes, Scandinavian Caledonides. II. Ion microprobe zircon U-Th-Pb: Contributions Mineralogy Petrology, v. 97, p. 205-217.

Williams, K.E., 1995, Tectonic subsidence analysis and Paleozoic paleogeography of Gondwana, *in* Tankard, A.J., Suárez-Soruco, R. and Welsink, H.J., eds., Petroleum basins of South America: Memoir American Association Petroleum Geologists, Volume 62, p. 79-100.

Wittig, R., 1976, Die Gamsberg-Spalten (SW Afrika) - Zeugen Karoo-zeitlicher Erdbeben: Geologische Rundschau, v. 65, p. 1019-1034.

Woodcock, N.J. and Fischer, M., 1986, Strike-slip duplexes: Journal Structural Geology, v. 8, p. 725-735.

Yates, A.M. and Warren, A.A., in press, The phylogeny of the "higher temnospondyls" (Vertebrata: Choanata) and its implications for the monophyly and origins of the Stereospondyli: Zoological Journal Linnean Society.

Zalán, P.V., Wolff, S., Astolfi, M.A.M., Wieira, I.S., Conceição, J.C.J., Appi, V.T., Neto, E.V.S., Cerqueira, J.R. and Marques, A., 1990, The Paraná basin, Brazil, *in* Leighton, M.W., Kolata, D.R., Oltz, D.F. and Eidel, J.J., eds., Interior cratonic basins: Memoir American Association Petroleum Geologists, Volume 51, p. 681-708.

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