SEQUENCE STRATIGRAPHIC AND CHEMOSTRATIGRAPHIC CONSTRAINTS ON THE EVOLUTION OF THE TERMINAL PROTEROZOIC TO CAMBRIAN NAMA BASIN, NAMIBIA

by

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ABSTRACT

Mixed siliciclastic and carbonate strata of the terminal Proterozoic Witvlei Group and the overlying terminal Proterozoic to Cambrian Nama Group of Namibia form an important record of tectonic, climatic, environmental, and biologic changes during the dawn of animal life on Earth, while radiometrically-dated volcanic ash beds, intercalated within the Nama Group, provide some of the only available constraints on the absolute timing of these events. This study combines sequence stratigraphic and cheemostratigraphic data to correlate stratigraphic units between widely separated regions on the Kalahari Craton. Marked C-isotope excursions constrain correlations of depositional sequences that cannot be directly traced across the craton, and thus make it possible to piece together a temporally calibrated chronostratigraphy. The Witvlei Group and its correlatives in the Naukluft Nappe Complex comprise two depositional sequences, each of which is bounded below by glacial tillite and records a strong negative carbon-isotope excursion. The tillite horizons may represent two distinct Varanger-age glaciations. Depositional sequences in the lower Nama Group (Kuibis Subgroup) record a marked positive carbon-isotope excursion and are correlated across more than 800 km of the Kalahari Craton from the northern border, where they conformably overlie the Witvlei Group, to near the southernmost exposures of the Nama Group, where they are overlain by the carbonate rich Schwarzzrand Subgroup. Moderately positive $\delta^{13}C$ values recorded by Schwarzzrand carbonates extend to within 1 m.y. of the Proterozoic-Cambrian boundary. By correlating cheemostratigraphic intervals around the world, comparing sediment accumulation rates, and extrapolating downward, a maximum age of approximately 565 Ma is estimated for the end of the younger Varanger glaciation. This estimate reduces the temporal separation between the Varanger ice ages and the Cambrian Period by 25 m.y. and strengthens stratigraphic and geochemical evidence for post-Varanger, Proterozoic-age glaciations, which now may be regarded as the last pulses of an extended epoch of repeated ice expansion and retreat.

Thesis Supervisor: John Grotzinger
Title: Professor of Geology
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DEDICATION

I dedicate this thesis to my father, who influenced me more than I could ever know, and more than I’d sometimes like to admit.
CHAPTER 1

INTRODUCTION

The end of the Proterozoic Eon was a time of tremendous global tectonic, environmental, and climatic change (Knoll and Walter, 1992). Some of these changes include the assembly of the supercontinent Gondwanaland (e.g. Hoffmann, 1991), shifts in climate leading to the expanse and retreat of ice sheets and the possible development of glaciers near the equator (see Eyles, 1993 chapter 13 and 14 for a review), plus large fluctuations in the chemical composition of the oceans and atmosphere (e.g. Derry, 1992, Des Marais, et al., 1992, Tucker, 1992, Kaufman and Knoll, 1995). In addition, it seems likely that these events may have set the stage for the advent of animal life during the terminal Proterozoic period and for its rapid (Bowring et al., 1993; Grotzinger et al., 1995) differentiation during the Cambrian Period (e.g. Tucker, 1992, Kaufman and Knoll, 1995). To understand how these different changes might be related, it is necessary to reconstruct what happened, when it happened, where, and how fast. Thus, research projects are underway on all continents attempting to correlate stratigraphic units locally and globally and reconstruct the rock record of terminal Proterozoic events. In the absence of a diverse fossil record and well-defined biostratigraphic zones, however, alternative correlation techniques have needed to be developed.

The most promising of these techniques is chemostratigraphy (see Kaufman and Knoll, 1995 for a review). Carbonate-bearing successions around the world record several large (as much as 10%) excursions of $^{13}$C to $^{12}$C ratio (expressed as $\delta^{13}$C). These excursions are larger than any seen in the Phanerozoic record and are much greater than the isotopic variability of the modern ocean. They maintain broadly consistent positions relative to other chronostratigraphic
markers such as glacial horizons, Ediacaran-type fossils, and stratigraphic changes in Sr isotope ratio recorded by limestone, and are interpreted to record global changes in the carbon isotope composition of the world oceans. Thus, in addition to being a record of environmental and chemical change in the Neoproterozoic oceans, they form a powerful tool for global correlation. Carbon isotope studies can not be done in a vacuum, however. Instead, they must be considered with within the context of their physical stratigraphic framework. Toward this end, the techniques of sequence stratigraphy (Van Wagoner et al., 1988; Sarg, 1988) are extremely powerful. Sequence stratigraphy provides a method for identifying the major unconformities or gaps in the stratigraphic record and also for defining relative time-lines for intrabasinal and possibly interbasinal correlation of Neoproterozoic successions (Cristie-Blick et al., 1988; Christie-Blick et al., 1995).

Mixed carbonate and siliciclastic rocks of the terminal Proterozoic to Cambrian Witvlei and Nama groups partially cover the Kalahari craton in southern Namibia. They were deposited, in part, in a foreland basin that developed on the Kalahari craton as a flexural response to convergence along the adjacent Damara and Gariep orogenic belts (Germs, 1983) during the suturing of east and west proto-Gondwanaland (e.g. Hoffman, 1991). They lie stratigraphically above glacial rocks (Hoffmann, 1989) and contain many of the principle features of terminal Proterozoic chronostratigraphy (Knoll and Walter, 1992), including assemblages of Ediacaran-type fossils, skeletalized fossils and trace fossils (e.g. Germs, 1983; Grotzinger et al., 1995), plus stratigraphic excursions in carbon isotope ratio ($\delta^{13}C$) (Kaufman et al., 1991). In addition, the Nama Group contains abundant ash beds which provide high-resolution radiometric calibration of terminal Proterozoic chronostratigraphy (Grotzinger et al., 1995). Thus, terminal Proterozoic rocks of the Witvlei and Nama groups form
an important, temporally-calibrated record of terminal Proterozoic biologic, tectonic, climatic and environmental events. The next three chapters, which were written as independent papers, present results of sequence stratigraphic and carbon-isotope chemostratigraphic studies aimed at reconstructing that record.

The Nama basin is divided into two sub-basins by a regional basement arch (Germs, 1983). Chapters 2 and 3 concern the physical stratigraphy of the lower Nama Group (Kuibis and Schwarzrand subgroups) in the sub-basin south of the arch. Chapter 2, which was co-authored with J. Grotzinger and G. Germs, was originally published in Precambrian Research (Saylor et al., 1995) and is reprinted here with the permission of Elsevier Science. It describes the main depositional facies and applies the concepts of sequence stratigraphy to divide the succession into unconformity-bounded depositional sequences. The addendum at the end of Chapter 2 supplements the Precambrian Research version; it provides additional description of facies in the carbonate platform near the top of the Schwarzrand Subgroup. Chapter 3 has been submitted to the Communications of the Geological Survey of Namibia as a paper coauthored with J. Grotzinger. It combines both stratigraphic analysis and structural mapping to reconstruct the geometry of the Proterozoic-Cambrian boundary unconformity in the southern Schwarzrand Subgroup. The unconformity surface, which lies at the top of a thick carbonate-platform succession and has relief of up to 500 m, is interpreted to modify a platform-to-basin transition telescoped by thrust faults into a relatively small area. Reconstruction of these structural and stratigraphic relationships is important because the carbonate platform extends to within 1 m.y. of the Precambrian-Cambrian boundary (Grotzinger et al., 1995), contains some of the youngest known Ediacaran-type fossils near its top (Grotzinger et al., 1995), and is thus an important Proterozoic-Cambrian boundary reference section.
Chapter 4 has been submitted to the *Journal of Sedimentary Research* as a paper co-authored with A. Kaufman, J. Grotzinger, and F. Urban. It presents carbon-isotope data from the Witvlei Group and the Kuibis and Schwarzrand subgroups of the Nama Group in four geographically separated regions of the Kalahari craton, including the northern and southern Nama sub-basins. Sequence stratigraphic and carbon-isotope chemostratigraphic frameworks establish constraints on correlation between the regions. Tie points along isotope profiles serve as time-lines for correlation of depositional sequences which can not be directly traced, while sequence boundaries provide independent tests for chemostratigraphic correlation within each sub-basin. Isotope profiles are tied to U-Pb zircon age constraints which significantly compress the duration of carbon-isotope chemostratigraphic intervals compared to previous estimates. A composite, temporally-calibrated isotope record is constructed and compared with the isotope record from terminal Proterozoic successions on other continents in order to consider the implications of available U-Pb age constraints for the partitioning of time in the terminal Proterozoic rock record.

Representative generalized and bed-by-bed measured sections, plus photographs of depositional facies are included in the Appendices at the end of the volume.

**REFERENCES**


Saylor, B.Z., 1995, Sequence stratigraphy and sedimentology of the Neoproterozoic Kuibis and Schwarzrand Subgroups (Nama Group), southwestern Namibia: Precambrian Research, v. 73, p. 153-171.


CHAPTER 2

Sequence stratigraphy and sedimentology of the Neoproterozoic Kuibis and Schwarzzrand Subgroups (Nama Group), southwestern Namibia

Abstract

The Kuibis and Schwarzzrand Subgroups of the Nama Group form a succession of shallow-marine and minor fluvial sedimentary rocks that is exposed over much of central and southern Namibia. Ediacaran-type body fossils and the stratigraphic carbon-isotope variability indicate a Vendian age for much of these strata; the Precambrian-Cambrian boundary is in the uppermost part of the Schwarzzrand Subgroup. Radiometric dating of abundant volcanic ash beds which span much of the Vendian and extend into the Cambrian part of the section could make the Nama Group the best calibrated reference section for Vendian chronostratigraphy. Here we describe the sedimentology and sequence stratigraphy of exposures of the Kuibis and Schwarzzrand Subgroups in southwestern Namibia. We place tighter environmental constraints on the paleontology and geochemistry and we identify seven depositional sequences. The boundaries of the depositional sequences correspond to the more important discontinuities in the Kuibis and Schwarzzrand Subgroups; their identification is crucial for defining rates of organism evolution and isotopic differentiation.

1. Introduction

The Vendian to Cambrian Nama Group is a 3000 m succession of shallow-marine and fluvial, siliciclastic and carbonate sedimentary rocks that covers extensive areas in central and southern Namibia (Fig. 1; Germs 1983). In large part, the Nama Group was deposited in a foreland basin that developed in response to collision along the adjacent Damara and Gariep orogenic belts (Fig. 1) during the Neoproterozoic assembly of Gondwanaland (Germs, 1974; Germs, 1983; Stanstreet et al., 1991). The Nama Group contains some globally recognized Vendian and Cambrian body-fossil and trace-fossil assemblages as well as endemic biota (Germs, 1972a, 1972b, 1995; Crimes and Germs, 1982). Preliminary determinations of the stratigraphic carbon-isotope variability in the lower Nama Group seem to correlate well with other Vendian sections around the world (Kaufman et al., 1991; Derry et al., 1992; Kaufman and Knoll, 1995). Furthermore, the Nama Group contains abundant ash beds, some of which were previously noted by Germs (1972c). During recent field work, we collected volcanic ash samples that span much of the Vendian part of the section and extend into lower Cambrian strata of the Nama Group (Germs, 1972c; Fig. 2). In conjunction with sedimentologic and stratigraphic controls relating key geochemical, paleontological, environmental and tec-
tonic features, radiometric dating of these volcanic rocks should make the Nama Group one of the best reference sections available for absolute temporal calibration of Vendian chronostratigraphy.

Here, we present preliminary results of an ongoing sedimentologic and stratigraphic investigation of the Nama Group. This paper focuses on exposures of the lower part of the Nama Group (Kuibus and Schwarzzand Subgroups) in southern Namibia (Fig. 1). It is from this region that the volcanic ash beds and carbon-}

ate isotope samples were collected, and in which most of the fossils have been discovered. The results are presented here as a tentative sequence-stratigraphic framework. They refine previous descriptions and interpretations of sedimentary facies (Germs, 1974, 1983) and place tighter environmental constraints on the paleontology and geochemistry of the Nama Group. The proposed sequence boundaries include erosional disconformities as well as more subtle, but possibly equally important, unconformities in the Nama Group (Fig. 2).

2. Stratigraphy, paleontology and geochemistry of the Nama Group

2.1. Sedimentology and stratigraphy

The Nama Group consists, from bottom to top, of the Kuibus, Schwarzzand and Fish River Subgroups. During deposition of the Kuibus and Schwarzzand Subgroups, a basement arch at Osis farm divided the Nama basin into two parts so that stratigraphic units thin or pinch out over it and change character across it (Germs, 1983). Germs' (1972c, 1983) descriptions of the sedimentology and stratigraphy of exposures of the Nama Group north and south of the Osis arch are briefly reviewed here.

South of the Osis arch, the Kuibus Subgroup consists of two successions, each of which is characterized by basal feldspathic conglomerate, followed upward by quartz sandstone, siltstone and limestone (Fig. 2). The successions thin and the limestone units pinch out eastward onto the craton. North of the Osis arch, the Kuibus Subgroup thickens northward toward the Damara orogen; the coarse sandstone and limestone grades laterally northward into limestone and shale (Germs, 1983). Germs (1983) interpreted the coarse siliciclastic rocks as mainly fluvial and the limestone and shale units as marine.

The Schwarzzand Subgroup is composed of siltstone, limestone and sandstone units (Fig. 2) that Germs (1983) interpreted as distal fluvial to offshore marine. North of the Osis arch, the siliciclastic units coarsen and thicken toward the Damara belt. Limestone is restricted to exposures south of the Osis arch, where it is the dominant lithology in the upper part of the Schwarzzand Subgroup. Similar to the Kuibus Sub-
group, the Schwarzrand Subgroup thins eastward as it encroaches the craton. Erosional valleys, which Germs (1983) interpreted as possible glaciogenic features, incise strata near the base and near the top of the Schwarzrand Subgroup.

The Fish River Subgroup unconformably overlies the Schwarzrand Subgroup. Germs (1983) identified five cycles with conglomerate or sandstone at the bases and mudstone-rich tops. He interpreted these cycles as fluvial clastic wedges with transgressive tidal-flat caps.

The fluvial sediments were derived, principally, from the eroding Damara orogenic belt (Germs, 1974, 1983).

2.2. Paleontology

The stratigraphic positions of various fossil identifications are shown in Fig. 2, and are reviewed by Germs (1995). Soft-bodied fossils of Ediacaran-affinity are present in the upper Kuibis and lower
Schwarzzand Subgroups and indicate a Vendian age for these strata: microfossil assemblages support this age interpretation (Germs et al., 1986). The small shelly fossil Cloudina (Germs, 1972a) and probable calcified metaphytes (Grant et al., 1991) are present in the limestone of the Kuibis and Schwarzzand Subgroups. These fossils represent some of the Earth’s oldest recognized calcifying organisms, and are consistent with a Vendian age (Grant, 1990; Grant et al., 1991). Trace fossils in the Kuibis and Schwarzzand Subgroups include some that are atypical of Vendian fossil assemblages (Germs, 1972b, 1995; Crimes and Germs, 1982).

An erosional unconformity incises thick Cloudina-bearing limestone near the top of the Schwarzzand Subgroup (Fig. 2). Above this horizon, sediments of the Nomitasas Formation and the overlying Fish River Subgroup contain the trace fossil Phycodes pedum, considered to be Early Cambrian in age (Crimes and Germs, 1982; Narbonne et al., 1987; Germs, 1995).

2.3. Chemostratigraphy

Kaufman et al. (1991) analyzed the isotopic compositions of organic and inorganic carbon from Neoproterozoic carbonates in Namibia, including the Nama Group, and found stratigraphic variations in the inorganic $^{13}$C abundances (Fig. 2). They used the stratigraphic isotope curves and glaciogenic horizons to suggest correlations among Neoproterozoic strata throughout Namibia. Furthermore, Kaufman et al. (1991) and Derry et al. (1992) suggested that the form of the isotope curves from Namibia is similar to isotope curves determined for other Neoproterozoic sections in Siberia and Norway. These similarities have led Kaufman et al. (1991), Derry et al. (1992), Knoll and Walter (1992) and others to infer global changes in ocean chemistry and to suggest that isotope stratigraphy might be a useful tool for correlating Neoproterozoic strata.

3. Sedimentology and sequence stratigraphy: Kuibis and Schwarzzand Subgroups

Recent field work has concentrated on identifying depositional sequences and sequence boundaries in the Kuibis and Schwarzzand Subgroups of the Nama Group. Depositional sequences are relatively conformable successions of strata separated by unconformities or their correlative conformities (Van Wagoner et al., 1988). The surfaces bounding sequences form during periods of relative sea-level fall and are recognized in the field by evidence of subaerial exposure, valley incision, or an abrupt superposition of disparate facies (Christie-Blick et al., 1988; Van Wagoner et al., 1988, 1990). Sequence boundaries are commonly quite subtle unconformities. Because these subtle unconformities can represent significant gaps in the sedimentary record, their recognition is a crucial part of any attempt to extract the depositional history of the Nama Group and its biological and geochemical features.

South of the Osis arch, we suggest dividing the Kuibis and Schwarzzand Subgroups into a minimum of seven depositional sequences, two in the Kuibis Subgroup and five, possibly six, in the Schwarzzand Subgroup. The positions of sequence boundaries, as well as formation and member boundaries, fossil assemblages, the carbon-isotope curves and volcanic ash beds, are shown in Fig. 2. Synopses of the sedimentary facies that make up these sequences are presented in Table 1.

3.1. Kuibis Subgroup

South of the Osis arch the Kuibis Subgroup comprises the Dabis and Zaris Formations and their component Kanies, Mara, Kliphoek and Mooifontein Members (Fig. 2). These members define two sequences with coarse siliciclastic bases (Kanies and Kliphoek Members) and carbonate tops (Mara and Mooifontein Members). Fig. 3 shows measured sections of the Kuibis Subgroup and a tentative sequence-stratigraphic framework.

3.1.1. K1 sequence

Coarse to pebbly sandstone of the basal K1 sequence generally overlies crystalline basement. The contact is planar, with no evident relief. In some localities near the Gariep orogenic belt and along the Orange River sequence K1 unconformably overlies older Proterozoic successions (Kröner and Germs, 1971; Hoffmann, 1989).

Kanies Member

Pebby to conglomeratic sandstone. At the base of sequence K1 there is commonly 1 to 2 m of pebbly to conglomeratic, medium- to thick-beded sandstone
The pebbly-conglomerate and sandstone is poorly sorted and immature. The contact with the crystalline basement can be difficult to distinguish due to a gradational transition from weathered granitoid gneiss to immature, basement-derived sandstone. The coarse grain-size and immaturity of these deposits are consistent with Germs' (1983) interpretation of a fluvial braidplain environment (Cotter, 1978).

**Cross-stratified sandstone.** Cross-stratified, coarse sandstone overlies the basal conglomeratic beds and is interbedded with planar-stratified sandstone (Fig. 3). The beds are 15 to 75 cm thick, laterally continuous and tabular in geometry. They are composed of amalgamated 5 to 15 cm sets of trough and tabular—planar cross-stratification separated by reactivation surfaces. Measured trough azimuths are unimodal and west-directed. This facies tends to be somewhat finer, better sorted and less feldspathic than the underlying conglomeratic sandstone.

The unidirectional paleocurrent measurements indicate strong current influence consistent with a fluvial interpretation. Also, the cross-stratified sandstone is interbedded with and grades westward (basinward) into planar-stratified, very shallow-marine, sheet sandstone (described next), which supports a fluvial to marginal-marine interpretation.

**Planar-stratified, medium sandstone.** Planar-stratified, medium sandstone is interbedded with cross-stratified sandstone in eastern exposures and completely replaces the cross-stratified facies in westernmost exposures (Fig. 3). Beds of planar-stratified sandstone are 15 to 50 cm thick. Small-scale wave ripples and polygonal desiccation cracks are present along bed tops, but are not abundant. An increase in fine material, including shale partings and siltstone intervals coincides with the westward transition from abundant cross to predominantly planar stratification.

The wave ripples and desiccation cracks indicate very shallow-water conditions with periods of emergence. The planar-stratified beds record distinct flow events. These features could record sheet-like, storm-initiated flows on a shallow, marginal-marine coastalplane (Johnson, 1975; Walker, 1984, 1985). In lulls between flow events, gentle wave action rippled the bed tops. Desiccation cracks developed either after retreat of the flood waters and return to normal subaerial conditions, or during rare periods of exposure.

**Mara Member**

Mixed fine siliciclastic and carbonate deposits. A few meters of interbedded siltstone and limestone sharply overlie the coarse sandstone of the Kanies Member (Fig. 3). This interval of mixed siliciclastic and carbonate deposits grades upward into thin- to medium-bedded calcisiltite and calcarenite of the Mara Member.

This transitional interval records a marine transgression that pushed the siliciclastic shoreline landward, trapping the coarse siliciclastic sediments on the craton. Subsequently, carbonate production commenced in the clear distal waters (Holmes and Evans, 1963; Mount, 1984).

**Calcarenite.** The Mara Member is composed of 5 to 20 cm thick beds of calcarenite with subordinate dolostite, calcisiltite and dololutite. Beds are tabular to irregular in geometry, commonly with sharp, scoured bases and irregular to wave-ripped internal lamination. In places, carbonate mud drapes the wave ripples. Siliciclastic shale forms partings between carbonate beds and rare shale beds range up to 20 cm thick. Flat-pebble, intraclast conglomerate is common and there are rare cracks. In western exposures there are a few thicker beds of trough cross-stratified ooid grainstone. In eastern exposures, calcarenite is interbedded with medium to coarse, planar-stratified sandstone.

The cross-stratified ooid grainstone and intraclast conglomerate indicate relatively strong flows. Variation in strength of the flows produced the interbedding of finer and coarser material (Demicco, 1983; Shinn, 1983). Therefore, the calcarenite is interpreted as a relatively low-energy, wave-influenced shallow-marine deposit. Cross-stratified ooid grainstones are more abundant westward, in the direction of a higher-energy open-marine setting. Siltstone beds are thicker and more abundant eastward, the direction from which they were sourced. Desiccation cracks record periods of supratidal exposure, but for the most part shallow-subtidal and intertidal conditions dominated (James, 1984; Demicco, 1985).

**3.1.2. K2 sequence**

Coarse and pebbly sandstone of the Kliphoek Member overlies Mara Member carbonate (Figs. 2 and 3). In eastern exposures, the carbonate-to-siliciclastic transition is abrupt. In western exposures, however, trough cross-stratified, coarse sandstone is interbedded on the
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<td>Cross-stratification, low-angle cross-stratified tabular sandstone</td>
<td>Multistorey cm-scale high and dark thin beds</td>
<td>Sharply defined bases</td>
<td>Type structures Neptunian dikes</td>
</tr>
<tr>
<td>Shale and siltstone</td>
<td>Calcite and calcareous mud</td>
<td>Mostly carbonate sand and silt with clasts</td>
<td>Basal contacts sharp, upper contacts gradational (Stage 3) and sharp (Stage 2)</td>
<td>Transgressive systems tract of Stage 2, Stage 3 and Stage 5</td>
</tr>
</tbody>
</table>

Table 1: Synthesis of the sedimentary facies

- **Facies**: Various facies types described with their characteristic features.
- **Lithology**: Textural characteristics of the deposits.
- **Sedimentary features**: Specific sedimentary structures observed.
- **Associations**: Other deposits associated with the described facies.
- **Interpretation**: Interpretative summary of the facies and its environment of deposition.
<table>
<thead>
<tr>
<th>Facies</th>
<th>Characteristics</th>
<th>Depositional Environment</th>
</tr>
</thead>
<tbody>
<tr>
<td>Massive-planar sandstone</td>
<td>Fine- to medium-grained, well-sorted sandstone</td>
<td>Laterally extensive incised valley fill at base of S2</td>
</tr>
<tr>
<td></td>
<td>~20 m thick massive unit</td>
<td>High velocity density flows</td>
</tr>
<tr>
<td></td>
<td>Meter-scale beds</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Massive to vaguely-planar stratification, pillow structures, or basal contact</td>
<td></td>
</tr>
<tr>
<td></td>
<td>erosion with several meters of relief</td>
<td></td>
</tr>
<tr>
<td>Large-scale cross-stratified sandstone</td>
<td>Well-sorted, fine-grained sandstone</td>
<td>Laterally extensive low-stand deposits of K2</td>
</tr>
<tr>
<td></td>
<td>Red bed thicknesses range from 50 to 200 cm</td>
<td>Lower and upper current-dominated shoreface</td>
</tr>
<tr>
<td></td>
<td>Amalgamated beds</td>
<td>Delta deposits</td>
</tr>
<tr>
<td></td>
<td>Tabular-planar, tangential and trough cross-stratification (50–200 cm high</td>
<td></td>
</tr>
<tr>
<td></td>
<td>cross-sets)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Scoured bases, channels</td>
<td></td>
</tr>
<tr>
<td>Shale with tabular-beded sandstone</td>
<td>Red shale</td>
<td>Laterally extensive lowstand and transgressive deposits of S1</td>
</tr>
<tr>
<td></td>
<td>Occasional to abundant interbeds of fine-grained sandstone</td>
<td>Very shallow marine storms, storm surge and flood-driven event beds</td>
</tr>
<tr>
<td></td>
<td>Sandstone beds are 5–30 cm thick</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Planar lamination, low-angle and hummocky cross-stratification</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Wave, interference and current ripples</td>
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<tr>
<td></td>
<td>Flute marks, gutter casts, ball and pillow structures</td>
<td></td>
</tr>
<tr>
<td></td>
<td>5–15 cm sandstone upward cycles</td>
<td></td>
</tr>
<tr>
<td>Small-scale cross-stratified sandstone</td>
<td>Poorly sorted, course-grained to pebbly sandstone</td>
<td>Laterally extensive interbedded with and graded easterward into planar-stratified sandstone</td>
</tr>
<tr>
<td></td>
<td>20–50 cm thick beds, tabular, sheet-like geometry</td>
<td>Low-stand deposits of K1</td>
</tr>
<tr>
<td></td>
<td>3–50 cm scale, commonly nested through cross-stratification</td>
<td>Fig. 3</td>
</tr>
<tr>
<td></td>
<td>Abundant mud chips</td>
<td></td>
</tr>
<tr>
<td>Planar stratified sandstone</td>
<td>Poorly sorted, fine- to coarse or pebbly sandstone Siltsone partings</td>
<td>Distal low-stand deposits of K1</td>
</tr>
<tr>
<td></td>
<td>20–50 cm beds</td>
<td>Wave-influenced, shallow marine flow event beds</td>
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<tr>
<td></td>
<td>Thick (1–3 cm), plane-parallel internal stratification, some low-angle cross-</td>
<td></td>
</tr>
<tr>
<td></td>
<td>stratification, some low-angle cross-stratification</td>
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<tr>
<td></td>
<td>Wave ripples and desiccation cracks</td>
<td></td>
</tr>
<tr>
<td>Mixed siliciclastic and car-</td>
<td>Fine-grained sandstone</td>
<td>Transitional interval between sandstone and limestone deposits of K1, K2, and S3.</td>
</tr>
<tr>
<td>bonate deposits</td>
<td>Siliciclastic and carbonate siltstone</td>
<td>Figs. 3 and 4</td>
</tr>
<tr>
<td></td>
<td>5–20 cm interbeds</td>
<td>Shallow subtidal to intertidal</td>
</tr>
<tr>
<td></td>
<td>Mostly planar stratification</td>
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</tr>
<tr>
<td></td>
<td>Wave-ripple laminations</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Possible desiccation cracks</td>
<td></td>
</tr>
<tr>
<td>Anulagamated wave-rippled</td>
<td>Green and gray siltstone to very fine sandstone</td>
<td>Lateral extensive in S1 and S2</td>
</tr>
<tr>
<td>sandstone and siltstone</td>
<td>Amalgamated, climbing oscillation and combined-flow ripples. Straight-crested</td>
<td>Deltic interchannel areas</td>
</tr>
<tr>
<td></td>
<td>and 3-dimensional 2–30 cm spacing</td>
<td></td>
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</tbody>
</table>
Fig. 3. Representative measured stratigraphic sections through the K1 and K2 sequences in the Kuibis Subgroup. Note that the Mooifontein Member was eroded along the S1 sequence boundary.
scale of a few meters with thin-bedded calcisiltite and trough cross-stratified ooid grainstone. Near the Fish River Canyon, there is an erosional unconformity at the base of the Kliphoek Member.

Probably, relative sea-level fall produced erosion in cratonic areas and caused the siliciclastic shoreline to prograde rapidly over the Mara Member carbonate, injecting coarse sand into the basinal areas (e.g., Holmes and Evans, 1963; Mount, 1984). Shoreline progradation was not sufficient to completely shut off carbonate production in the more distal parts of the basin such as Swartkloofberg where the Kliphoek Member consists of meter-scale interbeds of coarse sandstone and calcarenite. We place the K2 sequence boundary at the base of the lowest bed of coarse sandstone, consistent with Germs and Gresse (1991).

Kliphoek Member

Like much of the Kanies Member, the basal part of the Kliphoek Member consists of cross-stratified, coarse to pebbly sandstone. Beds are thicker and more irregular in geometry than in the Kanies Member. Cross-sets generally range between 1 and 2 m in height; there are abundant channelized and scoured surfaces.

Kliphoek Member sandstone records strong current influence and may be fluvial in origin. However, the large size of the cross-sets requires water depths of at least several meters and more. There is no evidence for exposure. These conditions may be more consistent with deposition in a marine environment strongly influenced by tidal or deltaic currents.

In western exposures, the Kliphoek Member consists of interbedded large-scale cross-stratified coarse sandstone and carbonate grainstone (Fig. 3). The westward transition from siliciclastic to carbonate sediments is consistent with deposition in a westward-deepening basin. In the basal exposures, backstepping, limestone-based cycles overlie a transgressive surface. In more proximal locations, the transgressive surface sharply separates uniformly coarse sandstone from recessive, fine to coarse sandstone and siltstone or limestone (Fig. 3).

Mooifontein Member

The Mooifontein Member is similar to the Mara Member (Fig. 3). However, it has overall thinner bedding and, except for a thin mixed siliciclastic and carbonate interval at its base, it lacks siliciclastic beds. Mixtures of carbonate silt- and sand-sized material, with subordinate ooids and Cloudina fragments make up fine- to coarse-interbeds that generally range from 3 to 7 cm thick.

Because of thin bedding and lack of evidence for exposure or strong flows, we interpret the Mooifontein Member as predominantly subtidal (Demicco, 1983). Transgression across sandstone of the Kliphoek allowed carbonate production to recommence.

3.2. Schwarzrand Subgroup

The Schwarzrand Subgroup consists of the Nudaus, Urusis, and Nomtsas Formations and sequences S1 through S5 (Fig. 2). Sequences S1 and S2 consist predominantly of siliciclastic sediments; measured sections of these sequences are shown in Fig. 4. Sequences S3 and S4 are part of a thick carbonate platform succession (Fig. 5). A spectacular erosional unconformity separates siltstone and conglomerate of sequence S5 from carbonate of sequence S4.

3.2.1. S1 sequence

In eastern, cratonic exposures of the Nama, Urusis, and Nomtsas Formations in the Neiderhagen Member of the Nudaus Formation overlies the Mooifontein Member limestone. The discoveries of erosional incisions into the Neiderhagen Member sandstone (Germs, 1983) and the Mooifontein Member limestone designate a sequence boundary.

At Klein Karas farm, sandstone of the Neiderhagen Member is incised (Germs, 1983). At Aussenkjer farm, limestone of the Mooifontein Member is absent; in its place are interbedded carbonate–clast conglomerate and thick-bedded sandstone (Fig. 3). Germs (1983) interpreted the erosional relief at Klein Karas farm as possibly glacial in origin. We, however, suggest that the valleys at Aussenkjer and at Klein Karas farms were incised during relative sea-level fall.

Sandstone of the Neiderhagen Member is interpreted as a highstand deposit. In western exposures of sequence S1 highstand siliciclastic deposits are absent, probably due to incomplete shoreline progradation. The S1 sequence boundary in the sections is at the sharp contact between subtidal limestone of the Mooifontein Member and near-shore shale of Vingerbroek Member of the Nudaus Formation (Figs. 3 and 4).
Vingerbreek Member

Shale with tabular-bedded sandstone. Much of sequence S1 consists of red and green shale with interspersed sharp-based, tabular sandstone beds (Fig. 4). The sandstone beds are arranged in 5 to 15 m sandier-upward cycles with sharp, shale bases.

The tabular-bedded sandstone is characterized by parallel lamination, but also displays low-angle and rare hummocky and trough cross-stratification. Generally, the bed bases and tops are flat. However, in some cases, small- to medium-scale, straight-crested and three-dimensional oscillation ripples are developed on the bed tops. Gutter casts, flute marks and ball-and-pillow structures are present.

Load structures and flute marks in the tabular-bedded, planar-stratified sandstone indicate sporadic, high-velocity flow-events. The wave ripples indicate shallow water-depths, less than fair-weather wave base. Only rare examples of hummocky cross-stratification indicate deeper, more offshore deposition. For these rea-
sons, the shale-based, sandier-upward cycles are interpreted as dominantly near-shore shoaling-upward cycles. The flow events may be driven by storm currents augmented by delta floods (Swift et al., 1987; McCormick and Grotzinger, 1993).

Wave-rippled sandstone and siltstone. Wave-rippled sandstone and siltstone is found interbedded with tabular-bedded sandstone and shale. More commonly, however, it is present as one to several meter thick intervals which lack distinct beds and which consist, instead, of amalgamated ripples. These sandier intervals are characterized by 2 to 10 cm scale, well-developed, generally straight-crested climbing, oscillatory and combined-flow ripples.

The small spacing of the wave ripples in this facies is similar to that of wave ripples found on some of the tabular sandstone beds and indicates similarly shallow water-depths. To produce the well-developed, thick accumulations of wave ripples requires significant sediment influx, but excludes strong currents. A possible scenario is that the wave-rippled facies developed between delta channels. The delta currents carried out sediment in suspension; some sediment settled in interchannel areas where currents were weak and wave processes dominated.

Amalgamated tabular-bedded sandstone. At the tops of some of the sandier-upward cycles, the tabular sandstone beds are amalgamated. In particular, near the top of sequence S1 there is a thick shale-based succession which has thicker and more closely spaced tabular sandstone beds upward, and has several meters of amalgamated sandstone at its top (Fig. 4). The stratification in this unit varies upward from dominantly planar to hummocky cross-stratification and trough cross-stratification.

Amalgamated tabular sandstone beds are prominent near the top of the sequence. The presence, in this part of the sequence, of hummocky cross-stratification and the lack of small-scale wave ripples indicate an increase in accommodation space. The cycles may be interpreted as aggrading followed by backstepping parasequences in lowstand to transgressive systems tracts. The thicker, amalgamated sandstone beds in the transgressive part of the sequence could have resulted from bypassing of the near-shore area (Myrow, 1992).

Cross-stratified calcarenite. In western exposures, near the top of sequence S1, there is a unit of calcarenite which caps amalgamated tabular-bedded sandstone (Fig. 4). The calcarenite is thin-bedded, and variably planar-stratified, cross-stratified or rippled. Generally, the calcarenite unit is sharply overlain by shale.

The stratification style of the calcarenite is similar to that of the underlying sandstone and probably indicates similar conditions of flow. A flooding surface separates the calcarenite beds from overlying shale.

3.2.2. S2 sequence

The base of the sequence S2 is a 10-20 m, massive to vaguely planar-stratified, thick-bedded sandstone unit that has a sharp, locally erosional base (Fig. 4). Erosion along the contact scours through underlying shale, limestone and tabular-bedded sandstone. The result is that in the westernmost exposures, such as Witputs farm, the basal S2 sandstone unit lies several meters above the calcarenite unit of sequence S1 (Fig. 4). The surface cuts down eastward so that at Kliphoek and Geel'perdshoek farms, the limestone is eroded. There the basal sandstone unit overlies amalgamated tabular-bedded sandstone and carbonate-clast conglomerate (Fig. 4).

In eastern exposures, sequence S2 consists mainly of sandstone. In western exposures it consists of sandstone, shale and limestone. In general, sequence S1 corresponds to Germs' (1983) Nasep Member of the Urusis Formation.

Nasep Member

Massive-planar sandstone. The sandstone that forms the base of sequence S2 is a distinct lithologic unit which is recognized over most of southern Namibia (Fig. 4). It consists of very thick, massive to vaguely planar-stratified sandstone beds. The sand is medium-grained and well-sorted.

A paucity of sedimentary structures makes interpretation of the massive-planar sandstone difficult. Such thick beds with vague planar stratification require strong, sediment-laden flow-events. These flow events may have occurred in broadly channelized distributaries that developed during relative sea-level fall.

Wave-rippled sandstone and siltstone. The massive-planar sandstone unit is sharply overlain by a 5 to 15 m thick interval of large-scale (20 to 40 cm spacing) wave-rippled (hummocky cross-stratified) gray to green siltstone (Fig. 4). Bedding surfaces show both straight-crested and three-dimensional ripple forms. The oscillation ripples are amalgamated so that, except
for the greater ripple-spacing, this facies is quite similar to the amalgamated, wave-rippled sandstone in sequence S1.

This thick interval of amalgamated wave-rippled silstone probably formed under similar hydrodynamic conditions as described for wave-rippled sandstone in sequence S1. The only difference is that the larger ripple-spacing indicates relatively deeper water. We interpret this facies as a transgressive deposit.

**Cross-stratified calcarenite.** In several western locations, limestone overlies the interval of large-scale wave-rippled sandstone (Fig. 4). The limestone unit in turn is overlain by green shale. The limestone unit has a sharp base, sharp top and thickens eastward. It consists of thin to medium beds of planar-stratified and trough and tabular cross-stratified calcarenite.

As in sequence S1, calcarenite is sharply juxtaposed above sandstone and below shale. Calcarenite units in both sequences S1 and S2 were deposited in current-influenced subtidal environments. In both cases, the switch from siliciclastic to carbonate deposition could have resulted from changes in sediment supply or delta lobe shifting. Alternatively, the production and dispersal of carbonate may have been in response to base-level changes.

**Green shale.** In western exposures, the middle part of sequence S2 consists of green shale. In contrast to shale of sequence S1, this shale contains only very rare sandstone beds. Because of the homogeneity of the green shale and the paucity of sedimentary structures and sandstone event beds within it, the shale is interpreted as the thickest offshore deposit described thus far. The shale interval includes the zone of maximum flooding for sequence S2 (Van Wagoner et al., 1988).

**Large-scale cross-stratified and planar-stratified sandstone.** The upper part of sequence S2 consists of medium sandstone. In western exposures, the sandstone sharply overlies the thick interval of green shale described above. In eastern exposures, however, the green shale is absent or greatly thinned, so that most of sequence S2 consists of sandstone. This sandstone facies is characterized by thick beds (0.5–2 m), with well-developed planar stratification, large-scale, west-directed tabular and trough cross-stratification, parting lineation, slump structures and soft sediment deformation.

This facies is interpreted as upper shoreface, probably deltaic deposits. The planar stratification and large-scale cross-stratification are evidence for steady, strong currents. The slump structures are consistent with failure on a delta slope (Martinsen, 1989).

The eastward change in thickness of shale and the sharp basal contact of the cross-stratified sandstone unit raise the suspicion that some erosion may have occurred beneath the unit. We have not found evidence for this erosion yet, and therefore see no need for a sequence boundary. Both features could have resulted purely from delta progradation (Van Wagoner et al., 1988, 1990).

### 3.2.3. S3 sequence

The S3 sequence boundary has erosional relief which increases eastward, reaching several tens of meters and incising through most of sequence S2. In eastern exposures, all that remains of sequence S2 are the massive-planar sandstone unit and locally some wave-rippled sandstone. At the S3 sequence boundary there is a 10 to 40 cm bed of very coarse to pebbly large-scale rippled sandstone. This bed is sharply overlain by limestone of sequence S3. In western exposures there is no evidence for erosion. The base of sequence S3 is placed at a sharp contact between the planar and cross-stratified sandstone of the uppermost S2 sequence and overlying mixed silts tone, limestone and sandstone.

The bulk of sequence S3 comprises carbonate platform deposits of the Huns Member. There is a thin transitional interval of mixed siliciclastic-and-carbonate rocks at the base of sequence S3 which records transgression and incipient carbonate platform development. There are no lowstand deposits; instead the sequence boundary and transgressive surface are amalgamated.

The carbonate platform is characterized by a cyclic interbedding of a variety of facies (Fig. 3), including

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*Fig. 5. (A) Measured section through the S3, S4 and S5 sequences at Swartkloofberg. Note the erosional unconformity that occurs through the Spitskopf limestone of the S4 sequence and is infilled with conglomerate of the Nortosa Formation (S5 sequence). This unconformity most likely represents the Vendian-Cambrian boundary. (B) Representative stratigraphic section through the S3 sequence at Aurb, a more cratonic setting than Swartkloofberg.*
green siltstone, thin-bedded calcisiltite,stromatolitic dolostones and a mixture of fine to coarse, cross-stratified limestone and breccia. Facies change markedly across depositional strike.

Huns Member

Green shale. Sharp-based units of green shale with minor planar-laminated, fine sandstone beds range from 2 to 10 m in thickness. Locally, subjacent carbonates are dolomitized, brecciated and eroded; more commonly, the contact is planar, with chert along it. Siltstone intervals are thicker and more abundant in eastern (cratonic) than in western (basinal) exposures.

The green shale is similar in appearance to the green shale that dominates the underlying S3 sequence. It probably records a similar quiet-water depositional environment that was disrupted by only rare high-energy flow-events.

Cross-stratified limestone. A mixture of fine to coarse and intraclast conglomeratic assemblages are grouped together here as cross-stratified limestone. Much of this facies consists of amalgamated beds of trough and tabular cross-stratified, fine calcisiltite or calcarenite. Coarser carbonate grains consist of rounded intraclasts of carbonate mudstone or sandstone, ooids, and Cloudina fragments. There are also interspersed 1 to 20 cm thick beds of flat-pebble conglomerate. Units of cross-stratified limestone are common in eastern exposures, thin westward and are nearly absent from western exposures of the sequence S3. Locally, tepee structures and dolomitized, sediment-filled neptunian dikes are present.

The absence of current-generated sedimentary structures and the abundance of coarse material indicate high-energy, shallow-subtidal environments (James, 1984; Demicco, 1985). Episodic storm events probably disrupted underlying beds and produced the common intraclast conglomerates. The tepee structures and neptunian dikes might have formed during periods of subaerial exposure (Assereot and Kendall, 1977) or, alternatively, by rupture of early-cemented limestone in the subtidal zone (Shinn, 1969).

Stromatolitic dolostone. Branching, elongate stromatolites are the most abundant form. Characteristically, stromatolites in sequence S3 form sharp-based, relatively continuous beds that commonly overlie dolomitized, flat-pebble conglomerate. They are interbedded with thin-bedded calcisiltite and to a lesser extent cross-stratified grainstone. Both stromatolite beds and intervals of thin-bedded calcisiltite are thicker and more abundant westward. In the vicinity of Swartkloofberg, stromatolites are part of meter-scale cycles, that are bounded by karstic surfaces with up to 2 m of relief. The karstic surfaces are commonly draped with stromatolites, which pass upward into calcisiltite.

Columnar stromatolites are limited to western (more distal) exposures of sequence S3 and, in particular, to the upper part of the sequence. The columnar stromatolites commonly overlie calcisiltite and are draped by calcisiltite, green siltstone or branching stromatolites.

The presence of stromatolite biostromes in more distal parts of the Huns carbonate platform suggests growth as buildups in the outer part of a ramp. A ramp geometry is suggested by the absence of shelf-derived breccias in calcisiltites, described below.

Calcisiltite. Massive and thin-bedded calcisiltite and calcilutite form regionally continuous units that are tens of meters thick. Minor coarser grainstones, flat-pebble conglomerate and siltstone beds are interspersed.

Similar to thin-bedded calcisiltite in the Mooifontein Member of sequence K2, calcisiltite intervals in the S3 sequence record relatively low-energy, but occasionally storm-influenced subtidal deposition. Because of the association of calcisiltite with stromatolites and with low-energy, more basinal parts of the platform, a distal carbonate ramp depositional environment is inferred.

3.2.4. S4 sequence

At Swartkloofberg farm, the rhythmically interbedded stromatolite and calcisiltite beds that characterize western exposures of sequence S3 (Fig. 5A) are sharply overlain by a unit of green siltstone with minor tabular-bedded sandstone (Feldschuhhorn Member).

The siltstone envelopes pinnacle reefs and grades sharply upward into dark limestone of the Spitskop Member of the Urusis Formation (Fig. 5). Because of abrupt vertical facies changes in this interval, two possible sequence boundaries are proposed.

Sequence S4a

Pinnacle reefs enclosed in green siltstone reach 60 m in height (Fig. 5). The reefs have a framework constructed predominantly of thrombolitic mesoclots and, to a lesser extent, stromatolites and marine cements. Cloudina shells are locally abundant as detritus filling
voids in the framework, but also are attached to the margins of thrombolitic mesoclasts. Fibrous, isopachous and botryoidal marine cements are now present as calcite pseudomorphs after former high-Mg calcite and aragonite. Marine cements encrust mesoclasts and stromatolites, but are most abundant where they fill neptunian dikes that cut the reefs. Some dikes are up to 50 cm wide and several meters deep. The margins of the reefs are locally flanked by coarse talus blocks of reefal origin.

The pinnacle reefs are exposed only where exhumed from the enclosing siltstone. Consequently, it is difficult to ascertain whether siltstone deposition was contemporaneous with, or post-dated, growth of the reefs. The pinnacle reefs must have formed in relatively deep-water and record a transgression-induced period of reduced carbonate production that forced reefs to grow as pinnacles. The green siltstone deposits might have by-passed nearer-shore, carbonate-producing areas and accumulated on the distal ramp (Grammer and Ginsburg, 1992) or they could record siliciclastic progradation during sea-level highstand. The surface at the base of the pinnacle reefs is certainly a significant flooding surface; we suggest that it might coincide with a sequence boundary.

**Sequence S4b**

In the green siltstone that encloses the pinnacle reefs there is a 10 m unit of laterally continuous columnar and branching stromatolites that is capped by a 50 cm thick flat-pebble conglomerate (Fig. 5). The matrix of the flat-pebble conglomerate is silified. Above this unit, more green siltstone and an additional pinnacle reef are present. The base of the conglomerate-capped stromatolite unit might also be a combined sequence boundary and transgressive surface.

**Spitskopf Member**

At the top of the Feldschuhorn Member the green siltstone grades abruptly into thin-bedded calcisiltite of the Spitskopf Member (Fig. 5). The transition consists of approximately 1 m of thinly interbedded siltstone and limestone. Much of the overlying calcisiltite shows incipient brecciation that grades laterally into fully developed platy-clast breccia. Platy-clast breccia beds range from 10 cm to 10 m in thickness. In some sections they are interbedded with 10 m thick interbeds of green siltstone.

The thin-bedded calcisiltite, platy-clast breccia and incipient breccia are interpreted as slope facies and slope breccias. During relative sea-level highstand, carbonate production resumed and platform deposits prograded out across the green siltstones (Grammer and Ginsburg, 1992).

**3.2.5. S5 sequence**

**Nomtsas Formation**

The base of sequence S5 is an erosional disconformity that is exposed well at Swartkloofberg (Fig. 5). There the erosional relief is at least 30 m and incises the slope limestone of the Spitzkopf Member. Further cratonward, at Sonntagsbrunn, there are additional incised valleys that incise sequence S3. Sonntagsbrunn valleys are anastomosing, irregular channels that have erosional relief of similar magnitude to the valley at Swartkloofberg. At both Swartkloofberg and Sonntagsbrunn, incised valleys are partially filled by conglomerates of the Nomtsas Formation; final fills consist of marine siltstones and interbedded diamicite.

Germs (1983) interpreted the incised valley and the valley fill as possible glaciogenic features. During recent field work, however, no demonstrably glacial features were identified along the S5 sequence boundary or within sequence S5; we suggest alternative depositional scenarios for facies in the Nomtsas Formation.

**Conglomerate.** At Swartkloofberg, there is interbedded matrix-supported and clast-supported conglomerate in the lower S5 sequence. Conglomerate clasts are angular and range up to 50 cm in diameter. Clasts consist of thin-bedded calcisiltite, stromatolitic dolostone and minor sandstone and siltstone. Thus, unlike the platy clasts in the slope breccias of sequence S4, this conglomerate is heterolithic and poorly sorted. Beds of matrix-supported conglomerate are generally massive to poorly stratified, have flat bases, and are inversely graded, with large (50 cm) boulders at their tops. These features are consistent with deposition by debris flows (Nardin et al., 1979; Hiscock and James, 1984). Beds of clast-supported conglomerate have erosional bases and planar stratification, consistent with deposition from tractional processes in what might have been a fluvial environment.

**Diamicite.** There is an abrupt transition from interbedded clast-supported and matrix-supported conglomerate to diamicite and siltstone. The position of
this transition corresponds to the stratigraphic level at which local incised valleys are completely filled (Fig. 5). The overlying diamictite interval is 30 m thick. It consists mostly of green siltstone with scattered angular to subrounded, 1 to 10 cm in diameter, carbonate and sandstone clasts. At a few horizons, there are trains of out-sized sandstone, carbonate and volcanic ash boulders. Individual boulders are up to 5 m in diameter. There are also minor interbeds of calcisiltite and sharp-based beds of clast-supported, platy-clast, carbonate conglomerate.

Significantly, no pebbles, cobbles or boulders are limestones within laminated mudstone; all carbonate and sandstone clasts are parts of well-mixed diamictite that can be interpreted as the result of sediment gravity flows or slumping. Much of the green siltstone could have accumulated in an offshore environment and then been incorporated into the sediment gravity flow events.

We interpret the coarse clasts of the lower Nomtsas Formation to have been derived from the flanks of incised valleys that became flooded during progressive onlap and transgression of the S4–S5 sequence boundary. Although much diamictite is stratigraphically above the incised valleys of the Swartkloofberg area, this is still consistent with diachronous flooding of the same surface in more cratonward locations. Accumulations of conglomerates in estuarine settings could have failed episodically producing mass flows and down-slope deposition (Swartkloofberg) in a deeper shelf environment that otherwise would not receive such coarse detritus.

The green siltstone with scattered clasts grades upward into clast-free green siltstone followed by interbedded siltstone and planar-stratified to low-angle cross-stratified, and rippled fine sandstone. As in underlying sequences, this facies is interpreted as recording upward-shoaling through shoreface sandstone event beds.

4. Discussion

We have described depositional sequences in the Vendian to Early Cambrian rocks of the lower part of the Nama Group. Depositional sequences in the Kuibis Subgroup (K1 and K2) consist of lowstand sandstone and truncated transgressive to highstand limestone.

Lowstand deposits in the depositional sequences of the Schwarzrand Subgroup (S1–S5) are generally absent or thin. However, based on vertical facies changes and the existence of cratonward erosional surfaces, we feel confident recognizing the depositional sequences and suggest that the sequence boundaries coincide with the transgressive surfaces. We have identified what should be the major gaps in the sedimentary record, and although we cannot yet constrain the magnitude of these gaps, we can consider some of their implications for the significance of the Nama Group to developing a global Vendian chronostratigraphy.

4.1. Lower Vendian biostratigraphy and chemostratigraphy

The K2 sequence boundary lies stratigraphically above Cloudina found in the Mara Member (Germs, 1983) and, below the lowest identified Ediacaran-type fossils found in the Kliphoek Member (Germs, 1983) (Fig. 2). Carbonates above and below the K2 sequence boundary have substantially different δ13C values. Thus, across the K2 sequence boundary, there are changes in the chemostratigraphy and the biostratigraphy of the Kuibis Subgroup.

The absence of soft-bodied fossils in sequence K1 might reflect conditions that were inappropriate for the early Vendian biota to live or be preserved. Alternatively, sandstones in sequence K1 may pre-date Ediacaran-type fossils. In this case, the period of Ediacaran radiation would coincide with the unconformity at the K2 sequence boundary. Similarly, the presence of this hiatus must be considered when attempting to correlate the positive shift in δ13C values of the Kuibis Subgroup with similar shifts in other Vendian sections.

4.2. Vendian–Cambrian boundary: biostratigraphy and chemostratigraphy

Germs (1983) indicated that the pinnacle reefs at Swartkloofberg developed above the S5 sequence boundary and erosional surface. However, field relationships from this study demonstrate that some of the pinnacle reefs are older than slope breccias that are incised by the S5 sequence boundary. None of the relationships require that any of the pinnacle reefs post-date valley incision. Therefore, most, and probably all,
of the pinnacle reefs are older than the S5 sequence boundary.

The relationship is important because the pinnacle reefs contain Cloudina, which Grant et al. (1991) suggested may be restricted to Vendian time. No Cloudina was recognized in deposits that demonstrably overlie sequence boundary S5. These deposits do, however, contain Phycodes pedum (Germs, 1983 and recent work), which is generally considered to be indicative of a Cambrian age. Our conclusions support the suggestion of Grant et al. (1991) by eliminating the need to recognize Cloudina in association with Phycodes pedum in sequence S5. The S5 sequence boundary is the most likely horizon for the Precambrian–Cambrian boundary.

Carbonate isotopic measurements from below and above the S5 sequence boundary show equivalent δ13C values of approximately +2 ppm (Kaufman et al., 1991). A characteristic negative shift in carbonate isotopic values that has been found in other Precambrian–Cambrian boundary sections (Kaufman and Knoll, 1995) is not apparent. Probably the rocks of that age in the Nama Group are absent. Instead, the time interval represented by the negative isotopic excursion is incorporated into the hiatus at the S5 sequence boundary.

4.3. Vendian glacial deposits

Sequences S4 and S5 record relative sea-level changes on the scale of several tens to hundreds of meters. One explanation for such high-amplitude changes in relative sea level is glacio-eustasy. However, evidence for glaciation in the Nama Group, and evidence for a worldwide latest Vendian or earliest Cambrian glaciation is inconclusive.

Deposits correlated with the Sturtian and Varanger glacial episodes, including Neoproterozoic rocks underlying the Nama Group are considerably older than the Nama Group (Hambrey and Harland, 1985). Harland (1989), based principally on glacial deposits in China, suggested that there might have been a third glacial period at approximately the Precambrian–Cambrian boundary. However, all prospective examples of these deposits have either uncertain glacial origin or uncertain age.

Simple disk-shaped fossils in the Twitya Formation underlying tillite beds in the Mackenzie Mountains of Canada are the only documented case of Proterozoic body fossils in glacial deposits (Hofmann et al., 1990). These simple forms are considered to be the oldest representatives of metazoan diversification and are older than either the S4 or S5 sequences of the Nama Group. Kilometer-deep incised valleys in the Wonoka Formation of Australia might have a partially glacio-eustatic origin. However, they lie well below the lowest Ediacaran assemblages (Jenkins, 1984; Christie-Blick et al., 1990) and so also presumably are considerably older than the Nama Group. Indeed, the emerging Vendian chronostratigraphy (Knoll and Walter, 1992) strongly implies the absence of any glacial deposits stratigraphically above diverse Ediacaran faunas.

5. Summary

(1) The Kuibis and Schwarzrand Subgroups can be divided into seven depositional sequences (K1–K2, S1–S5) separated by unconformities or their correlatives conformities. The boundaries of these sequences are the most important disconformities in the lower Nama Group. Recognition of these disconformities is necessary for constraining the rates of organism evolution and isotopic differentiation recorded by the Nama Group.

(2) The K1 and K2 sequences are similar. They consist of coarse silicilastic bases and carbonate tops which record predominantly shallow-marine depositional environments.

(3) The S1 and S2 sequences consist mostly of near-shore to offshore shale, siltstone and sandstone event beds. Both sequences have relatively thick successions of deltaic sandstone at their tops which are incised by the overlying sequence boundaries.

(4) The S3 and S4 sequences are largely platform carbonate deposits composed of a wide variety of facies including siltstone, cross-stratified limestone, slope breccia, stromatolitic bioherms and pinnacle reefs.

(5) The S5 sequence boundary forms a spectacular, regional network of incised valleys that scours through slope breccia in sequence S4. At Swartkloofberg, the valleys are infilled by conglomerate and overlain by diamicite. Because we identified no demonstrable glacial features, we tentatively reject glaciogenic hypotheses for the origin of sediments and other features at this stratigraphic level. This interpretation is supported by the global inventory of confirmed Neoproterozoic
glacial deposits, none of which are stratigraphically above diverse Ediacaran faunas.

(6) The first appearance of Ediacaran-type fossils and the Precambrian–Cambrian boundary coincides with the K2 sequence boundary and the S5 sequence boundary.

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References


Thick carbonate platform facies of the Huns, Feldschuhhorn and Spitskop members in the middle of the Schwarzrand Subgroup were briefly described in the main body of this chapter. Here the stacking patterns and lateral variation of platform facies are described in more detail.

Platform Geometry

The Huns/Spitskop carbonate platform is bounded above and below by erosional unconformities (Fig. A) and consists of mixed carbonate and siliciclastic facies. Carbonate facies are best exposed between the Sonntagsbrunn and Swartkloofberg; they pinch-out northward and eastward, but expand southwestward to where the carbonate platform reaches a maximum thickness of nearly a kilometer.

Depositional Facies

Clastic carbonate facies of the Huns/Spitskop carbonate platform consist of parted- to nodular, thin-bedded, fine-grained limestone (similar to Cambrian "ribbon rock"; e.g. Demicco, 1983; Cowen and James, 1993), medium- to thick-laminated calcisiltite and calcarenite, and pellet/intraclast/ooid packstone, grainstone, and rudstone. Thin-bedded limestone facies occur together with grainstone facies in coarsening-upward cycles, but also occur in thick non-cyclic units in the upper part of the platform (Spitskop Member). Both wave- and current-generated sedimentary structures are common in the laminated calcisiltite and coarser-grained facies. Microbial facies occur in the western and central exposures of the platform. They consist of elongate, bifurcating, laterally-linked stromatolites, pedestal stromatolitic and thrombolitic bioherms, larger (1-2
m) dome-shaped stromatolitic bioherms and, locally, giant (60 m) pinnacle reefs (Grotzinger and Khetani, 1995). Elongate stromatolites commonly occur in association with grainstone but also in association with carbonate and siliciclastic mudstone in coarsening-upward cycles in the central and western exposures of the platform. Larger-sized, equant microbial bioherms are generally associated with carbonate mudstone or parted limestone. Pinnacle reefs are only recognized at one horizon, above a flooding surface at the top of the Huns Member.

Many of these facies are common members of terminal Proterozoic and Cambrian carbonate platform successions (Cowan and James, 1993; Knoll et al., 1995; Pelechaty et al., 1996). The coarser-grained facies and the elongate stromatolite facies all record the strong influence of waves and currents and are interpreted to have formed on the proximal part of the platform in wave- and current-swept subtidal environments. A greater abundance of coarse-grained, cross-bedded facies in the westernmost platform exposures suggests deposition in high-energy shoal areas. Thick accumulations of parted and thin-bedded limestone contain rare intraclast-conglomerate beds, but generally lack evidence for wave and current influence. These facies and the associated columnar microbialites are interpreted to have formed in deeper-water, quiet-subtidal environments (e.g. Cowan and James, 1993). Pinnacle reefs are interpreted to have formed in a deeper water environments following temporary drowning of the platform.

Evidence for exposure is generally limited to karst surfaces which cap subtidal grainstone in the platform margin area. No extensive peritidal facies are recognized. Slope facies comprising deep-water limestone, shale and platy-clast breccia are restricted to the westernmost exposures of the Spitskop Member where a series of thrust sheets telescope the platform-to-basin transition into a relatively small area (Chapter 3).
Siliciclastic facies in the Huns/Spitskop carbonate platform consist of shale and fine sandstone. Hummocky cross-stratification, wave-ripples and planar-lamination are common and high-angle cross stratification is locally present indicating that these facies were deposited in a variety of wave-agitated and current-influenced environments (Dott and Bourgeois, 1982; Soegaard and Eriksson, 1985; Southard and Boguchwal, 1990).

**Meter-scale cycles**

Mixed carbonate and siliciclastic facies, particularly in the central and western exposures of the Huns Member, are arranged into meter-scale (1-10 m) cycles. Two cycle-types are recognized: siliciclastic-rich cycles (Fig. B1) and grainstone-cycles (Fig. B2). The main distinguishing feature is the presence of siliciclastic mudstone and sandstone near the base of siliciclastic-rich cycles. Both cycle types may or may not have microbial layers at their base. Both cycle types commonly contain upward-coarsening grainstone near their tops. Both are commonly capped by layers of rudstone and grapestone containing large (3 to 5 cm) micrite-coated clusters of intraclasts. Locally, particularly in the westernmost exposures of the platform, grainstone cycles are capped by karst surfaces with potholes up to 1.5 m deep (Figs. B3 and C).

Evidence for karst development at the tops of meter-scale cycles suggests episodic periods of exposure when sea-level dropped below level of the carbonate platform. Based on U-Pb age constraints (Grotzinger et al., 1995; Chapter 4), an average sediment accumulation rate of between 300 and 400 m/m.y. can be calculated for the platform, yielding a duration of less than 100,000 years for each cycle. Consequently, high-frequency relative sea-level variation is interpreted to have been an important control on cycle development. As a whole, the coarsening upward nature of the cycles, which grade from
siliciclastic or carbonate mudstone to siliciclastic or carbonate sandstone followed by coarse grainstone, and the upward-increasing abundance of trough cross-stratification is interpreted to record increasingly higher-energy depositional environments upward and thus suggests a shallowing-upward asymmetry to the cycles, which are interpreted as shallowing-upward parasequences.

**Larger-scale parasequences**

A larger-scale (10's of meters) cyclicity is manifested by the alternation of siliciclastic-rich and carbonate-rich units in the eastern exposures of the Huns Member and in the Spitskop Member (Figs. A & C). The basal contacts of siliciclastic units are sharp. Commonly, underlying carbonate is iron stained and, rarely, dolomitized or associated with potholes and karst development. Similar to the siliciclastic-rich meter-scale cycles, siliciclastic mudstone grades upward into siliciclastic sandstone followed by limestone. The siliciclastic-to-carbonate transition is gradational over the scale of a meter. Decameter-scale siliciclastic-carbonate cycles in the eastern exposures of the Huns Member grades westward into carbonate-dominated meter-scale cycles that characterize the westernmost exposures of the member.

The sharp base to siliciclastic units, the coarsening-upward nature of the facies, and the gradational upward contact with limestone facies in both meter-scale and decameter-scale cycles is similar to other described siliciclastic-carbonate cycles (Aitken, 1966; Grotzinger, 1986, Chow and James, 1987). In many of these cases, the sharp surfaces at the bases of cycles have been interpreted as flooding surfaces and the cycles are interpreted to shoal upward. The abundance of sand in the cycles described here is somewhat enigmatic, however, because generally, siliciclastic-based, shallowing-upward cycles, contain only very fine siliciclastic material (Aitken, 1966; Grotzinger, 1986, Chow
and James, 1987). A possible mechanism to account for the sand is partial siliciclastic progradation during relative sea-level fall, followed by redistribution during flooding. As carbonate production outstepped the redistribution of the siliciclastic sand, carbonate facies replaced the siliciclastic facies. Thus, although, tectonically, or climatically driven changes in terrigenous influx may also have been important influences on siliciclastic progradation, the larger-scale cycles, similar to the meter-scale cycles, are interpreted, as shallowing-upward parasequences, bounded by flooding surfaces. Correlations of these larger-scale parasequences are shown in Fig. C.

In the original version of this paper it was suggested that the surface at the top of the Huns Member, above which the pinnacle reefs developed, may correspond to a sequence boundary in addition to being a flooding surface. Correlation of parasequences across the carbonate platform showed no evidence for an erosional unconformity developed at this surface, however, and it may be regarded, instead, as a surface of maximum flooding. A sequence boundary may separate predominantly coarse-grained meter-scale cycles, which lie above that flooding surface and record platform recovery, from non-cyclic deeper ramp facies, which predominate in the upper Spitskop Member at the top of the carbonate platform.

Origin of sea-level fluctuations

The abundance of meter-scale karst-capped cycles, the absence of tidal-flat facies and the presence of pinnacle reefs are all typical features of carbonate platforms developed during times of major continental glaciation and influenced by high-frequency, high-amplitude glacioeustatic sea-level fluctuations (Read, 1996). Although the possibility of tectonic effects from the bordering Gariep deformational belt can not be ignored, the depositional patterns of the
Huns/Spitskop carbonate platform are suggestive of glacioeustatic control on platform development.

References


Figure Captions

Fig. A: Cross section of the Kuibis and Schwarzrand Subgroups between the farms Swartkloofberg and Sontaagsbrun (see Fig. 1 in main body of chapter for location map). The Huns-Spitskopf carbonate is bounded by erosional unconformities at its base and at its top.

Fig. B: General characteristics of meter-scale cycles in the Huns-Spitskopf carbonate platform.

Fig. C: Representative measured sections along a cross-section of the Huns-Spitskopf carbonate platform showing proposed correlation of decimeter-scale cycles. Detailed measured sections are included in Appendix C at the end of this volume.
CHAPTER 3

THRUST-RELATED DEFORMATION OF THE NAMA GROUP IN SOUTHWESTERN NAMIBIA: RECONSTRUCTION OF AN IMPORTANT PROTEROZOIC-CAMBRIAN BOUNDARY SECTION

Abstract
Northwest-southeast striking northeast-vergent thrust faults divide exposures of the Kuibis and Schwarzzand subgroups on the farms Swartkloofberg (95), Swartpunt (74), Nord Witputs (22), Tierkloof (75) and Aub (81) into an autochthon and three allochthonous thrust plates. Broad wavelength basement-involved cross-folds create significant structural relief, exposing deep structural levels where the thrust faults merge and pass into basement. These thrust faults are considered to be the leading edge of the Gariep deformational belt.

The Proterozoic-Cambrian boundary, contained within a major unconformity near the top of the Schwarzzand Subgroup, is exposed in each of three thrust plates in the study area. In the lowest thrust plate, the unconformity lies above a 500 m thick section of the Spitskop Member, comprising shelf limestone and siliciclastic rocks. Exposures of the Spitskop Member in the middle and upper thrust plates consist of slope facies, are a maximum of 60 m thick and locally are completely eroded by the overlying Proterozoic-Cambrian boundary unconformity. Exposures of the Spitskop Member in each of the thrust plates are interpreted as widely geographically separated facies on a slope-to-basin transition which is telescoped by the thrust faults into a relatively small area. The total relief along the top of the Spitskop Member, more than 500 m, is a combination of depositional thinning across the shelf-to-basin transition and erosional incision along the Proterozoic-Cambrian boundary unconformity.

Introduction
This paper describes some of the structural and stratigraphic relationships involving the Vendian to Cambrian Nama Group on Nord Witputs (22), Swartkloofberg (95), and Swartpunt (74) farms in southwestern Namibia (Fig. 1). Exposures of the Nama Group in this area span the Precambrian-Cambrian boundary. Although the boundary is contained within a major erosional unconformity, U-Pb zircon geochronology on volcanic ash beds, combined with
global correlations based on biostratigraphy and carbon-isotope
chemostratigraphy (Chapter 4) indicate that the Vendian part of the section
extends to within 1 m.y. of the Cambrian System and contains, near its top, some
of the youngest known Ediacaran-type fossils (Fig. 2; Grotzinger et al., 1995).
Thus, the stratigraphic succession forms an important Precambrian-Cambrian
boundary reference section.

The study area lies along the eastern-limit of the syn-sedimentary Gariep
deforrmational belt (Davies and Coward, 1982; von Veh, 1988). Compressional
structures cross-cut the area, resulting in the repetition of stratigraphic units.
Fortuitously, the Precambrian-Cambrian boundary is repeated in three different
thrust plates (Fig. 2) which telescope exposures of spatially separate
paleogeographic domains into a relatively small and easily accessible region.
Here we report results of stratigraphic and structural studies aimed at
reconstructing the original succession through the Precambrian-Cambrian
boundary.

General Geology

The Nama Group consists of, in ascending order, the Kuibis, Schwarzrand
and Fish River subgroups (Germs, 1983). It was deposited in a foreland basin that
subsided in response to convergence along the Damara and Gariep
compressional belts (Germs, 1983; Germs and Gresse, 1991) and is deformed
along its northern and western margins by compressional structures related to
these belts (Martin, 1965; Martin, 1974; Coward, 1983; Miller, 1983). In southern
Namibia, mixed siliciclastic and carbonate rocks of the Kuibis and Schwarzrand
Subgroups thicken southwestward toward the Gariep belt, reaching their
maximum total thickness (more than 2000 m) in the study area where they are
deformed by thrust-related structures at the leading edge of the Gariep belt (Figs. 1 & 2).

The Precambrian-Cambrian boundary, as recognized on the basis of biostratigraphy and carbon-isotope chem stratigraphy, is contained within a regionally extensive erosional unconformity near the top of the Schwarzrand Subgroup (Grotzinger et al., 1995; Chapter 4). Ediacaran-type fossils have been discovered in the study area less than 60 m below this unconformity (Grotzinger et al., 1995). In addition, carbon-isotope data from the Ediacaran-fossil bearing section resemble carbon isotope profiles from terminal Neoproterozoic sections in Siberia (Pelechaty et al., 1996), arctic Canada (Narbonne et al., 1994), and other areas, reinforcing the terminal Neoproterozoic age inferred from the biostratigraphy (Fig. 2; Grotzinger et al., 1995; Chapter 4). Since typical latest Neoproterozoic isotope values extend up to the unconformity, with no evidence of the negative isotope excursion which underlies the boundary in other Proterozoic-Cambrian boundary sections, and since the overlying Nomtsas Formation contains Cambrian trace fossils (Germs, 1983; Grotzinger et al., 1995), the Proterozoic-Cambrian boundary is inferred to be contained within the unconformity.

A U-Pb zircon age of 543.3±1 Ma for a volcanic ash bed located 130 m below the Proterozoic-Cambrian boundary unconformity and 90 m below the Ediacaran fossils is a maximum for both the age of the fossils and the age of the Precambrian-Cambrian boundary in the study area (Grotzinger et al., 1995). This age is the same, within the limits of analytical error, as the age of lowermost Cambrian rocks (543.8±5.1/-1.3) in Siberia (Bowring et al., 1993), making the Ediacaran fossils in the study area some of the youngest known representatives of their kind (Grotzinger et al., 1995). A U-Pb zircon age of 538.8±1 Ma for a volcanic ash bed just above the unconformity is a minimum for the age of the
Proterozoic-Cambrian boundary in Namibia and constrains the duration of the unconformity to less than 5 m.y (Grotzinger et al., 1995).

Stratigraphy

The sedimentology and stratigraphy of the Kuibis and Schwarzrand subgroups have been described and interpreted elsewhere (Germs, 1983; Saylor, 1993; Saylor et al., 1995; Chapter 2), and the specifics of the stratigraphy in the study area are only briefly described here. Discussion focuses on stratigraphic units near the Proterozoic-Cambrian boundary and how these units change across each of three thrust plates in the study area.

Kuibis Subgroup

Exposures of the Kuibis Subgroup are restricted to the southwestern part of the map area and to the lower and upper thrust plates (Fig. 1). The Kuibis Subgroup overlies local outcrops of unnamed stratigraphic units consisting of older diamictite and cream-colored dolostone, or nonconformably overlies crystalline basement. The Kuibis Subgroup is almost 300 m thick and comprises two units of coarse sandstone (Kanies and Kliphoek Members), and two units of dominantly fine-grained carbonate (Mara and Mooifontein Members) (Saylor, 1993; Saylor et al. 1995; Chapter 2). The siliciclastic-dominated units are interpreted to have formed in fluvial to margin-marine environments and the carbonate-dominated units in shallow-subtidal settings (Germs, 1983; Saylor et al., 1995).

Lower Schwarzrand Subgroup

The lower Schwarzrand Subgroup, comprising the Nudaus Formation and the Nasep Member of the Urusis Formation, crops out across much of the map
area (Fig. 1). It is 400 m thick and is dominated by siliciclastic mudstone and sandstone, interpreted to have formed in mid-shelf to near-shore and deltaic environments (Germs, 1983; Saylor, 1995, Chapter 2). The Nudaus Formation is 150 m thick and consists of sandier-upward parasequences comprising green mudstone with intercalated thin- to medium-beded sandstone. The upper Nudaus Formation and the lower Nasep Member form a 60 m interval comprising 5 to 20 m thick-units of medium-beded sandstone, shale and grainstone. The middle 100 m of the Nasep Member consists of green shale, and the upper 60 m consists of cross-beded and planar laminated, locally slumped, fine sandstone.

Upper Schwarzrand Subgroup

The middle part of the Schwarzrand Subgroup, comprising the Huns, Feldschuhhorn and Spitskop Members of the Urusis Formation, is interpreted as a carbonate ramp succession (Saylor et al., 1995, Chapter 2). It reaches a total thickness in the study area of nearly a kilometer. The Precambrian-Cambrian boundary unconformity lies at the top of the Spitskop Member and cuts down section, so that the overlying Cambrian Nomtsas Formation rests on progressively lower strata of the Spitskop, Feldschuhhorn and Huns members. The details of this erosional incision and changes in the underlying carbonate platform are pieced together here from structurally and geographically isolated outcrops in each of the three thrust plates.

Huns Member

The Huns Member is a nearly 300 m thick unit formed predominantly of limestone. The basal 40 m of the Huns Member is recessive, comprising shale with limestone and sandstone interbeds. The remainder consists of meter-scale,
upward-shallowing cycles (Chapter 2). A few siliciclastic units are correlatable across the study area, but are never more than a few meters thick. The carbonate cycles consist of elongate, dolomitized stromatolitic columns overlain by upward-coarsening cross-stratified pellet and intraclast grainstone and commonly are capped by karst-surfaces. They are interpreted to have formed in a high-energy, wave-swept shoal area that was strongly influenced by high frequency relative sea-level oscillations (Saylor et al., 1995; Chapter 2). A unit of pink lime mudstone, with meter-high thrombolitic and stromatolitic columns and domes, and intercalated green shale forms the top of the Huns Member. This distinctive unit is recognizable and correlatable across the map area (Fig. 4).

Reefs

Exhumed pinnacle reefs are preserved in the middle thrust plate (Fig. 1). Initially these reefs were interpreted as part of the Cambrian Nomtsas Formation and were thought to have grown in an erosional valley that incised through the Spitskop Member (Germs, 1983). However, they lie above light-colored stromatolitic limestone similar to that at the top of the Huns Member and locally lie below dark, thin-bedded Spitskop limestone. Consequently, they have been reinterpreted as part of the Huns Member (Figs. 3 & 4; Saylor, 1993; Saylor et al., 1995, Chapter 2).

In one location (Fig. 5; see Fig. 1 for location) two pinnacle reefs are enveloped in shale and conglomerate of the Cambrian Nomtsas Formation. They project from a layer of stromatolitic limestone which appears to abut Spitskop limestone along the steep wall of an incised valley. However, the juxtaposition against the valley wall is not stratigraphic. Instead, it is the result of structural repetition along a small normal fault with only a few meters of offset. The stromatolitic layer is part of the Feldschuhhorn Member. In the hanging wall of
the normal fault the layer can be traced to where it underlies the Spitskop Member. Pinnacle reefs at this level resemble the stromatolitic layer from which they project and, rather than recording a second period of reef nucleation, may have formed by mantling of stratigraphically lower pinnacles. Much like their present exposure, the reefs most likely formed resistant structures that were exhumed from surrounding shale during canyon incision. The reefs were covered again with conglomerate as the canyon filled with Nomtsas Formation. Nowhere do the stratigraphic relationships require that any of the pinnacle reefs lie within the Nomtsas Formation (Saylor et al., 1995; Chapter 2). In contrast, they are all considered to have developed from a single horizon at the top of the Huns Member during a period of reduced carbonate production following flooding and drowning of the platform.

*Feldschuhhorn Member*

The Feldschuhhorn Member is a 60 m thick green-shale unit. It stratigraphically overlies and envelopes the pink stromatolitic limestone and pinnacle reefs at the top of the Huns Member (Figs. 1 & 3). Light-colored stromatolitic limestone mantles the pinnacle reefs. The upper part of the Feldschuhhorn Member contains interbedded sandstone and grades upward into black, fine-grained limestone of the Spitskop Member.

*Spitskop Member*

The Spitskop Member in the lower thrust plate has not previously been described in any detail. It is 500 m thick (Fig. 4c) and consists of alternating carbonate and siliciclastic units, each of which is several 10's to 100 or more meters thick. The lower carbonate units consist largely of meter-scale, karst-capped cycles, which, similar to cycles of the Huns Member, are interpreted to
have formed in a high-energy wave-swept environment. Higher carbonate units consist of thin-bedded, fine-grained limestone with local development of thrombolitic and stromatolitic domes and the upper carbonate units consist almost entirely of thick-laminated to thin-bedded limestone with rare beds of flat-pebble, intraclast breccia, commonly with mounded tops. With the exception of intraclast breccias, which are interpreted to have formed during occasional storms, which ripped-up and reworked the thin-bedded limestone (e.g. Sepkoski, 1982), higher carbonate units show little evidence for the influence of strong waves or currents. They resemble Cambrian "ribbon-rock" and are similarly interpreted to have formed in low-energy, shallow- to deeper-subtidal environments on a carbonate ramp (Demicco, 1983; Cowan and James, 1993). Thick siliciclastic units form coarsening-upward successions, each with green mudstone at the base followed by ripple-laminated or thick-bedded, planar-laminated and hummocky cross-stratified very fine to fine sandstone at the top. These siliciclastic facies are interpreted to have been deposited in outer- to mid-shelf environments.

The Spitskop Member in the middle and upper thrust plates (Figs 4a, b) is less than 60 m thick. It consists of fine-grained, black, thick-laminated to thin-bedded limestone, breccia and shale. The breccia beds are composed of platy clasts that resemble the surrounding limestone. Breccias range from incipient fracture zones, to fully-developed, matrix-supported, disorganized debrites. The relative proportion of breccia, particularly debrites, and shale increases westward. These sections of the Spitskop Formation resemble the Cow Head Formation of Newfoundland (e.g. James and Mountjoy, 1983) and are interpreted to have formed by slope failure and mass-wasting on a carbonate slope (Chapter 2).
The combination of ramp-type facies and slope-type facies that constitutes the Spitskop Member is characteristic of distally-steepened ramps (Read, 1985). Exposures of the Spitskop Member in the study area are interpreted as representative sections along a shelf to basin transition, extending from outer ramp to slope positions (e.g. James and Mountjoy, 1983), which have been structurally telescopied into a relatively small area.

**Nomtsas Formation**

Exposures of the Nomtsas Formation are restricted to the northern part of the map area. They overlie and form the fill of erosional canyons incised through the Spitskop Member.

Outcrops of the Nomtsas Formation in the lower thrust plate are inferred to overlie the Spitskop Member, but the contact is covered. Exposures of the stratigraphically underlying Spitskop Member near the covered zone are dolomitized and brecciated. The dolomitization and brecciation are local features, however, and this horizon can be traced laterally to intact limestone approximately 120 m below the top of the Spitskop Member (see Fig. 1). It is unclear whether the Nomtsas Formation lies directly on this dolomitized horizon, or if the remaining 120 m of the section is buried beneath intervening Quaternary alluvium. However, based on the similarity of this horizon to a dolomitized and brecciated horizon which directly underlies exposures of the Nomtsas Formation in the middle thrust plate, we suggest that it may correspond to the Spitskop-Nomtsas contact and that the basal erosion surface of the Nomtsas Formation in the lower thrust plate may have cut down through as much as 120 m of the Spitskop Member (Fig. 4c).

In the middle structural level, the Nomtsas Formation overlies a 60 m thick section of the Spitskop Member and locally the basal erosion surface has cut
entirely through the member down to the level of pinnacle reefs at the top of the Huns Member (Fig. 4b). Stromatolitic and thrombolitic carbonate at the top of the Spitskop exposures and immediately underlying the Nomtsas Formation is dolomitized and characterized by extensive development of fitted breccias.

Nomtsas successions in the middle and upper thrust plate are lithologically and stratigraphically similar comprising matrix- and clast-supported conglomerate, overlain by shale-rich diamictite, followed by clast-free mudstone with interbedded sandstone. Disorganized and locally inverse-stratified boulder and pebble conglomerate in a sandy matrix and shale-rich diamictite with outsized clasts as large as 3m across are interpreted to have formed by debris-flow and slump processes (Saylor, 1993; Saylor et al., 1995; Chapter 2). Conglomerate clasts, which include limestone, dolostone, sandstone and volcanic ash, are lithologically similar to stratigraphically lower units and are interpreted as debris shed from the flanks of incised valleys (Saylor, 1993; Saylor et al., 1995; Chapter 2). Clasts similar to the dolomitized unconformity surface may indicate that dolomitization preceded valley infilling.

In the western part of the map area the surface corresponding to the sub-Nomtsas unconformity may be represented by an abrupt change in conglomerate facies from Spitskop Member slope-breccia, composed of platy-clasts of deep-water slope-facies limestone, to megabreccia with large (up to 3 m) boulders of stromatolitic, thrombolitic and massive fine-grained dolostone and limestone similar to the more shallow-water ramp facies (Fig. 4a). Slope breccias on distally-steepened ramps such as the Spitskop Member generally consist entirely of deep-water clasts (James and Mountjoy, 1983). Thus, although the megabreccia has no sand in the matrix and differs from other exposures of the Nomtsas Formation conglomerate, the change to more shallow-water clasts probably records shelf-edge exposure and erosion (e.g. Hiscott and James, 1984) during
relative sea-level fall and the initial development of the up-dip, sub-Nomtsas unconformity.

**Structure**

Three northwest-southeast oriented, northeast-vergent thrust faults transect the map area dividing it into an autochthon and three main thrust plates (Fig. 2). These faults are interpreted as the frontal thrusts along the leading edge of the Gariep orogenic belt. Younger, (Permian?) basement-involved deformation has resulted in broad upwarps with significant structural relief, such that the deepest structural levels are exposed in the southeast part of the map area (fig. 1), where the two upper faults merge and root into basement.

This study focused on the upper two thrust faults. Structure along the eastern and western borders of the map area was mapped only in a reconnaissance fashion based largely on air photo interpretation.

The lowermost thrust plate contains a stratigraphic section extending from the top of the Kuibis Subgroup up to the Nomtsas Formation, and including the 500 m thick section of the Spitskop Member (Figs 1 & 4c). The middle and upper fault planes merge so that the middle thrust plate is only locally exposed. It appears in the middle of the map area (Fig. 1) as a half-klippe of Huns Member thrust over Spitskop Member and overthrust by lower Schwarzrand rock (Nasep Member; Fig. 6). Toward the northwest the two thrust faults cut up section and splay apart. They are inferred to extend northwestward on either side of the array of pinnacle reefs (Figs 1 and 4b). Toward the southeast the middle and upper thrust faults merge and root in basement indicating their thick-skinned nature. The uppermost thrust plate contains a complete stratigraphic section extending from basement up to the base of the sub-Nomtsas unconformity (Fig. 4a).
Rocks in the upper thrust plate are deformed by numerous linked folds (Figs 1 & 2). The folds parallel the thrust faults, plunge to the northwest, and are generally open and upright, but northeast-vergent. Locally the folds are isoclinal or are cut by small-displacement (a few meters) thrust faults. The middle thrust plate contains an anticline and is itself folded. Deformation in the lower thrust plate is less intense. Near the bounding faults there are a few fault-parallel folds.

Fig. 7 shows a possible balanced cross-section (Suppe, 1983) through the middle of the map area (see Fig. 1 for location). This cross-section meets the stratigraphic constraint of successively more distal exposures westward and the structural constraint of merging, basement rooted faults (A and B).

The oldest fault in the area is fault B, which forms the base of the middle thrust plate. The upper thrust plate lies in a splay (fault A) of fault B. Later motion and development of a thrust ramp along fault C may have folded faults A and B. The horizontal offset across fault C is approximately 1 km. Since fault A is a splay which postdates fault B, the offset along A is restored by matching the hanging wall cutoff of the Huns Member with the footwall cutoff of the Huns Member in the hanging wall of fault B (Fig. 7b). The horizontal displacement is 1 km. A hanging wall anticline in the Huns Member of the middle thrust plate is a tie point for restoring fault B which has a horizontal offset of 3 km.

Paleogeographic restoration

The stratigraphic succession from the Kuibis Subgroup up through the Huns Member of the Schwarzrand Subgroup shows little change across the three thrust plates. Thicknesses remain approximately the same and distinct beds and marker horizons can be recognized across the map area. The Spitskop Member, however, changes significantly. In the lower thrust plate it is 500 m thick, but in the middle and upper thrust plates it is 0 to 60 m thick. In the lower thrust plate
the Spitskop Member consists of outer ramp limestone and deltaic sandstone, and in the upper two thrust plates it consists of deeper-water limestone, slope breccia and shale.

These substantial differences are not the result of modern erosion patterns because in all thrust plates contacts with the underlying Feldschuhhorn Member and the overlying Nomtsas Formation are preserved. Instead, similar to the shelf margin of the Permian Grayburg Formation (Franseen, 1989) in west Texas, these differences reflect some combination of stratigraphic thinning and facies change across the shelf-to-basin transition and erosional truncation along a basinward-dipping unconformity beneath the Nomtsas Formation (Fig. 7).

The total amount of relief along the top of the Spitskop Member, 500 m over a palinspastically restored distance of 8 km, is similar in scale to, although somewhat greater than the relief across the Grayburg shelf margin (350 m over 4.5 km). Over a distance of 5 km in the lower thrust plate, there may be as much as 130 m of erosional incision along the sub-Nomtsas unconformity. In the middle thrust plate, there is at least 60 m of incision expressed by the steep walls of the unconformity surface (Fig. 3 b, c). Consequently, the remaining 310 m of relief must be accounted for over the distance (3 km) represented by the horizontal displacement of the lower thrust fault.

There is essentially no depositional thinning in the upper thrust plate and most of the shelf-to-basin transition has been cut out along the lower thrust fault. Since no marker horizons in the Spitskop Member can be correlated across this fault, the proportion of the 310 m of relief which is represented by pre-erosion depositional thinning across the shelf-to-basin transition is unknown. Assuming the maximum possible amount of depositional thinning, 310 m across the palinspastically restored distance of 3.5 km, the calculated depositional slope is 6°. Although somewhat high, 6° is reasonable for the distally steepened part of a
ramp and is consistent with the abundance of breccias in the slope sections (Read, 1985). The original depositional profile may have been much less steeply inclined, however, because slopes of only 2° are sufficient to cause slope failure. Since erosional incision effected both shelf exposures in the lower thrust plate and basinal exposures in the middle thrust plate, incision and the development of an erosional escarpment probably modified the depositional shelf-to-basin profile substantially.

It is uncertain whether the erosional relief across the shelf margin developed everywhere by sub-aerial exposure and fluvial incision or, if, like the Grayburg escarpment, some parts, particularly those down-dip, developed by submarine erosion and mass wasting. The unconformity surface is recognized 100 km east (cratonward) of the study area where it is associated with incised canyons a few tens of meters deep. There, the canyons are partially infilled by planar-stratified conglomerate and trough cross-stratified sandstone (unpublished data), facies which record strong current influence. The great lateral extent of erosional topography developed along this unconformity surface and the evidence for strong currents are hard to account for by mass wasting in an exclusively sub-marine environment so that the more proximal canyons are tentatively interpreted to have formed by fluvial incision. Facies of the Nomtsas Formation in the lower thrust plate, above shelf facies of the Spitskop Member, and in the middle thrust plate, above slope facies of the Spitskop Member, are strikingly similar and seem to have been deposited in similar depositional environments, suggestive of deposition during diachronous transgression and back-filling of the escarpment, following a large relative sea-level drop. The coarse conglomerate and diamicite facies consists of debris derived from the escarpment and deposited by sediment gravity flows, probably in a deeper marine environment consistent with the preceding slope setting. Fine
conglomerate and sandstone facies may have been introduced by mass wasting of facies derived from upslope estuarine environments (Saylor et al., 1995; Chapter 2). Dolomitization fronts and mantling breccias associated with the unconformity in both the lower and middle thrust plates in the study area may have been formed by mixing of meteoric and marine waters following exposure and karst weathering (e.g. Badiozamani, 1973; James and Choquette, 1987), but may also have been formed much later by fluids that traveled through the porous Nomtsas Formation (e.g. Mussman et al., 1987). Support for early (pre-Nomtsas) dolomitization is provided by the presence of dolomitized clasts within the Nomtsas Formation.

The Proterozoic-Cambrian boundary unconformity is interpreted to record a major fall in relative sea-level. A minimum estimate for drop in relative sea-level is a few tens of meters, sufficient to expose and incise updip portions of the unconformity. If dolomitization along downdip exposures of the unconformity surface is related to karst processes, relative sea-level fall might have been 500 m or more.

The Nama basin was a tectonically active foreland basin (Germs, 1983) and tectonic uplift associated with thrust deformation may have been an important factor in the development of the erosional unconformity. Stratigraphic patterns throughout the Schwarzrand Subgroup, including westward-thickening stratigraphic units and eastward downcutting erosional unconformities, indicate tectonically-driven flexural subsidence of the western part of the basin and slower subsidence and episodic uplift of the eastern part of the basin (Germs and Gresse, 1991; Gresse and Germs, 1993; Chapters 2). Erosion along the eastern extent of the Proterozoic-Cambrian boundary unconformity surface may be related to uplift of a flexural bulge (Germs and Gresse, 1991). Evidence for rapid subsidence of the western part of the basin, however, indicates that it lies in
front of the bulge and that development of the unconformity in the study area can not be explained by bulge uplift. The development of an erosional unconformity in both the rapidly subsiding western part of the basin and above the slowly subsiding flexural bulge may best be explained by a eustatic sea-level fall. The recognition of approximately coeval unconformities at the Proterozoic-Cambrian boundary in Siberia (Pelechaty et al., in press), Canada (Narbonne et al., 1994) and the western United States (Cooper and Fedo, 1995) demonstrates the global nature of this unconformity and provides additional evidence that the sea-level fall may be eustatic in nature (Runnegar et al., 1995).

Conclusions

1) Thrust-related deformation divides exposures of the Kuibis and Schwarzzrand Subgroups on the farms Swartkloofberg (95), Swartpunt (74), Nord Witputs (22) Tierkloof(75) and Aub(81) into an autochthon and three thrust plates, separated by northwest-southeast striking, northeast-vergent thrust faults. The thrust faults, which root into basement, are interpreted as the leading edge of the Gariep deformatinal belt. The Proterozoic-Cambrian boundary, contained within a major unconformity near the top of the Schwarzzrand Subgroup, is exposed in each of the thrust plates.

2) The lower thrust plate contains a nearly complete stratigraphic section extending from the top of the Kuibis Subgroup up to the Cambrian Nomtssas Formation at the top of the Schwarzzrand Subgroup. The unconformity at the base of the Nomtssas formation lies above a 500 m thick section of the Spitskop Member and has erosional relief of more than 130 m.

3) The middle thrust plate contains only the middle and upper part of the Schwarzzrand Subgroup. Pinnacle reefs are restricted to the middle thrust plate. The reefs lie structurally above the upper Spitskop Member and were originally
interpreted as part of the Nomtsas Formation, but have been reinterpreted to lie stratigraphically at the top of the Huns Member. Exposures of the Spitskop Member above the pinnacle reefs in the middle structural level are a maximum of 60 m thick and locally the Proterozoic-Cambrian unconformity cuts down to the level of the reefs.

4) The upper thrust plate contains a complete section from basement up to conglomerate thought to correlate with the sub-Nomtsas unconformity. The Spitskop Member is a maximum of 60 m thick and consists of toe of slope facies.

5) The character of the Spitskop Member changes across the structural levels from ramp to slope to toe of slope facies. These facies formed on a shelf-to-basin transition along a distally steepened ramp which was transected and telescoped into a relatively small area by the thrust faults. The total relief along the Proterozoic-Cambrian unconformity, more than 500 m, is a combination of depositional thinning and erosional incision across this transition.

6) The Proterozoic-Cambrian boundary unconformity in Namibia developed during a major fall in relative sea-level. Development of significant erosional topography along the downdip exposures of the unconformity surface, well in front of the flexural bulge may be explained best by a eustatic sealevel fall. This interpretation is strongly supported by Runnegaret et al. 's (1995) suggestion that approximately coeval falls in relative sea-level in Siberia, Canada and the western United States indicate a eustatic sealevel event at this time.

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Figure Captions

Figure 1: Geologic map of the farms Swartkloofberg, Swartpunt and NordWitputs. Inset shows location of map area in southwestern Namibia relative to exposures of the Kuabis and Schwarzrand subgroups (shaded).

Figure 2: Tectonic map of the study area showing the autochthon and the three thrust plates.

Figure 3: General stratigraphic column for the Kuabis and Schwarzrand subgroups in the region near the study area (see Chapter 4 for stable isotope data and Grotzinger et al., 1995 for radiometric age constraints). See Fig. 4 for key.

Figure 4: Measured sections from the top of the Huns Member (a distinctive pink stromatolitic unit) up to the Nomtsasas Formation for the lower, middle and upper thrust plates. See appendix for detailed measured stratigraphic sections through the Huns and Spitskopf Members in each thrust sheet.

Figure 5: Line drawing of structural and stratigraphic relationships involving pinnacle reefs on the farm Swartkloofberg (95). See text for discussion and Fig. 1 for location. (H: upper Huns Member; R: pinnacle reefs in Huns Mbr.; F: Feldschuhhorn Mbr.; S: Spitskopf Mbr.)

Figure 6: Photograph of the middle and upper thrust plates. Siliciclastic rocks of the Nasep Member are thrust over pinnacle reefs, shale and dark carbonate of the upper Huns, Feldschuhhorn and Spitskopf members. Tick marks on hanging wall. (H: upper Huns Member; R: pinnacle reefs in Huns Mbr.; F: Feldschuhhorn Mbr.; S: Spitskopf Mbr.)

Figure 7: a) Coss-section showing the lower, middle and upper thrust plates. Locations of cross-section shown in fig. 1. b) Partially restored cross-section showing geometry prior to movement along fault C and splay A.

Figure 8: Schematic cross section through the shelf-to-basin transition of the Spitskopf Member. The total relief along the Precambrian-Cambrian boundary, which is shown by the thick, solid line, is a combination of depositional thinning and erosional truncation of the shelf margin. Key in Fig. 1.
CHAPTER 4

THE PARTITIONING OF TERMINAL NEOPROTEROZOIC TIME: CONSTRAINTS FROM NAMIBIA

ABSTRACT

Integrated sequence stratigraphic and chemostratigraphic data yield a high resolution framework for correlations of stratigraphic units in the terminal Proterozoic to Cambrian Witvlei and Nama Groups of Namibia. Coupled with precise U-Pb zircon age constraints, these correlations make it possible to construct a composite reference section for use in calibrating terminal Proterozoic chronostratigraphy. The Namibian reference section starts with two distinct glacial horizons and extends up to within 1 million years of the Proterozoic-Cambrian boundary. The two glacial horizons may represent each of two distinct Varanger-age glaciations better known from the north Atlantic region. From the higher of the two glacial horizons up, the composite stratigraphy preserves one of the thickest and most complete available records of carbon-isotope excursions in post-Varanger terminal Proterozoic seawater. Four carbon-isotope chemostratigraphic intervals are recognized: 1) a post-glacial negative δ¹³C excursion (Npg interval); 2) a rising interval (Pᵢ interval) of increasing positive δ¹³C values; 3) a falling interval (Pᵢ interval) characterized by decreasing positive δ¹³C and culminating in near zero or negative values; and 4) an interval of moderately positive, relatively invariant δ¹³C values (I interval) that extends up to the unconformity that contains the Proterozoic-Cambrian boundary. Each of these chemostratigraphic intervals can be recognized in widely separated equivalent sections around the world. By comparing sediment accumulation rate in the radiometrically calibrated Namibian stratigraphy with sediment accumulation rates in correlative sections in Arctic Canada and Oman a maximum age of approximately 565 Ma is estimated for the end of the younger Varanger glaciation. This age is 25 m.y. younger than previous age estimates.

INTRODUCTION

Historically, the rapid diversification of Earth's early animals was thought to have begun during an interval of equitable climate tens of millions of years after the major ice sheets of the Varanger ice age had waned
and millions of years before the dawn of the Cambrian Era (Glaessner, 1984). Recent discoveries of Ediacaran-type fossils within radiometrically dated beds in Namibia (Grotzinger et al., 1996), however, have extended the range of these fossils to within one million years of the Proterozoic-Cambrian boundary. In addition, strong secular variations in the carbon- and strontium-isotopic composition of terminal Proterozoic seawater (now preserved in sedimentary carbonate and organic C) have made it possible to correlate widely separated sections around the world, providing a framework for comparison of fossil ranges and glacial horizons (Knoll and Walter, 1992; Kaufman and Knoll, 1995; Kaufman et al., in press). Based on these isotopic correlations two distinct episodes of Varanger glaciation have been distinguished (Kaufman et al., in press), so that it now appears that the range of diverse Ediacaran fossils in Australia and Canada may extend down to, or below, the level of the younger Varanger glaciation, even though direct evidence for ice-sheet development is limited or absent (Pell et al., 1993; Jenkins, 1995; Kaufman et al., in press).

Unfortunately, the timing of these climatic and biologic events is poorly constrained. U-Pb zircon ages from volcanic rocks lying thousands of meters below (606+3.7/-2.9 Ma; Krogh et al., 1988; 602±3 Ma; Kaye and Zartman, 1980) and thousands of meters above (565±5 Ma; Benus, 1988) inferred Varanger-equivalent tillites in the Avalon terrane of Newfoundland and Massachusetts bracket one of the Varanger glaciations. In the absence of a chemostratigraphic framework for these tillites, however, it is unclear which of the two Varanger glaciations they represent.

In this paper sequence stratigraphy (Christie-Blick et al., 1988; Posamentier and Vail., 1988; Sarg, 1988; Van Wagoner et al., 1988) and carbon-isotope chemostratigraphy are combined to correlate stratigraphic units of
terminal Proterozoic age on the Kalahari craton of Namibia. The composite reference section extends from post-glacial transgression at the end of the younger Varanger ice age up to within a million years of the Proterozoic-Cambrian boundary. The resulting isotope profile is directly tied to U-Pb zircon ages (Grotzinger et al. 1995) and hence places constraints on the ages of other terminal Proterozoic successions world-wide. These data significantly compress the duration of key features of terminal Proterozoic chronostratigraphy relative to previous estimates and indicate that the end of the Varanger glacial epoch may be much younger than originally inferred. Thus, by comparing the temporally calibrated Namibian record with coeval successions on other continents the timing of the end of younger Varanger ice age is estimated.

GEOLOGY AND AGE CONSTRAINTS

Terminal Proterozoic strata on the Kalahari Craton of Namibia, comprise the Witvlei and lower Nama groups and their correlatives. These units contain many of the key features of known terminal Proterozoic chronostratigraphy, including glacial horizons (Hoffmann, 1989), a diverse fossil record containing both Ediacaran-type fossil imprints and calcified skeletons (Crimes and Germs, 1982; Germs, 1983; Germs et al., 1986; Grant, 1990; Grotzinger et al., 1995a), and strong carbon- and strontium-isotopic excursions preserved in carbonate rocks (Kaufman et al., 1991; Derry et al., 1992; Kaufman et al. 1993; Kaufman et al., in press).

The Witvlei Group, interpreted as a passive margin succession, is defined in the area near Gobabis (Fig. 1; Hoffmann, 1989; Hegenberger, 1993). The basal Court Formation lies directly above glacial diamicomite of the Blaubeker Formation. An unconformity between the Court Formation and
the overlying Buschmannsklippe Formation is thought to correlate laterally with a second glacial diamicite (Hoffmann, 1989; Hofmann et al., 1995), which has been attributed to the Varanger ice age (Hoffmann, 1989, Kaufman et al., in press). This younger glacial horizon, the Bläskranz Formation, is known to be preserved only in the Naukluft Nappe Complex, a series of thrust nappes containing Witvlei- and Nama-equivalent strata that been emplaced above autochthonous strata of the Nama Group (Figs. 1, 2a and 2b).

The fossiliferous Nama Group overlies the Witvlei Group in the Gobabis area, but extends well beyond the limits of Witvlei deposition (Fig. 1). Much of the Nama Group was deposited in a foreland basin that developed during convergence along the bordering Damara and Gariep compressional belts (Germs, 1983; Gresse and Germs, 1993; Germs, 1995). During deposition of the lower Nama Group, the Nama basin was partitioned into northern and southern sub-basins by a basement arch culminating near Ösis (Germs, 1983).

The Proterozoic part of the Nama Group comprises the Kuibis and most of the Schwarzrand subgroups. Typical terminal Proterozoic fossils, including Ediacaran-type fossils (Germs, 1972; Germs, 1983; Grotzinger et al. 1995), Vendotaenids (Germs et al., 1986), and Cloudina (Germs, 1983, Grant, 1990), plus goblet-shaped calcified fossils (Grotzinger et al. 1995) are present both north and south of Ösis (Fig. 2b-d). The highest known Ediacaran-type fossils lie 130 m below the Proterozoic-Cambrian boundary, which is contained within a regionally extensive erosional unconformity near the top of the southern Schwarzrand Subgroup (Grotzinger et al., 1995; Chapter 3). Overlying units contain Phycodes pedum and other Cambrian-type trace fossils (Germs, 1983; Geyer and Uchman, 1995; Grotzinger et al., 1995).

U-Pb zircon ages for ash beds intercalated in the Nama Group provide high-resolution absolute age control (Grotzinger et al., 1995). The oldest dated
ash bed, which is from the northern Kuibis Subgroup, yielded an age of 548.8±1 Ma (Fig. 2c). Two ash beds from 350 m and 130 m below the Proterozoic-Cambrian boundary unconformity in the southern Schwarzrand Subgroup, yielded ages of 545.5±1 Ma and 543.3±1 Ma respectively (Fig. 2d). Both of these ash beds lie below recently discovered horizons of Ediacaran-type fossils (Grotzinger et al., 1995), and the younger is the same age, within the limits of error, as lowermost Cambrian volcanic rocks in Siberia (Bowring et al., 1993), suggesting that the Proterozoic part of the succession extends to within one million years of the Proterozoic-Cambrian boundary. The youngest dated ash bed in Namibia is from a few meters above the Proterozoic-Cambrian boundary unconformity in the southern sub-basin; it yielded an age of 538.8±1 Ma (Fig. 2d).

STRATIGRAPHY

Sequence stratigraphic and carbon-isotope chem stratigraphic data sets are described independently for successions in each of four geographic areas on the Kalahari craton of Namibia: 1) the Witvlei Group and the Kuibis Subgroup in the Gobabis area; 2) equivalents of the Witvlei Group in the Naukluft Nappe Complex; 3) the Kuibis Subgroup in the main exposures of the northern Nama sub-basin; and 4) the Kuibis and Schwarzrand subgroups in the southern Nama sub-basin. Isotopic and elemental analysis followed the methods described in detail by Kaufman et al. (1991), Derry et al. (1992), Narbonne et al. (1994) and Kaufman and Knoll (1995). Isotope data from this study (Table 1) are supplemented by data from Kaufman et al. (1991; Table 2), which are regrouped according to location and plotted in approximate positions relative to measured stratigraphic sections. Together, the integrated sequence stratigraphies and carbon-isotope chem stratigraphies provide high-
resolution time-lines for correlation of stratigraphic units between each of the geographic areas.

**Sequence Stratigraphy: Gobabis Area**

**Court Formation.**—The Court Formation (generalized section shown in Fig. 2a) is limited in lateral extent and pinches out to the north and west, where it is overstepped by the Buschmannsklippe Formation. It is a maximum of 200 m thick and consists of thin-laminated carbonate rhythmite at the base, overlain by mixed shale, sandstone and dolostone, followed by coarse sandstone with locally developed conglomerate and stromatolitic dolostone (Hegenberger, 1993). Hegenberger (1993) recognized that sedimentary features of the Court Formation, such as carbonate rhythmite, intraclast conglomerate, ooid grainstone, and stromatolitic dolostone are typical of a marine environment, but suggested that the Court Formation may have had a limited original extent, and thus concluded it may have been deposited in a restricted or lacustrine basin. The Court Formation is capped by an erosional unconformity and has been interpreted by Hegenberger (1993) as a single, complete depositional sequence. Based on lithostratigraphic correlation Hoffman (1989) suggested that the unconformity at the top of the sequence correlates with glacial diamicrite of the Bläskranz Formation, preserved in allochthonous strata of the Naukluft Nappe Complex, more than 100 km to the southwest.

**Buschmannsklippe Formation.**—The Buschmannsklippe Formation thickens northwestward across the Gobabis area to a maximum of 200 m (Fig. 2a). At the base of the formation, the Bildah Member consists of homogeneous, fine-grained, massive and finely laminated pink/tan dolostone (Hegenberger,
1993). Laminae define distinctive meter-high, irregular domes and slump structures and are truncated by vertical, spar- and sediment-filled tubes, up to a few centimeters across and tens of centimeters long. Hegenberger (1987) interpreted these tubes as gas-escape structures and compared them to similar tube-like structures in the terminal Proterozoic Noonday Formation of the western U.S. The Bildah Member grades upward into the La Fraque Member, which is composed of thinly bedded, ripple cross-stratified pink/tan limestone and purple siltstone and sandstone. In-situ, precipitated layers of upward-divergent calcite fans are present near the base of the member. The high concentration of strontium in these fans (>1000 ppm; Kaufman et al., 1993) suggests they were most likely aragonite before neomorphism. Edgewise-oriented intraclast conglomerate and hummocky cross-stratified sandstone beds increase in abundance upward through the La Fraque Member into the lower Okambara Member. The Okambara Member, in turn, grades upward into ooid-intraclast grainstone, followed by trough cross-stratified ooid grainstone interbedded with digitate stromatolites, microbialaminite with cauliflower chert (probably replaced anhydrite, e.g. Milliken, 1979), and mud-cracked and quartz-rich dolarenite. The contact with coarse sandstone of the basal Nama Group (Weissberg Member) is gradational, consisting of interbeds of carbonate, sandstone with wave ripples, and thin mudstone drapes with desiccation cracks.

The Bildah Member was deposited primarily in a low-energy, lagoonal (Hegenberger, 1993) or sub-tidal ramp setting (e.g. Read, 1985). Crystal fans of the La Fraque Member precipitated directly on the sea floor (e.g. Grotzinger and Knoll, 1995) and may correspond to the zone of maximum flooding, formed during a period when both carbonate and terrestrial clastic sediment influx was reduced. The homogeneity of the fine-grained carbonate, the tube-
like structures, and the evidence for direct precipitation of carbonate on the sea floor in the lower Buschmannsklippe Formation, compare well with distinctive carbonate units overlying Proterozoic glacial diamictite world wide (Williams, 1979; Aitken, 1991; Hoffmann, 1991; Grotzinger and Knoll, 1995; Kennedy, in press), and particularly with carbonate overlying the Blässkranz diamictite in the Naukluft Nappe Complex. Similar to these carbonate units (e.g. Fairchild and Hambrey, 1995, Kennedy, in press) the Bildah Member is interpreted to record transgression consistent with deposition during post-glacial sea-level rise.

The upper La Fraque Member and the Okambara Member shoal upward to mid-shelf facies with evidence for strong storm and wave influence (e.g. Kreisa, 1981), followed by restricted lagoonal/peritidal facies, with evidence for evaporation and desiccation. Significantly, the contact with the Weissberg Member corresponds to a gradational change in lithology, but there is no evidence for a corresponding change in water depth. Thus, although the Buschmannsklippe Formation is interpreted as a transgressive-regressive succession that forms a complete depositional sequence (Hegenberger, 1993), the conformable transition at the top of the sequence is interpreted as evidence for conformity between the Witvlei Group and the Kuibis Subgroup (Hegenberger, 1993; Hoffmann et al., 1995).

**Kuibis Subgroup.**—The Kuibis Subgroup in the Gobabis area consists of two units of sandstone, each interpreted as shallow-marine in origin and each overlain by carbonate-rich strata interpreted to have formed in a sub-tidal environment (Hegenberger, 1993). Together these units are interpreted as two depositional sequences, each with lowstand- to transgressive-sandstone at the base and transgressive- to highstand carbonate at the top.
*C*emostratigraphy: *Methods and Diagenetic Alteration.*

Carbonate samples from the Witvlei and Nama groups in the Gobabib area were collected from multiple exposures, and their corresponding isotope values are plotted only in approximate positions relative to generalized stratigraphic columns. Although carbonate sample from the Naukluft Nappe Complex are from a single locality, they are not directly tied to a measured section and their positions are approximate. The remainder of the data presented in this study, however, are directly tied to measured sections. The finest-grained most pristine-looking samples were collected at three to ten meter intervals. Carbonate samples from each of the four geographic areas were evaluated petrographically and geochemically for diagenetic alteration following the methods reviewed by Kaufman and Knoll (1995).

Samples consist generally of fine-grained limestone and dolostone, or more rarely of intraclast or ooid packstone/grainstone. Fluorescence under cathodoluminescence is variable, and highly luminescent areas were avoided as much as possible during microsampling. $\delta^{18}$O values range between -0.1 and -16.7‰; most lie between -5 and -10‰. Mn/Sr ratios range between 0.05 and 22.24; most are less than 2, and nearly all are less than 10. Fe/Sr ratios range between 1.3 and 24.3; most are less than 10. Following Narbonne et al. (1994) samples with $\delta^{18}$O > -10‰ and Mn/Sr < 2 are considered to be the least altered. According to these criteria, carbonate samples from the Gobabib area show the most evidence for alteration of all the studied sections. Nonetheless, in most cases these samples passed at least one of the diagenetic tests. In addition, $\delta^{13}$C values for samples that failed one of the tests still agree well with $\delta^{13}$C values from stratigraphically nearby samples that passed both, suggesting the alteration of $\delta^{13}$C is minimal. Smooth and consistent variation
of $\delta^{13}$C values in each of the areas reveals clear stratigraphic trends (Fig. 2). There is no evidence that the trends are related to lithology, as interbedded limestone and dolostone samples yielded equivalent $\delta^{13}$C values; nor is there any evidence that stratigraphic trends were controlled by sedimentary facies, as stratigraphic trends continue irrespective of facies changes. Thus, as a whole, carbonate samples from the Witvlei and Nama groups in each of the studied areas are interpreted to record near-primary variations in the carbon-isotope composition of seawater.

*Carbon-Isotope Variations: Gobabis Area.*

Carbon-isotope data from the Court Formation and the unconformably overlying Buschmannsklippe Formation and Kuibis Subgroup are plotted in approximate positions relative to the composite stratigraphic column in Fig. 2a. Dolostones from the Court Formation indicate a wide range in $\delta^{13}$C values (-4.3 to +5.8‰; Table 1). Significantly, the most negative $\delta^{13}$C values are from samples near the base of the formation and $^{13}$C abundances increase up section. Exact stratigraphic trends cannot be determined, however, because the positions of highly $^{13}$C-enriched samples is uncertain. It is possible that diagenetic alteration in a restricted or lacustrine basin can account for the isotopic variations of the Court Formation, but there is minimal independent geochemical or petrographic evidence to indicate such alteration. Given that the Court Formation lies directly above glacial tillite and consists of typical marine facies, and given the world-wide recognition of multiple Proterozoic glacial horizons, each associated with similar negative-to-positive $\delta^{13}$C excursion pairs (Kaufman et al. in press), it seems, on balance, that the carbon-isotope variations are most-likely primary, and that the Court Formation
records global-seawater conditions, including a post-glacial negative-to-positive carbon-isotope excursion.

δ\(^{13}\)C values from the Buschmannsklippe Formation decrease from -2.6‰ near the base of the formation to a minimum near -5‰ corresponding to the calcite fans and associated facies near the maximum flooding zone. Values then increase up section to near -3.5‰. The uppermost Kuibis sample is relatively enriched \(^{13}\)C (+0.2‰), however. The consistency of negative δ\(^{13}\)C values up into the lower Kuibis Subgroup supports physical stratigraphic evidence that the transition from the Buschmannsklippe Formation to the Kuibis Subgroup is conformable (Hegenberger, 1993). Consequently, the Buschmannsklippe Formation is interpreted as a conformable transition extending from post-glacial transgression up into the lower Nama Group in the Gobabis area.

**Naukluft Nappe Complex**

The Naukluft Nappe Complex consists of six major thrust nappes comprising Witvlei- and Nama-equivalent strata that were emplaced above autochthonous units of the Nama Group (Hartnady, 1978; Miller, 1983; Hoffmann, 1989; Hoffmann et al. 1995). Stratigraphic equivalents of the Witvlei Group include the Bläskranz and Tsabisis formations. The 200 m thick Bläskranz Formation includes a laminated-shale unit, with outsized limestone clasts interpreted as dropstones, and an overlying unit of massive diamictite (Hoffmann, 1989; Hoffmann et al., 1995). It is interpreted as glacial in origin and is of particular interest here because of suggested correlations between it and the younger of the two Varanger tillite horizons in Spitsbergen (Hoffmann, 1989; Hoffmann et al. 1995; Kaufman et al., in press). The stratigraphically overlying Tsabisis Formation consists of a basal few
meters of massive- to thinly laminated, fine-grained, pink/tan dolostone overlain by thinly-interbedded dolostone and purple shale. It bears a strong resemblance to, and has been correlated lithostratigraphically with, the lower Buschmannsklippe Formation of the Witvlei Group in the Gobabis area (Hoffmann, 1989; Hoffmann et al. 1995).

Carbon-isotope compositions of samples from the basal Tsabisis Formation are near -3.1‰. Higher up carbonate samples are slightly more 13C depleted, ranging between -5.7 and -4.9‰. Such negative δ13C values are typical of carbonate units immediately overlying Proterozoic glacial horizons (reviewed in Kaufman and Knoll, 1995). In addition, both the absolute values and the stratigraphic trend resemble carbon-isotope values through the lower Buschmannsklippe Formation. This similarity strengthens lithostratigraphic arguments for correlation of the Tsabisis and Buschmannsklippe formations and supports the interpretation that both were deposited as part of the same depositional sequence (W2) during sea-level rise following the end of Bläskranz glaciation (Hoffmann, 1991; Hofmann et al., 1995).

*Sequence Stratigraphy: Northern Nama Sub-Basin*

North of Osis the Kuibis Subgroup (Fig. 3) thickens from 0 m at Osis to a maximum of more than 500 m (Germs, 1972b; Germs, 1983; Figures 2c and 3). It consists of the Dabis Formation, which is a thin basal unit composed principally of coarse sandstone, and the Zaris Formation, which forms a northward-thickening carbonate platform. The overlying Schwarzrand Subgroup also thickens northward from Osis, forming a succession, nearly 800 m thick, of alternating shelf mudstone and deltaic sandstone (Germs, 1983).
**Dabis Formation.**—The Dabis Formation consists principally of trough cross-stratified and planar-laminated, pebbly feldspathic sandstone. Generally it is only a few meters thick and grades upward into mixed shale, sandstone and limestone of the Zaris Formation. Locally, however, (e.g. at Zaris Pass, Fig. 3d) it is more than 50 m thick and consists of two units of coarse-grained sandstone separated by an intervening carbonate-dominated unit. This unit consists largely of fenestral, microbial dolostone with thin layers of coarsely recrystallized grainstone and coarse sandstone.

As in the Gobabis area (Hegenberger, 1993) and the southern Nama sub-basin (Saylor et al., 1995), coarse sandstone of the Dabis Formation is interpreted as largely fluvial- to marginal-marine in origin (Germs, 1983). Locally-developed fenestral, microbial dolostone is interpreted to have formed in a quiet-water, near-shore environment that was episodically inundated by sand. A sequence boundary is interpreted to lie at the base of the upper thick sandstone unit, which is similar to other coarse-sandstone units in the Kuibis Subgroup, and which most likely formed by siliciclastic progradation during relative sea-level low-stand.

**Zaris Formation.**—The Zaris Formation consists, from bottom to top, of the Omkyk, Hoogland and Urikos members. The carbonate-dominated Omkyk and Hoogland members grade upward and northward into the shale-dominated Urikos Member (Germs, 1983). The Omkyk Member consists of three coarsening-upward successions (Germs, 1983), each 10's of meters thick with mudstone-rich facies at the base, coarser-grained, heterolithic facies with abundant hummocky cross-stratification and intraclast conglomerate in the middle, and cross-stratified grainstone commonly overlain by thrombolitic
and stromatolitic biohermal and biostromal dolostone at the top (Figs. 2c, 3, and 4). Thinly interbedded limestone, sandstone and shale beds of the overlying Hoogland Member fine upward. They grade into the green shale-dominated Urikos Member, which contains subordinate, thin beds (<2 cm) of limestone and sandstone. Thicker beds of coarse-grained carbonate with large-scale wave-rippled tops (e.g. coarse-grained ripples of Cheele and Leckie, 1992) and sandstone beds with hummocky cross-stratification are common in the upper 30 m of the Urikos Member. These facies grade upward into planar-laminated sandstone of the overlying Schwarzrand Subgroup.

Coarsening-upward successions of the Omkyk Member are interpreted as shallowing-upward parasequences (Germs, 1983). Each grades upward from storm-dominated carbonate ramp facies (e.g. Kreisa, 1981) to cross-bedded grainstone-shoal facies, which are overstepped by microbial biostromes (e.g. Read, 1985). The Hoogland Member is interpreted to record upward-deepening from a storm-influenced ramp environment with abundant hummocky cross-stratification to rhythmite and shale with no evidence for storm-wave influence. The Urikos Member records upward-shallowing from facies deposited below storm-wave base to storm-dominated mixed carbonate and siliciclastic facies (e.g. Cheele and Leckie, 1992), followed conformably by siliciclastic strata of the Schwarzrand Subgroup.

The Omkyk Member and immediately underlying, transgressive sandstone of the Dabis Formation together are interpreted as a complete depositional sequence. Fine-grained carbonate and shale of the lowest cycle constitute the transgressive systems tract. Upper cycles, which are thinner and dominated by coarser-grained, higher-energy shoal facies, constitute the highstand systems tract. The Hoogland and Urikos Members together form a transgressive-regressive succession, also interpreted as a complete
depositional sequence. The upper boundaries of both sequences are locally conformable transitions.

**Carbon-Isotope Variations: Northern Nama Sub-basin.**

Across the northern sub-basin negative δ¹³C values of samples from the Dabis Formation are followed by increasingly ¹³C-enriched samples from the Zaris Formation which reach a maximum between +4‰ and +6‰ (Fig. 3). This positive-isotope excursion is best represented by the Zebra River section (Figs. 2c and 4) where δ¹³C values increase through the Omkyk Member from near -1‰ at the base to a peak of +4.5‰, then decrease through the Hoogland Member to values straddling 0‰ in the overlying Urikos Member. Although samples from the Urikos Member are depleted in ¹⁸O, perhaps recording the influence of diagenetic fluids circulating through the surrounding shale, high Sr abundances and Mn/Sr values less than 2, similar to those from samples lower in the section, indicate the δ¹³C alteration is minimal. The timing of the positive-isotope excursion is well constrained by a U-Pb zircon age of 548.8 ± 1 Ma (Grotzinger et al., 1995a) for an ash bed near the base of the Hoogland Member in the Zebra River section (Figs. 2c and 4).

The negative-isotope excursion (N_pg interval), and the rising (P_r interval) and falling (P_f interval) limbs of the positive-isotope excursion constrain correlation of shallowing-upward parasequences and depositional sequences across the Kuibis Subgroup of the northern sub-basin, including exposures of the Kuibis Subgroup in the Gobabis area (Figure 3). Three depositional sequences are recognized. The lowest sequence (K1) is best developed in the region near Zaris Pass and in the Gobabis area. It is characterized by the negative δ¹³C values of the N_pg interval. The middle sequence (K2) consists of the three upward-shallowing parasequences of the
Omkyk Member and records the rising $\delta^{13}$C values of the $P_r$ interval. The highest sequence (K3) records the maximum $\delta^{13}$C values, which define the top of the $P_r$ interval, and also falling $\delta^{13}$C values of the $P_f$ interval. The $P_f$ interval ends in a minimum zone with $\delta^{13}$C values less than 0‰.

**Sequence stratigraphy: Southern Nama Sub-basin**

The southern Kuibis Subgroup thickens southwestward, toward the Gariep belt, expanding from 0 m at Osis to a maximum of 200 m near Swartkloofberg (Germs, 1983; Figs. 2d and 3). The southern Schwarzrand Subgroup also thickens southwestward, from 200 m at Osis to more than 1200 m near Swartkloofberg (Fig. 2d). Saylor et al. (1995) recognized six depositional sequences in the southern Kuibis and Schwarzrand subgroups, and traced the intervening sequence boundaries across much of the southern sub-basin.

**Kuibis Subgroup.**—The southern Kuibis Subgroup forms two depositional sequences (K1, K2), each of which is capped by a locally erosional sequence boundary (Saylor, 1993; Saylor et al., 1995) (Fig. 2d). Each sequence has a coarse sandstone base (Kanies and Kliphoek Members), comprising fluvial braidplain to shallow-marine facies and a carbonate-rich top (Mara and Mooifontein Members), composed principally of sub-tidal dolarenite and calcarenite (Germs, 1983; Saylor, 1995). Carbonate rocks of the lower sequence thin and grade into shale and sandstone eastward and northward (Germs, 1983). Carbonate at the top of the upper sequence is locally truncated by incised canyons with more than ten meters of relief (Chapter, 2; Germs, 1995).

**Schwarzrand Subgroup.**—The lower part of the Schwarzrand Subgroup, comprising the Nudaus Formation and the Nasep Member of the Urusis
Formation, consists principally of fine-grained siliciclastic mudstone and sandstone interpreted to have been deposited in a range of siliciclastic-shelf and deltaic environments (Fig. 2d; Germs, 1983; Saylor et al., 1995). It constitutes two depositional sequences (S1 and S2), both of which are bounded above by erosional surfaces that cut down section northeastward (Saylor et al., 1995).

The middle part of the Schwarzrand Subgroup constitutes carbonate platform rocks of the Urusis Formation (Huns, Feldschuhhorn and Spitskop members) and thickens southwestward to a maximum of nearly a kilometer near Swartkloofberg (Fig. 2d). Carbonate facies in the lower part of the platform (Huns Member) consist largely of meter-scale, coarsening-upward cycles, which consist of stromatolites, calcisiltite, and pellet/intraclast grainstone, and are commonly capped by karst-surfaces with up to 1.5 m of relief (Chapter 2). Pinnacle reefs that developed above a flooding surface at the top of the Huns Member are enveloped in shale of the Feldschuhhorn Member. Carbonate facies in the upper part of the platform (Spitskop Member) are generally finer grained than in the lower part. They consist largely of thin-bedded calcisiltite, with locally developed stromatolites and thrombolites, and are interpreted to have been deposited in low-energy subtidal settings (Chapters 2 & 3)

An unconformity with extensive erosional canyons caps the carbonate platform. Near the western margin of the platform the erosion surface cuts down to the level of the Huns Member. The total amount of relief, more than 500 m, is a combination of depositional thinning and erosional truncation across a shelf-to-basin transition telescoped by thrust sheets into a relatively small area (Chapter 3). Conglomerate and shallow marine sandstone and
shale of the Cambrian Nomtsas Formation partially infill the incised valleys.

*Carbon-isotope Variations: Southern Nama Sub-basin*

The Kuibis Subgroup in the region near Swartkloofberg (Figs. 2d and 3), where the succession reaches its thickest point in the southern sub-basin, records an excursion from negative $\delta^{13}$C values (minimum of -3.5‰) in the Mara Member and lower Kliphoek Member to positive $\delta^{13}$C values (maximum +2.8‰) in the Mooifontein Member. However, near Osis (Fig. 3), where the basal units of the Kuibis Subgroup pinch out, only high-positive values (+6‰) are documented.

Carbon-isotope profiles from the southern Kuibis generally resemble, and are correlated with the Npg and P$_r$ intervals of the northern Kuibis Subgroup. Consistent with the physical stratigraphy, the pinch-out of the Npg interval between Swartkloofberg and Osis (Fig. 3) indicates stratigraphic pinchout of sequence K1 and stratigraphic onlap of sequence K2 across the basement arch at Osis. Sequence K3 of the northern Kuibis Subgroup corresponds to the P$_f$ interval. It is unclear how it correlates across the Osis arch, however, because there is no evidence for the P$_f$ interval in the southern sub-basin, and no sequence boundaries have been traced across the arch. Sequence K3 may correlate either with the unconformity at the top of the Kuibis Subgroup or, alternatively, with sequence S1 of the basal Schwarzrand Subgroup south of Osis. Stratigraphic correlations suggested by Germs (1983), plus field data from this project show no evidence that limestone of sequence K3 grades southward into shale and sandstone of sequence S1. Consequently, sequence K3 is tentatively correlated with the unconformity at the top of the southern Kuibis Subgroup. No overlap is recognized between the northern Kuibis Subgroup and the southern
Schwarzrand Subgroup. Instead, the base of the southern Schwarzrand Subgroup is interpreted as younger than the top northern Kuibis Subgroup.

Carbon-isotope data from the carbonate platform rocks of the lower and middle Schwarzrand Subgroup lie in a narrow band of moderately positive, relatively invariant values (I interval; cf. Pelechaty et al., 1996) that decrease slightly upward from near +2%o at the base of the platform, to near +1%o just below the unconformity that contains the Proterozoic-Cambrian boundary (Fig. 2d). Ediacaran-type fossils, discovered in the Spitskop Member near the top of the platform, lie above a 543.3±1 Ma volcanic ash bed (Grotzinger et al., 1995). The fossils, the ash bed and the carbon-isotope data were all collected from the same measured section of the Spitskop Member.

DISCUSSION

Global Correlation of Namibian Chronostratigraphy

A composite reference section (Fig. 5a), constituting the thickest, most complete component sections, is constructed on the basis of the proposed carbon-isotope chemosтратigraphic and sequence stratigraphic correlations. This reference section includes a lower part, comprising the Witvlei Group (Court and Buschmannsklippe formations corresponding to sequences W1 and W2) and the basal Kuibis Subgroup (sequence K1) in the Gobabis area (Fig. 2a), plus the Bläskranz tillite from the Naukluft Nappe Complex (Fig. 2b), a middle part, corresponding to sequences K2 and K3 of the Kuibis Subgroup in the Zebra River section (Fig. 2c) of the northern Kuibis Subgroup, and an upper part consisting of the entire Schwarzrand Subgroup of the southern sub-basin in the region near Swartkloofberg (Fig. 2d). Two glacial horizons, with an intervening interval of variable negative to positive δ13C values lie at the base of the reference section. A negative δ13C excursion
(N$_{pg}$ interval) above the Bläskranz Formation is followed by the rising ($P_r$ interval) and falling ($P_f$ interval) limbs of an overlying large positive $\delta^{13}C$ excursion, which, in turn, is followed by an interval of moderately positive, relatively invariant isotope values (I interval) that extends up to the Proterozoic-Cambrian boundary unconformity. The most reliable $^{87}Sr/^{86}Sr$ values available for limestones in the succession range from 0.7081 near the base of the Buschmannsklippe Formation to 0.7084 in the Schwarzrand Subgroup.

The general form of the composite-isotope profile from Namibia resembles the preliminary isotopic data presented by Kaufman et al. (1991), and is consistent with previously suggested global correlations. As suggested by Kaufman et al. (in press) based on carbon- and strontium-isotopic correlations, tillites of the Blaubeker and Bläskranz formations may represent two intervals of Varanger glaciation, each followed by a similar negative-to-positive carbon-isotope excursion pair. Thus, the Blaubeker Formation may correlate with the older of the Varanger tillites in Spitsbergen and with the glaciogenic Ice Brook Formation of Western Canada (Kaufman et al. in press). The Bläskranz Formation would then correlate with the younger of the two Varanger tillites and, although unrepresented by tillites, with strata in the middle Sheepbed Formation of western Canada that lie between two positive carbon-isotope excursions (Kaufman et al., in press). Two positive carbon-isotope excursions are also recorded by terminal Proterozoic strata of Oman (Fig. 5d; Burns and Matter, 1993). Thus, extending the correlations to the Middle East, the Blaubeker Formation may be equivalent to tillite beneath the Abu Mahara Formation in Oman; the base of the Buschmannsklippe Formation may correlate with the Khufai-Shurman
boundary, where a sharp excursion to negative $\delta^{13}C$ values defines the base of the $N_{pg}$ interval.

Chemostratigraphic intervals in the Ediacaran-fossil bearing, post-Bläskranz part of the Namibian reference section compare well with several other post-Varanger, Ediacaran-fossil bearing terminal Proterozoic sections including sections in arctic Canada, Oman, and Siberia (Burns and Matter, 1993; Narbonne et al., 1994; Kaufman and Knoll, 1995; Jenkins, 1995; Knoll et al., 1995; Pelechaty et al., 1996; Kaufman et al., in press). Tie-points at the boundaries of chemostratigraphic intervals shown in Fig. 5 indicate preferred correlations among these sections. The $N_{pg}$ interval is clearly developed in Oman, but its exact stratigraphic position must be inferred in arctic Canada. Strata representative of this interval do not appear to be preserved in Siberia. The $P_t$, $P_f$ and $I$ intervals are clearly developed in arctic Canada, Oman and Siberia. Even the excursion to slightly negative $\delta^{13}C$ values that defines the top of the $P_f$ interval in Namibia appears to have correlatives in Oman, and possibly in Siberia, strengthening the argument that it reflects secular, global seawater isotopic excursions. In Siberia and Canada (Narbonne et al., 1994; Knoll et al., 1995, Pelechaty et al., 1996), however, there is also a sharp negative carbon-isotope excursion ($N_{sub C}$ interval) just above the $I$ interval and just below the Proterozoic-Cambrian boundary which does not appear in the Namibian reference section. This excursion is inferred to be contained within the Proterozoic-Cambrian boundary unconformity (Grotzinger et al. 1995). Radiometric dates in Namibia and Siberia constrain the duration of this hiatus to less than one million years (Bowring et al., 1992; Grotzinger et al., 1995).

The Namibian reference section, from the base of the Buschmannsklippe Formation up to the top of the Spitskop Member, is
interpreted as a relatively complete and continuous record of carbon-isotope variations, extending from the transgression at the end of the Varanger epoch to within one million years of the Proterozoic-Cambrian boundary. The general form of the isotope record resembles Knoll and Walter's (1992) summary of post-Varanger terminal Proterozoic carbon-isotope variability. However, U-Pb zircon ages from Namibia (Grotzinger et al., 1995a) revise the temporal calibration of this generalized isotope curve. Specifically, the I interval and overlying $N_{subC}$ interval, originally perceived as encompassing more than two thirds of terminal Proterozoic time (Knoll and Walter, 1992), extending from $\sim$580 to $\sim$545 Ma, are relegated to only the last 5 or 6 million years (between 548.8 $\pm$ 1 Ma and 543.3 $\pm$ 1 Ma). Unfortunately, the dated ash beds in Namibia span only from the decline of the youngest positive carbon-isotope excursion ($P_f$ interval) through the Proterozoic-Cambrian boundary. The older part of terminal Proterozoic time is still relatively poorly calibrated. Given the excellent radiometric and chronostratigraphic control in Namibia it is worthwhile to ask how the older part of the record may be affected due to compression of the younger part.

**Extrapolation of Age Constraints**

Graphic correlation (Shaw, 1960; Shaw, 1964; Mann and Lane, 1995) was used to assess variations in average sediment accumulation rate through the Namibian reference section. Correlation diagrams were constructed by plotting the positions of stratigraphic tie points in the Namibian reference section against the corresponding positions in stratigraphically equivalent comparison sections. Most commonly the first and last occurrences of fossils are used to define stratigraphic tie-points, but in this case, the boundaries of the chemostratigraphic intervals defined the tie-points (Fig. 5). Where a
boundary is not clearly developed in a section because of limited carbon-isotope data, its position is inferred. A slope of one for a line connecting the tie points indicates that sediment accumulation rates in the two sections were equal. Differences in sediment accumulation history related to variations in sediment accumulation rate, including the effects of unconformities, change the slope of the line.

Fig. 6 graphically correlates the terminal Proterozoic reference section from Namibia with time-equivalent sections in Canada and Oman, the two sections that have the most continuous carbon-isotopic trends. Almost all slopes are less than one, indicating that sediment accumulation rates throughout each of these terminal Proterozoic sections was less than in the Namibian reference section. In addition, there is a sharp bend at the boundary between the \( P_r \) and \( P_I \) intervals; steeper slopes are associated more with the \( N_{pg} \) and \( P_r \) intervals than with the \( P_I \) and I intervals, indicating a relative increase of sediment accumulation rate through the \( P_I \) and I intervals in the Namibian reference section compared to Oman and Canada. This increase is relative; in absolute terms, average sediment accumulation rates could have slowed in the other successions while remaining constant in Namibia.

The lower part of the Namibian reference section is interpreted to have been deposited on a passive margin and the upper part in a foreland basin. Since foreland basin subsidence is expected to increase over time (Allen and Homewood, 1986), the most likely effect was that of an absolute increase in the subsidence and hence sediment accumulation rate in Namibia. Thus, the average sediment accumulation rate calculated for the I interval in Namibia represents a maximum rate of accumulation and can be extrapolated downward to estimate minimum constraints on older tie points along the isotope curve. Using 3.5 m.y. as the minimum possible duration for the I
interval (U-Pb age constraints span between 548.8 ± 1 and 543.3 ± 1 Ma) and 1400 m as the thickness of the I interval, a maximum average sediment-accumulation rate of 400 m/m.y is calculated for the part of the section corresponding to the I interval in Namibia. Extrapolating downward yields minimum - estimated ages of 549 Ma for the base of the Pr interval and 551 Ma for the base of the Npg interval (base of Buschmannsklippe Formation) and the end of the Naukluft glaciation.

If the alternative interpretation is accepted, i.e. that sediment accumulation rates decreased across the Pr to Pf boundary in Oman, then an average sediment accumulation rate calculated for the I interval in Oman is a minimum for the succession and extrapolation downward yields a maximum age estimate for the base of the Npg interval. Choosing the maximum possible duration for the I interval of 7.3 m.y. (maximum possible span between 548.8 ± 1 and 543.3±1), the average sediment accumulation rate during deposition of the 275 m thick I interval in Oman yields a minimum sediment accumulation rate 38 m/m.y. Extrapolating downward, a maximum age of ~553 Ma is estimated for the base of the Pr interval and ~564 Ma for the base of the Npg interval in Oman, and, by correlation, for the base of the Buschmannsklippe Formation and the end of the Naukluft glaciation.

The maximum age for the end of the Bläskranz glaciation in Namibia is approximately the same as a volcanic horizon (U-Pb zircon age of 565±3 Ma) interbedded with Ediacaran-type fossils that lie several hundred meters above Avalon glacial rocks in Newfoundland (Benus, 1988; Smith and Hiscott, 1984; Myrow, 1995). Albeit only rough estimates, extrapolated ages for tie points along the isotope curve in Namibia illustrate the point that the end of glaciation in Namibia is significantly younger than the 600 to 590 Ma age range assigned to the Avalon tillites and commonly considered to represent
the span of the Varanger glaciation (Knoll and Walter, 1992). The younger age estimated for the Blässkranz Formation, which is correlated with the younger of the Varanger tillites, places an upper limit on additional isotope excursions recently documented from Australia, Canada, Spitsbergen, and Oman to fit between the two Varanger-equivalent glacial horizons (Fig. 7; Kaufman et al., in press). Thus, the Avalon tillites probably correlate with the older of the Varanger tillites. Diverse Ediacaran-type fossils overlying the Avalon tillite in Newfoundland may be close in time, and possibly older than, the younger of the two Varanger glaciations. This graphical exercise suggests that the Varanger glacial epoch spanned 20 to 30 m.y., beginning less than 600 Ma and ending some time after 570 Ma.

Implications

The new age constraints (Grotzinger et al. 1995) and the recognition of multiple glacial episodes and associated carbon- and strontium-isotopic excursions (Kaufman et al. in press) expand the interval of Varanger glaciation and shrink the post-Varanger terminal Proterozoic epoch (Fig. 7). Rather than a single period of intense and wide-spread glaciation, which may have created an Earth completely covered in snow (Harland, 1964; Kirschvink, 1982), the Varanger glacial epoch may be interpreted, instead, as multiple distinct glaciations, possibly of more local distribution, but recorded globally as changes in seawater chemistry. The distinctive facies and relative depletion in $^{13}$C that characterize carbonate rocks overlying Proterozoic glacial tillites are not singular anomalies but, instead, are repeated features related to the dynamics of ocean circulation associated with periods of ice expansion and retreat (Kaufman et al., in press; Knoll et al., 1996; Grotzinger and Knoll, 1995; Kaufman et al., 1991; Tucker, 1992).
A possible corollary of this interpretation is that younger negative-to-positive carbon-isotope excursions, at the top of the Kuibis Subgroup in Namibia, and at the Proterozoic-Cambrian boundary in arctic Canada and Siberia, may also be related to glacial events (Kaufman et al., 1991), but perhaps without preservation of extensive tills or other glaciogenic facies. Features of possible glacial origin, including iron-formation, diamictite, and groove marks, have been reported from the Kuibis-Schwarzrand boundary in Namibia (Schwellnus, 1942, Germs, 1995), close to the muted negative-isotope excursion at the top of the Kuibis Subgroup. In addition, stacked karst-capped shallowing-upward cycles in carbonate platform rocks of the middle Schwarzrand Subgroup (Chapter 2) resemble carbonate cycles formed during other major periods of glaciation in Earth history (Read 1996). Combined with the impressive erosional canyons at the Proterozoic-Cambrian boundary, they indicate high-frequency and high-amplitude relative sea-level oscillations during Schwarzrand deposition. Although tectonic forces may have driven these relative sea-level changes in the tectonically active Nama foreland basin, the unconformity recognized at the Proterozoic-Cambrian boundary is present world-wide (Runnegar et al., 1995), suggesting a possible eustatic origin. Continued stratigraphic analyses of terminal Proterozoic carbonate platforms in stable tectonic settings may yield additional evidence for glacioeustatic sea-level oscillations and global ice-house conditions during the latter, post-Varanger stages of terminal Proterozoic time.

Sedimentologic and isotopic evidence for glaciations near the Proterozoic-Cambrian boundary have been regarded with skepticism (Chapter 2), in part because they were thought to be tens of millions of years younger than the Varanger glacial epoch. Within the framework of the time-scale proposed here, however, they post-date the younger of the two Varanger
glaciations by only a few million years and may be regarded as the last pulses of an extended glacial epoch, which featured repeated episodes of ice expansion and retreat.

The recognition of multiple terminal Proterozoic glacial episodes, some of which may have occurred near the Proterozoic-Cambrian boundary, has significant implications for understanding the relationship between the Ediacaran-radiation and glaciation. Although no diverse Ediacaran fossils are known to underlie terminal Proterozoic glacial horizons, simple Ediacaran-like forms from arctic Canada underlie rocks thought to represent the older of the Varanger glaciations (Hofmann et al., 1990; Kaufman et al., in press.). Chemostratigraphic correlations of fossil horizons in Canada and Australia indicate that the range of diverse Ediacaran-type fossils extends down near to, or below, the younger of the Varanger tillites. In addition, the age of diverse Ediacaran fossils in Newfoundland overlaps with the estimated age of the Younger Varanger glaciation. These relationships suggest that the radiation of Ediacaran organisms may have begun during the Varanger glacial epoch. Inferred glacial horizons within post-Varanger Ediacaran-fossil bearing terminal Proterozoic strata indicate episodic glaciation may have continued throughout much of the period of Ediacaran radiation.

CONCLUSIONS

1) Terminal Proterozoic strata of the Witvlei and groups and their correlatives preserve relatively unaltered records of carbon- and strontium isotopic excursions in seawater. Thus, combining chemostratigraphy and sequence stratigraphy, it is possible to correlate widely separated section across the Kalahari craton of Namibia, and hence construct a composite reference section.
2) Tillite of the Blaubeker Formation at the base of the Witvlei group and tillite of the Bläskranz Formation, interpreted to correlate with an unconformity in the middle of the Witvlei Group, mark two distinct glacial horizons which likely represent two distinct phases of Varanger-age glaciation similar to what is recorded in the north Atlantic region.

3) Carbon-isotope excursions recorded by the Buschmannsklippe Formation and the Kuibis and Schwarzrand subgroups, interpreted to lie stratigraphically above the Bläskranz Formation, are divided into four chemostratigraphic intervals, a post-glacial negative isotope excursion (N_g interval), the rising (P_r interval) and falling (P_f interval) limbs of a positive isotope excursion, and an interval of moderately positive, relatively invariant values (I interval) that extends to the Proterozoic-Cambrian boundary unconformity. These chemostratigraphic intervals are recognized in isotope profiles from other terminal Neoproterozoic successions around the world.

4) Carbon-isotope data from Namibia are tied directly to U-Pb radiometric age constraints from intercalated volcanic ash beds. These data constrain the duration of the I interval to less than 6 m.y. Extrapolating sediment accumulation rates calculated for the I interval in Namibia and Oman, a minimum age of ~552 Ma and a maximum age of ~565 Ma are estimated for the end of the younger Varanger glaciation.

ACKNOWLEDGEMENTS

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        | -1.1 | N.D. | N.D. | N.D. | N.D. | N.D. |
P440A-B.93 Tsabise | -3.1 | N.D. | N.D. | N.D. | N.D. | N.D. |
        | -3.1 | N.D. | N.D. | N.D. | N.D. | N.D. |
P438A-J.93 Tsabise | -5.5 | N.D. | N.D. | N.D. | N.D. | N.D. |
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        | -5.4 | N.D. | N.D. | N.D. | N.D. | N.D. |
        | -4.9 | N.D. | N.D. | N.D. | N.D. | N.D. |

Northern Nama

Sub-basin
Zebra River
H1 Omhylw | 4 | -1.3 | -12.5 | 3.38 | 13.44 | 0.020 |
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H3        | 23 | 0.6  | -13.2 | 0.14 | 1.3   | 0.009 |
H4        | 35 | 0.6  | -10.6 | 0.77 | 3.3   | 0.010 |
H5        | 58 | 1.2  | -7.7  | 0.15 | 1.3   | 0.009 |
H6        | 87 | 0.7  | -9.5  | 0.49 | 1.8   | 0.005 |
H7c       | 99 | 1.4  | -9.0  | 1.87 | 19.5  | 0.012 |
H7m       | 99 | 1.6  | -8.6  | 0.73 | 2.1   | 0.006 |
H8        | 102 | 2.5  | -8.8  | 0.13 | 0.9   | 0.005 |
H9        | 114 | 1.7  | -8.5  | 0.23 | 9.1   | 0.008 |
H10       | 121 | 3.1  | -8.6  | 0.20 | 2.3   | 0.015 |
H11       | 132 | 3.2  | -8.8  | 0.20 | 2.2   | 0.009 |
H12       | 155 | 3.5  | -9.9  | 0.25 | 2.7   | 0.008 |
H14       | 185 | 4.1  | -7.7  | 0.05 | 1.8   | 0.009 |
H15       | 197 | 4.5  | -8.4  | 0.06 | 7.1   | 0.115 |
H16       | 241 | 3.9  | -8.4  | 0.51 | 6.1   | 0.018 |
H17       | 262 | 3.1  | -9.8  | 0.48 | 3.8   | 0.021 |
H18       | 278 | 3.3  | -9.9  | 0.24 | 2.2   | 0.008 |
H19       | 279 | 3.6  | -8.7  | 0.09 | 1.7   | 0.008 |
H20       | 289 | 2.5  | -8.8  | 0.24 | 2.4   | 0.008 |
H21       | 313 | 2.6  | -7.2  | 0.13 | 2.2   | 0.008 |
H22       | 318 | 2.4  | -7.3  | 0.15 | 3.3   | 0.009 |
H23       | 320 | 2.7  | -8.5  | 0.08 | 1.8   | 0.010 |
N1 Uitkops | 321 | 1.6  | -9.3  | 0.05 | 2.1   | 0.009 |
N2        | 333 | 2.2  | -8.1  | 0.17 | 2.1   | 0.014 |
N3        | 354 | 1.6  | -9.4  | 0.31 | 4.2   | 0.007 |
U21       | 269 | 1.3  | -12.5 | n.a. | n.a.  | n.a. |
U22       | 360 | 1.2  | -12.5 | n.a. | n.a.  | n.a. |
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| | C141-13 | * | 45 | 0.2 | -6.6 | N.D. | N.D. | N.D. |
| | C151-33 | Ominky | 56 | 3.3 | -8.3 | N.D. | N.D. | N.D. |
| | C151-44 | 57 | 3.2 | -12.2 | N.D. | N.D. | N.D. |
| | C151-48 | 78 | 3.3 | -12.2 | N.D. | N.D. | N.D. |
| | C151-82 | 80 | 2.5 | -11.7 | N.D. | N.D. | N.D. |
| | C151-86 | 108 | 4.0 | -9.2 | N.D. | N.D. | N.D. |

<p>| Southern Name | | | | | | | | |
| Sub-basin: | Swartdooiberg area | | | | | | | |
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| | bsz34-3 | * | 61 | -2.0 | -5.3 | 3.34 | 10.25 | 0.423 |
| | bsz34-4 | * | 62 | -2.9 | -7.3 | 8.12 | 24.34 | 0.428 |
| | bsz34-5 | * | 63 | -3.4 | -7.8 | 2.74 | 14.78 | 0.433 |
| | bsz34-6 | * | 66 | -1.0 | -8.2 | 3.27 | 14.38 | 0.449 |
| | bsz34-8 | * | 75 | -3.1 | -3.6 | 3.54 | 20.23 | 0.350 |
| | bsz34-9 | * | 76 | -2.1 | -7.0 | 1.54 | 6.33 | 0.429 |
| | bsz34-12 | * | 101 | -2.6 | -6.6 | 4.00 | 6.51 | 0.401 |
| | bsz34-13 | * | 106 | -1.6 | -12.0 | 0.51 | 7.17 | 0.133 |
| | bsz33-2 | Koelfontein | 178 | 0.1 | -15.8 | N.D. | N.D. | N.D. |
| | bsz33-3 | * | 177 | 2.5 | -14.0 | N.D. | N.D. | N.D. |
| | bsz33-4 | * | 178 | 1.2 | -14.0 | N.D. | N.D. | N.D. |
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| | bsz33-9 | * | 187 | 1.6 | -13.7 | N.D. | N.D. | N.D. |
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| | bsz33-12 | * | 188 | 2.6 | -12.4 | N.D. | N.D. | N.D. |
| | bsz33-14 | * | 194 | 1.7 | -12.3 | N.D. | N.D. | N.D. |
| | FV1-0 | Spitskop | 8 | 1.9 | -5.0 | 0.01 | 0.81 | 0.001 |
| | FV1 | 20 | 1.8 | -7.3 | 0.00 | 1.18 | 0.001 |
| | FV1-2 | 32 | 2.3 | -7.5 | 0.00 | 0.76 | 0.001 |
| | FV1-3 | 54 | 2.1 | -6.5 | 0.00 | 0.39 | 0.001 |
| | FV1-4 | 66 | 1.9 | -7.4 | 0.01 | 1.67 | 0.001 |
| | FV1-5-2 | 78 | 1.9 | -6.8 | 0.02 | 1.64 | 0.002 |
| | FV1-5-3 | 78 | 1.8 | -6.2 | 0.02 | 1.64 | 0.002 |
| | FV1-6 | 80 | 1.9 | -7.7 | 0.01 | 2.24 | 0.002 |
| | FV1-7-1 | 92 | 1.3 | -8.0 | 0.02 | 1.55 | 0.003 |
| | FV1-7-2 | 92 | 1.7 | -8.5 | 0.03 | 1.49 | 0.003 |
| | FV1-8 | 110 | 1.6 | -7.3 | 0.02 | 0.83 | 0.005 |
| | FV1-9 | 122 | 2.1 | -6.0 | 0.08 | 0.63 | 0.005 |
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FIGURES

Figure 1: Map showing exposures of the Nama and Witvlei groups in Namibia and location of measured sections in Fig. 2.

Figure 2: Stratigraphic columns and corresponding carbon-isotope data through the Witvlei Group and Kuibis and Schwarzrand subgroups. Carbon-isotope data from the Gobabis area (a) and the Naukluft Nappe Complex (b) are plotted in approximate positions relative to the generalized columns. Except for the data from Kaufman et al., (1991), for which the sample locations are approximate, carbon-isotope data from the Zebra River section of the northern Nama sub-basin and from the composite stratigraphic column from near Swartkloofberg in the southern Nama sub-basin are tied to measured sections. See Fig. 1 for locations. (Note: Position of Naukluft Nappe Comlex is not palispatically restored)

Figure 3: Measured stratigraphic sections and carbon-isotope data along a north-south transect of the Buschmannsklippe Formation and the Kuibis Subgroup showing chemostratigraphic and sequence stratigraphic correlations.

Figure 4: Outcrop photograph of the Zebra River Section of the Zaris Formation in the northern Nama sub-basin showing position of parasequence boundaries, volcanic ash bed (vvv), and superimposed generalized carbon-isotope profile.

Figure 5: Stratigraphic columns (modified from Kaufman and Knoll, 1995) showing carbon-isotope excursions, fossil data, the most reliable limestone Sr-isotope values, and glacial horizons in a) the Namibian reference section and other correlative sections. All sections drawn to the same scale as the Namibian reference section. Boxed letters are chemostratigraphically-defined points of correlation. Sources of data: a) this paper; b) Narbonne et al. (1994); Kaufman et al. (1996); c) Knoll et al. (1986), Fairchild and Spiro (1987), Kaufman and others (1993), Kaufman et al. (in press) d) Burns and Matter (1993); g) Knoll et al. (1995). Wi=Wisonbreen Formation.

Figure 6) Graphic correlation of Namibian reference section with correlative comparison sections in Oman and arctic Canada indicates that average sediment accumulation rate in Namibia increased through the P1 and I intervals relative to the comparison sections. Downward extrapolation sediment accumulation rates calculated for the I interval in Namibia and Oman yield maximum and minimum estimates for ages of stratigraphically lower chemostratigraphic tie-points.

Figure 7) Suggested revised terminal Proterozoic chronostratigraphy calibrated with respect to available U-Pb zircon age constraints.
Fig. 7

T: Twitya disks
E: Ediacaran-type fossils
C: Cloudina and goblet-shaped fossils
S: Anabarites and Cambrotubulous

- U/Pb age which is tied to chemostratigraphic framework
- U/Pb age which is not tied to chemostratigraphic framework
APPENDIX A

Reference section for the southern Kuibis Subgroup
Location of Kulbis Reference Section
27°15' S, 16° 46'E
Kliphoek Farm (72)
West side of road #727
2.5 km S of Nootgedacht farmhouse
APPENDIX B
Mesured sections in the Lower Schwarzrand Subgroup
Nudaus Formation (sequence S1) and Nasep Member (Sequence S2)

Part 1: Photographs of typical facies
Part 2: Locations of measured sections
Part 3: Generalized measured sections
Part 4: Selected measured sections
PART 1:
Photographs of Facies in the
Lower Schwarzrand Subgroup
(Nudaus Formation and Nasep Member)
Section 93 on Anusi Farm. S1-S2 sequence boundary indicated.

S1-S2 sequence boundary. Fred Vendenbergh for scale.
Ripple cross-lamination (large-scale)

Ripple cross-lamination (small-scale)
Tabular-bedded sandstone

Planar-laminated sandstone

Cross-bedded and slumped sandstone
PART 2:
Locations of Measured Sections in the
Lower Schwarzrand Subgroup
(Nudaus Formation and Nasep Member)
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<th>Longitude E</th>
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<td>16° 51'</td>
<td>Geelperdshoek (76)</td>
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<tr>
<td>12</td>
<td>27° 25'</td>
<td>16° 47'</td>
<td>Tierkloof (75)</td>
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<td>52</td>
<td>27° 32'</td>
<td>16° 36'</td>
<td>NorthWitputs (22)</td>
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<td>62</td>
<td>27° 35'</td>
<td>16° 35'</td>
<td>Bobbejaankrans (180)</td>
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<td>75</td>
<td>26° 25'</td>
<td>17° 20'</td>
<td>Brackwasser (144)</td>
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<td>27° 10'</td>
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<td>Geigoab (95)</td>
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<td>17° 21'</td>
<td>Nuichas (94)</td>
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<td>85</td>
<td>27° 22'</td>
<td>17° 18'</td>
<td>Huns (106)</td>
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<td>86</td>
<td>27° 30'</td>
<td>17° 04'</td>
<td>Abos (80)</td>
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<tr>
<td>87</td>
<td>27° 29'</td>
<td>17° 12'</td>
<td>Sekelberg Mtn.</td>
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<td>88</td>
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<td>Kolke (84)</td>
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<td>27° 16'</td>
<td>16° 38'</td>
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PART 3:
Generalized Measured Sections in the Lower Schwarzrand Subgroup
(Nudaus Formation and Nasep Member)
Dominant Facies

- Carbonate grainstone
- Pebble conglomerate
- Cross-bedded sandstone
- Massive-weathering sandstone
- Planar-laminated sandstone; some cross-bedding
- Ripple-laminated siltstone with distinctive large spacing
- Ripple-laminated sandstone and siltstone
- Tabular-bedded sandstone and mudstone
- Mudstone

---

Volcanic ash bed
Sequence boundary
PART 4:
Selected Measured Sections in the
Lower Schwarzrand Subgroup
(Nudaus Formation and Nasep Member)
**PRINCIPLE BED OR LAMINAE CHARACTERISTICS**

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<th>Description</th>
<th>Image</th>
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<td>thickly and evenly parallel-laminated</td>
<td><img src="thick-bedded.jpg" alt="Image" /></td>
</tr>
<tr>
<td>low-angle cross-laminated</td>
<td><img src="low-angle.jpg" alt="Image" /></td>
</tr>
<tr>
<td>thinly and evenly parallel-laminated</td>
<td><img src="thinly.jpg" alt="Image" /></td>
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<tr>
<td>thin-bedded, mud-and sand-rich layers</td>
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<td>nodular, irregularly bedded</td>
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<td>medium-bedded</td>
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<td>volcanic ash bed</td>
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<td>thick-bedded; massive-weathering, vaguely planar-laminated</td>
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<td>large-scale trough cross-bedded</td>
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<td>mudstone with thin sand-sized interbeds</td>
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<td>elongate stromatolites</td>
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<td>columnar/pedestal stromatolites/thrombolites</td>
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**COMPOSITION**

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<td>mixed siliciclastic and carbonate sandstone</td>
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**GRAINSIZE**

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<tr>
<td>v very fine</td>
<td>w/l laminite/calcisiltite</td>
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<tr>
<td>f fine</td>
<td>a arenite/fine grainstone</td>
</tr>
<tr>
<td>m medium</td>
<td>r rudstone</td>
</tr>
<tr>
<td>c coarse</td>
<td>b breccia</td>
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SEDIMENTARY STRUCTURES

- planar-lamination
- low-angle cross-lamination
- tabular cross-stratification
- trough cross-stratification
- hummocky cross-stratification
- 3-D wave-ripple
- ripple cross-stratification
- wave ripples
- current ripples
- slump structures
- water-escape structures
- flute casts
- gutter casts
- mud cracks

GRAIN-TYPES

- pellets of rounded mud intraclasts
- pellets of rounded mud intraclasts
- coated grains; ooids
  - grapestone
- tabular intraclasts
- sub-mm platy-grains
  - skeletal grains
  - mud chips
  - form-fitted intraclast breccia
  - microbreccia:
    - form-fitted breccia with mm-scale intraclasts
    - quartz-pebble conglomerate

PALEOCURRENT INDICATORS

- unidirectional current indicator for e.g. flute mark (FM), current ripple (CR), and trough azimuth (TA)
- bidirectional current indicator for e.g. gutter cast (GC), wave-ripple trend (RT), stromatolite elongation (S), and parting or current lineation (PL, CL)
SECTION 62
Bobbejaankrans Farm (180)
Mooifontein Member up into Huns Member
BOBBEJAANKRANS (62: 0-25 m)

25 cm beds

5-25 cm thick beds; irregular bases and tops; irregular internal laminations; intercalated coarser and finer grains

m | a | r | b carbonate
m v f m c sandstone
BOBBEJAANKRANS 180 (62: 25-50 m)

25 meters

m l a r b carbonate
m v f m c sandstone
BOBBEJAANKRANS 180 (62: 50-75 m)

50 meters

m a r b carbonate
m v f m c sandstone
BOBBEJAANKRANS 180 (62: 100-125 m)

Nasep Mbr. / S2
Nudaus Fm. / S1

100 meters

m l a r b carbonate
m v f m c sandstone
BOBBEJAANKRANS 180 (62: 125-150 m)

150
145
140
135
130

125 meters

m l a r b carbonate
m v f m c sandstone

Huns Mbr./S3
λ: 40 cm; h: 15 cm
Nasep Mbr. /S2
BOBBEJAANKRANS 180 (62: 200-225 m)

200 meters

m | a | r | b | carbonate

m | v | f | m | c | sandstone
BOBBEJAANSKRANS 180 (62: 225 m - end)

225 meters

m l a r b carbonate

m v f m c sandstone

limestone is red near top
SECTION 75
Brackwasser Farm (144)
BRACKWASSER 144 (75: 0-25 m)

- 2.5 cm spacing

- m a r b carbonate

- m s f m c sandstone

- Levels and stratigraphic units

- Measure points and markers

- Section and profile view
BRACKWASSER 144 (75: 50-75 m)

- 10 - 30 cm spacing

PL 200

U 190

10 cm spacing

Naudus Fm.

S1

S2

50 meters

m l a r b carbonate

m s f m c sandstone
SECTION 82
Geigoab Farm (95)
GEIGOAB 95 (82: 0-25 m)

0 meters

m v f m c sandstone

m l a r b carbonate

25
20
15
10
5

meters
GEIGOAB 95 (82: 25-50 m)

Nunaus Fm.

25 meters

m l a r b carbonate
m v f m c sandstone
GEIGOAB 95 (82: 50-75 m)

- S3
- S2
- quartz pebbles
- MP
- λ: 30 cm
- wavy, erosional base
- mlarb carbonate
- mvfmc sandstone
SECTION 84
Kaalberg Farm (94)
KAALBERG 93 (84: 25-50 m)

\[ \lambda : 20 \, \text{cm} \]

\[ \lambda : 40 \, \text{cm} \]

irregular, scoured bases

m, l, a, r, b carbonate
m, v, f, m, c sandstone
KAALBERG 93 (84: 75-100 m)

Nudaus Fm.

75 meters

mlarb carbonate

mvmc sandstone
KAALBERG 93 (84: 125 m - end)
SECTION 85
Huns Farm (106)
HUNS 106 (85: 50-75 m)

50 meters

Nudlaus Fm.

m l a r b carbonate
m f m c c sandstone
SECTION 86
Abos Farm (86)
ABOS 80 (86: 75-100 m)

Nagep Mbr.

Urusis Fm.

Nadaus Fm.

75 meters

m f m c c sandstone

m l a r b carbonate

S1

S2

MP

100

95

90

85

80
ABOS 80 (86: 100-125 m)

Nasep Mbr.

>30 cm spacing

>15 cm spacing

m i a r b carbonate
m f m c c sandstone

100 meters
ABOL 80 (86: 150-175 m)

175
170
165
160
155
150 meters

m I a r b carbonate
m f m c c sandstone

Nasep Mbr.

210 PL
260 PL
SECTION 87
Sekelberg Mtn.
on border between
Moedhou (182) and Quaggaspoort (79)
SEKELBERG (87: 25-50 m)

Nudaus Fm.

rt 170

λ: 5 cm

m l a r b carbonate

m v f m c sandstone
SEKELBERG (87: 100-125 m)

125
120
115
110
105

Nasep Mbr.

100 meters

m a r b carbonate
m v f m c sandstone

λ: 25 cm
SEKELBERG (87: 175-200 m)

Huns Mbr.

175 meters

m | a | r | b | carbonate
m | v | f | m | c | sandstone
SEKELBERG (87: 200 m to end)
SECTION 92
Kliphoek Farm (72)
Kliphoek 72 SW (92: 0-25 m)

Nudaus Fm.

- 25
- 20
- 15
- 10
- 5
- 0

meters

0 1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20 21 22 23 24 25

m a r b carbonate

m f m c c sandstone

15 cm
Kliphoek 72 (92: 50-75 m)

Nudaus Fm.

50 meters

m | a | r | b carbonate
m f m c c sandstone

2 cm
245

15 cm
KLIPHOEK 72 SW (92: 75-100 m)

75 meters

Nudus Fm.

m f m c c sandstone
m i a r b carbonate

2 cm
15-20 cm
4 cm 145
13 cm
KLIPHOEK 72 SW (175-200 m)

Nasep Mbr.

175 meters

200
195
190
185
180

m l a r b carbonate
m f m c c sandstone
KLIPHOEK 72 SW (200-225 m)

Nasep Mbr.

200 meters

mlarb carbonate
mfccc sandstone
KLIPHOEK 72 SW (225-250 m)

225 meters

Nasep Mbr.

m a r b carbonate

m f m c c sandstone
SECTION 93
Anusi Farm (73)
ANUSI (93: 0-25 m)

Nudaus Fm.

25 meters

m, l, a, r, b, carbonate
m, f, m, c, c, sandstone
ANUSI (93: 25-50 m)

Nudaus Fm.

25 meters

m f m c c sandstone
m t a r b carbonate

2 cm spacing

EVB
ANUSI (93: 50-75 m)

50 meters

m a r b carbonate
m f m c c sandstone

irregular base

PL 200 top, 2 cm
ANUSI (93: 75-100 m)

Nudaes Fm.

75 meters

m l a r b carbonate
m f m c c sandstone

top, 2 cm
ANUSI (93: 100-125 m)

125

120

115

110

105

100 meters

Naves Fm.

m l a r b carbonate

m f m c c sandstone

faint parallel laminations

220, foreset dip direction

top of cliff
ANUSI (93: 125-150 m)

Nudaus Fm.

150
145
140
135
130

125 meters

m l a r b carbonate
m f m c c sandstone
ANUSI (93: 150-175 m)

175
170
165
160
155

150 meters

m f m c c sandstone
m l a r b carbonate

Nudaqus Fm.
ANUSI (93: 175-200 m)

needle breccia

PL 210, 215

m l a r b carbonate
ms f m c sandstone

175 meters
ANUSI (93: 200-225 m)

225
220
215
210
205
200 meters

Nudaus Fm.

Nasop Mbr.

S2
S1

m l a r b carbonate
m s f m c sandstone

MP
MP
MP
There is no text material missing here. Pages have been incorrectly numbered.

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APPENDIX C

Measured sections in the middle Schwarzrand Subgroup
(Huns, Feldschuhhorn and Spitskop members)

Part 1: Photographs of typical facies
Part 2: Locations of measured sections
Part 3: Generalized measured sections
Part 4: Selected measured sections
Part 1:

Photographs of facies in the middle Schwarzrand Subgroup
(Huns, Feldschuhhorn and Spitskop members)
Exposures of the middle Schwarzrand Subgroup (Huns, Feldschuhhorn, and Spitskopf members) on the farm Arimas (20).
Cross-stratified grainstone

Interbedded grainstone and calcisiltite

Laminated calcisiltite
Karst surface

Rudstone
Elongate, laterally-linked stromatolites

Large dome-shaped stromatolite

Columnar stromatolites/thrombolites
Part 2:
Locations of sections in the middle Schwarzrand Subgroup
(Huns, Feidschuhhorn and Spitskop members)
<table>
<thead>
<tr>
<th>Number</th>
<th>Latitude S</th>
<th>Longitude E</th>
<th>Farm</th>
</tr>
</thead>
<tbody>
<tr>
<td>10</td>
<td>27° 35'</td>
<td>17° 00'</td>
<td>Aub (87)</td>
</tr>
<tr>
<td>13</td>
<td>27° 25'</td>
<td>16° 47'</td>
<td>Tierkloof (75)</td>
</tr>
<tr>
<td>14</td>
<td>27° 30'</td>
<td>16° 45'</td>
<td>Swartpunt (74)</td>
</tr>
<tr>
<td>16</td>
<td>27° 25'</td>
<td>16° 47'</td>
<td>Tierkloof (75)</td>
</tr>
<tr>
<td>17</td>
<td>27° 32'</td>
<td>16° 36'</td>
<td>North Witputs (22)</td>
</tr>
<tr>
<td>20</td>
<td>27° 41'</td>
<td>17° 02'</td>
<td>Arimas (33)</td>
</tr>
<tr>
<td>24</td>
<td>26° 34'</td>
<td>16° 43'</td>
<td>North Witputs (22)</td>
</tr>
<tr>
<td>27</td>
<td>27° 28'</td>
<td>16° 31'</td>
<td>Swatkloofberg (95)</td>
</tr>
<tr>
<td>36</td>
<td>27° 26'</td>
<td>16° 28'</td>
<td>Swatkloofberg (95)</td>
</tr>
<tr>
<td>62</td>
<td>27° 35'</td>
<td>17° 26'</td>
<td>Bobbejaankrans (180)</td>
</tr>
<tr>
<td>64</td>
<td>27° 32'</td>
<td>17° 26'</td>
<td>Moedhou (182)</td>
</tr>
<tr>
<td>81</td>
<td>27° 08'</td>
<td>17° 16'</td>
<td>Geigoab (195)</td>
</tr>
<tr>
<td>115</td>
<td>27° 30'</td>
<td>16° 35'</td>
<td>Swatkloofberg (108)</td>
</tr>
<tr>
<td>134</td>
<td>27° 30'</td>
<td>16° 55'</td>
<td>Tierkloof (75)</td>
</tr>
<tr>
<td>140</td>
<td>27° 38'</td>
<td>16° 55'</td>
<td>Arimas (83)</td>
</tr>
<tr>
<td>141</td>
<td>27° 40'</td>
<td>17° 03'</td>
<td>Arimasi (73)</td>
</tr>
<tr>
<td>142</td>
<td>27° 42'</td>
<td>17° 11'</td>
<td>Arimas (72)</td>
</tr>
<tr>
<td>145</td>
<td>27° 29'</td>
<td>17° 00'</td>
<td>Abos (73)</td>
</tr>
<tr>
<td>146</td>
<td>27° 24'</td>
<td>16° 30'</td>
<td>Sonntagsbrunn (146)</td>
</tr>
<tr>
<td>160</td>
<td>27° 30'</td>
<td>16° 34'</td>
<td>Swatkloofberg (108)</td>
</tr>
<tr>
<td>162</td>
<td>26° 34'</td>
<td>16° 43'</td>
<td>North Witputs (22)</td>
</tr>
</tbody>
</table>
Part 3:
Generalized sections in the middle Schwarzrand Subgroup
(Huns, Feldschuhhorn and Spitskop members)
Part 4:
Measured sections in the middle Schwarzrand Subgroup
(Huns, Feldschuhhorn and Spitskop members)
SEDIMENTARY STRUCTURES

- planar-lamination
- low-angle cross-lamination
- tabular cross-stratification
- trough cross-stratification
- hummocky cross-stratification
- 3-D wave-ripple
- ripple cross-stratification
- wave ripples
- current ripples
- slump structures
- water-escape structures
- flute casts
- gutter casts
- mud cracks

GRAIN-TYPES

- pellets of rounded mud intraclasts
- pellets of rounded mud intraclasts
- coated grains; ooids
- grapestone
- skeletal grains
- mud chips
- form-fitted intraclast breccia
- microbreccia: form-fitted breccia with mm-scale intraclasts
- quartz-pebble conglomerate

PALEOCURRENT INDICATORS

- unidirectional current indicator for e.g. flute mark (FM), current ripple (CR), and trough azimuth (TA)
- bidirectional current indicator for e.g. gutter cast (GC), wave-ripple trend (RT), stromatolite elongation (S), and parting or current lineation (PL, CL)
### PRINCIPLE BED OR LAMINAE CHARACTERISTICS

- **thickly and evenly parallel-laminated**
- **low-angle cross-laminated**
- **thinnily and evenly parallel-laminated**
- **parallel-laminated**
- **thin-bedded, mud- and sand-rich layers**
- **nodular, irregularly bedded**
- **medium-bedded**

<table>
<thead>
<tr>
<th>Characteristic</th>
<th>Description</th>
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</thead>
<tbody>
<tr>
<td>thick-bedded; massive-weathering, vaguely planar-laminated</td>
<td><strong>MP</strong></td>
</tr>
<tr>
<td>large-scale trough cross-bedded</td>
<td></td>
</tr>
<tr>
<td>coarse-grained lenses</td>
<td></td>
</tr>
<tr>
<td>mudstone with thin sand-sized interbeds</td>
<td></td>
</tr>
<tr>
<td>elongate stromatolites</td>
<td></td>
</tr>
<tr>
<td>domal stromatolites</td>
<td></td>
</tr>
<tr>
<td>columnar/pedestal stromatolites/thrombolites</td>
<td></td>
</tr>
</tbody>
</table>

### COMPOSITION

- **limestone**
- **siliciclastic sandstone or mudstone**
- **mixed siliciclastic and carbonate sandstone**

<table>
<thead>
<tr>
<th>Composition</th>
<th>Description</th>
</tr>
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<tbody>
<tr>
<td>limestone</td>
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</tr>
<tr>
<td>siliciclastic sandstone or mudstone</td>
<td></td>
</tr>
<tr>
<td>mixed siliciclastic and carbonate sandstone</td>
<td></td>
</tr>
<tr>
<td>siliciclastic shale</td>
<td></td>
</tr>
<tr>
<td>carbonate mudstone</td>
<td></td>
</tr>
<tr>
<td>volcanic ash</td>
<td>vvv</td>
</tr>
</tbody>
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### GRAINSIZE

<table>
<thead>
<tr>
<th>Grain Size</th>
<th>Siliciclastic</th>
<th>Carbonate</th>
</tr>
</thead>
<tbody>
<tr>
<td>m</td>
<td>mud</td>
<td>m</td>
</tr>
<tr>
<td>v</td>
<td>very fine</td>
<td>w/l</td>
</tr>
<tr>
<td>f</td>
<td>fine</td>
<td>a</td>
</tr>
<tr>
<td>m</td>
<td>medium</td>
<td>r</td>
</tr>
<tr>
<td>c</td>
<td>coarse</td>
<td>b</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Description</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>laminite/calcsiltite</td>
<td></td>
</tr>
<tr>
<td>arenite/fine grainstone</td>
<td></td>
</tr>
<tr>
<td>rudstone</td>
<td></td>
</tr>
<tr>
<td>breccia</td>
<td></td>
</tr>
</tbody>
</table>
SECTION 10
Aub Farm (87)
AUB (10:0-25 m)

25
20
15
10
5
0

Nasep Mbr.

Huns Mbr.

m a w g b carbonate
m f m c c sandstone

meters
AUB (10: 50-75 m)

50 meters

m a w g b carbonate
m f m c c sandstone

1 cm
AUB (10: 75-100 m)

Huns Mbr.

m a w g b carbonate
m f m c c sandstone

75 meters
SECTION 14
Swartpunt Farm (74)
Swartpunt (14: 400-480 m)
SECTION 20
Arimas Farm (83)
ARIMAS (20: 75-100 m)

coarsens upward tp 3-5 c
scale intraclasts at top

m l a r b carbonate
m f m c c sandstone
ARIMAS (20: 125-150 m)

15-20 cm high, 3-5 cm wide

light gray

m l a r b carbonate

m f m c c sandstone
ARIMAS (20: 175-200 m)

2 m domes with 5-10 cm LLH along tops
light gray

3-5 cm intraclasts @ base;
fines upward

m l a r b carbonate
m f m c c sandstone
ARIMAS (20: 200-225 m)

200 meters

m l a r b carbonate
m f m c c sandstone
SECTION 27
Swartkloofberg Farm (95)
SWARTKLOOFBERG 76 (27: 0-10 m)

meters

m i a r b carbonate
m v f m c sandstone
SWARTKLOOFBERG 76 (27: 10-35 m)

10 meters

m | a | r | b | carbonate
m | v | f | m | c | sandstone

10 cm λ
SWARTKLOOFBERG 76 (2: 115-140 m)

115 meters

m i a r b carbonate
m v f m c sandstone

dolomitized fills

nodular dolomite <10%

starved
SWARTKLOOFBERG 76 (27: 140-165 m)

m | a | r | b | carbonate
m | v | f | m | c | sandstone

starved
SWARTKLOOFBERG 76 (27: 165-190 m)

140 cm deep karst pothole
Patchy, thrombolitic material
covering pothole surface

110 cm deep karst pothole

100 cm deep karst pothole

45 cm deep karst potholes
filled with intraclast breccia

m l a r b carbonate
m v f m c sandstone
SECTION 36
Swartkloofberg Farm (95)
Swartkloofberg (36: 160-240 m)

Feldschuhorn

Huns

Distinctive large (1-3 m)
pink stromatolites and thrombolites

160 meters
SECTION 62
Bobbejaankrans Farm (180)
see appendix B
SECTION 64
Moedhou Farm (182)
MOEDHOU 182 (64: 0-25 m)

m l a r b carbonate
m v f m c sandstone

0 meters
MOEDHOU 182 (64: 25-50 m)

---

m l a r b carbonate
m v f m c sandstone

---

25 meters

---

50
45
40
35
30
MOEDHOU 182 (64: 100-125 m)

- Chert nodules

- Carbonate

- Sandstone
MOEDHOU 182 (64: 200-225 m)

- Sandy at base
- Chert along surface
- Generalized: too steep to measure

200 meters

m l a r b carbonate
m v f m c sandstone
MOEDHOU 182 (64: 225-250 m)

scour and fill

m a r b carbonate

m v f m c sandstone
MOEDHOU 182 (64: 250-275 m)

- Some scour and drape up to 3 cm long
- Needle clasts 5-25 cm long clasts
- LM: a b carbonate
- MFMC: sandstone
MOEDHOU 182 (64: 275-300 m)

275 meters

m v f m c sandstone

m i a r b carbonate
MOEDHOU 182 (64: 300 m - top)

300 meters

m l a r b carbonate
m v f m c sandstone
SECTION 81
Geigoab Farm (195)
SECTION 115
Swartkloofberg Farm (195)
Swartkloofberg (115: 0-80 m)

- Nodular to irregularly bedded
- Carbonate
- Siliciclastic
SECTION 134
Tierkloof Farm (75)
SECTION 160
Swartkloofberg Farm (95)
Swartkloofberg (160: 0-80 m)

- Carbonate
- Siliciclastic
Swartkloofberg (160: 160-240 m)

160 meters

carbonate

siliciclastic
North Witputs (162: 240-320 m)

- Top of Hung Mbr.
- Thin-beaded carbonate mud and shale
- Carbonate siliciclastic
SECTION 162
North Witputs Farm (22)
North Witputs (162: 160-240 m)

poorly outcropping
same as
below

carbonate
North Witputs (162: 240-320 m)