GEOLOGICAL SURVEY OF NAMIBIA

MINISTRY OF MINES AND ENERGY

STRATIGRAPHY AND SEDIMENTOLOGY OF THE LATE PRECAMBRIAN WITVLEI AND NAMA GROUPS, EAST OF WINDHOEK

by

W. HEGENBERGER

MEMOIR 17

1993

Front cover: Stromatolites forming a characteristic layer in the uppermost portion of the Okambara Member. Photograph taken on the farm Okambara 219 by K.-H. Hoffmann. This section is about 27 cm long.

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ABSTRACT

East of Windhoek, outliers of Witvlei and Nama Group rocks occur in two synclinoria on the Southern Foreland of the late Proterozoic to early Palaeozoic Damara Orogen.

Those sediments are underlain by conglomerate, sandstone, shale and carbonate of the Tsumis Group which was deposited during early Damara rifting, followed by predominantly fluvial arenites of the Nosib Group (Kamtsas Formation). A patchy tillite (Blaubeker Formation) could either form the top of the Nosib Group or the base of the overlying Witvlei Group.

The Witvlei Group overlies the Tsumis and Nosib Groups unconformably and consists of two different sedimentary cycles (= sequences). The Court Formation at the base started with “deep-water” carbonate (laminite) passing into shallow-water carbonate (both of the Gobabis Member) and was followed by flood plain mudstone with carbonate intercalations (Constance Member). Both members were deposited in a confined, shallowing-upward basin, probably a lake. The cycle is terminated by fluvial sandstone of the Simmenau Member which received its sediments mainly from uplifted blocks in the southwest and can be interpreted as the latest stage of the rifting phase.

The younger cycle of the Witvlei Group, the Buschmannsklippe Formation, overlies the Court Formation unconformably and starts with light-coloured intertidal dolomite of the Bildah Member, passing into shallow subtidal limestone towards the top. It is followed by the marly La Fraque Member which was deposited in a muddy lagoonal to subtidal environment. The predominantly shallow subtidal lower unit of the Okambara Member consists of limestone with arenaceous intercalations and is characterised by edgewise conglomerate and hummocky cross-stratification. This type of crossbedding is also present in the arenaceous middle unit whereas the upper unit consists of dolomite with sandy intercalations containing flat -pebble conglomerate and, near the top, stromatolites and pseudomorphs after an evaporite mineral, indicative of regression. The Buschmannsklippe basin was most probably connected with the Damara ocean in the west and the formation is interpreted as a passive margin series, deposited on the edge of the Kalahari Craton during opening of the Damara ocean. Its fine-grained siliciclastic sediments were transported by tidal and storm-generated currents from the west.

The Witvlei Group occurs only in the study area and has been correlated with similar rocks in the Naukluft and Witputs-Rosh Pinah regions in western-central and southwestern Namibia.

A far-reaching transgression over a peneplaned surface initiated deposition of the Nama Group. This major flooding surface is a sequence boundary and is interpreted to indicate the reversal from spreading to convergence of the plate motion, about 650 million years ago. In the study area, however, the Nama basin may have developed as a continuation of the Buschmannsklippe basin and their common boundary would be the correlative conformity of the generally unconformable lower contact of the Nama Group. The Weissberg Member (Dabis Formation) at the base consists of mainly intertidal quartzite. The overlying Zaris Formation comprises several facies belts from shallow subtidal to fluvial environments. It reflects the onlap of the lower Nama Group sediments towards the Kalahari Craton. Most of the siliciclastic material was derived from the Kalahari Craton; a delta fan deposit, however, received its coarse-grained sediments from the north.

By comparing the situation of the study area in the Nama basin with that of a deep borehole in western Botswana, it can be inferred that the Schwarzrand and Fishriver Subgroups were deposited in the study area but have since been eroded.

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1. INTRODUCTION

The late Precambrian to early Palaeozoic Damara Orogen is part of the Pan-African orogenic belt that underlies a large tract of Namibia where it can be subdivided into several geological provinces (Miller, 1983; Hoffmann, 1987, and Fig. 1.1). Three metamorphic belts, the Kaoko Belt in the northwest, the northeast-trending Intercontinental Belt and the Gariep Belt in the south, are accompanied by two external zones in the north and southeast respectively.

During its history, the orogen passed through a Wilson cycle, evolving through the successive stages of rifting, spreading, plate convergence, subduction, and continental collison (Miller, 1983).

Initial rifting is characterised by fault grabens filled with fluvial, lacustrine, shallow marine and volcanogenic deposits with rapid facies changes. On the Kalahari Craton these deposits belong to the Tsumis Group, comprising the Bitterwater, Aubures, Doornpoort, Klein Aub and Eskadron Formations (Hoffmann, 1989) and were formerly correlated with the Sinclair Sequence (SACS, 1980).

The overlying Nosib Group consists mainly of fluvial arenites and represents later rift filling. The contact between sediments of the Tsumis Group and the Nosib Group is unconformable to paraconformable, indicating the absence of a major deformation event between these groups.

During the further evolution of the Damara Orogen, sedimentation on the two external zones of the orogen differed widely, due to the different structural histories of the zones. A carbonate platform lasted for most of the passive margin stage in the north, whereas erosion prevailed to the south of the Damara basin. Sedimentation started only toward the end of the spreading phase of the Intercontinental Belt or even after subduction had begun (Miller, 1983), resulting in a relatively thin mixed carbonate and clastic wedge in a continental-shelf sea on the margin of the Kalahari Craton. Initially the clastic sediments were mainly derived from the Kalahari Craton in the east, and later, as a molasse, from sources within the rising Damara Belt to the north and west (Germs, 1983).

The deposits of the southern external zones underlie a vast area in the southeastern portion of the country, as well as in adjoining parts of South Africa and Botswana (Fig. 1.1; Germs, 1972a, 1983; Gresse, 1986; Reeves, 1979; Aldiss, 1988). In the study area they have been subdivided into a lower Witvlei Group and an upper Nama Group.

Deposition of the Witvlei and Nama Groups on the southern external zone took place between about 650 and 530 million years ago (Miller, 1983). These approximate age limits are based on the presence of an Ediacara-type fauna in the lower portion of the Nama Group (Germs, 1972a, b; Cloud and Glaessner, 1982; Glaessner, 1984), on the age of detrital mica contained in certain Nama beds (Ahrendt *et al.*, 1985; Horstmann, 1987; Weber *et al.*, 1983), on the age of white micas of diagenetic origin in Nama rocks from the environs of the Naukluft mountains (Ahrendt *et al.*, 1977), on the intrusive age of the Bremen Complex (SACS, 1980), as well as on palaeomagnetic investigations (Kröner *et al.*, 1980). All results indicate that the Witvlei Group and the lower portion of the Nama Group, the only units dealt with in this memoir, are Vendian (Upper Proterozoic) in age, while the upper Nama Group sediments are Cambrian (Hartnady, 1978).

Most of the strata deposited in the southern external basin after rifting had ceased belong to the Nama Group. In two areas, however, these are unconformably to paraconformably underlain by sediments which are younger than the Nosib Group. Along the Gariep Belt they can be followed from near Witpütz to south of the Orange River, where they are named the Witputs and Gariep Groups (Hoffmann, 1989; Kröner and Germs, 1971; Martin, 1965a; McMillan, 1968). The second occurrence is in the study area (Figs 1.1 and 1.2) where the new name “Witvlei Group” is introduced. It encompasses the Court and the Buschmannsklippe Formations of the previous classification, which formed the basal part of the Nama Group (Hegenberger and Seeger, 1980; SACS, 1980).

Germs (1972a, 1983) and SACS (1980) distinguished three subgroups of the Nama Group, viz. the Kuibis Subgroup at the base, overlain by the Schwarzrand Subgroup and the Fish River Subgroup at the top. The development of the passive to an active margin is revealed by a major shift in the provenance areas of the clastic sediments. During Kuibis and lower Schwarzrand times, the detritus was mainly derived from the Kalahari Craton situated to the east of the basin, whereas erosion of the rising Damara Orogen situated to the north and west of the foreland basin provided most of the material of the upper Nama Group (Germs, 1983) which is thus interpreted as a molasse of the Damara Orogen.

The floor of the Southern Foreland basin consists of various stratigraphic units. A zone transitional between the facies of the externides and the internides was overridden during Damaran orogeny by nappes, and partly eroded along a major uplift along the southeastern margin of the orogen. The external and internal facies are therefore generally separated by a belt of pre-Witvlei rocks. Only in the Naukluft Nappe Complex have sediments been preserved which can be interpreted as a facies transition between the Witvlei-Nama Groups and the stratigraphic units of the internal zone (Miller, 1983; Hoffmann, 1989).

Most of the external succession of the Damara Sequence extends from the Aranos Syncline into the Nosop, Ncojane and possible further northeastward into the Passarge Synclines in Botswana (Fig. 1.1). These sediments probably did not extend eastward beyond the Kalahari Line, situated in central Botswana (Reeves, 1979; Meixner, 1984a; see Fig. 1.1). The Aranos, Ncojane and Nosop Synclines are probably younger than the Damara Sequence, the deposition of which they did not influence. They did, however, form sedimentary basins during Karoo times, below which the Witvlei and Nama sediments were preserved from erosion.

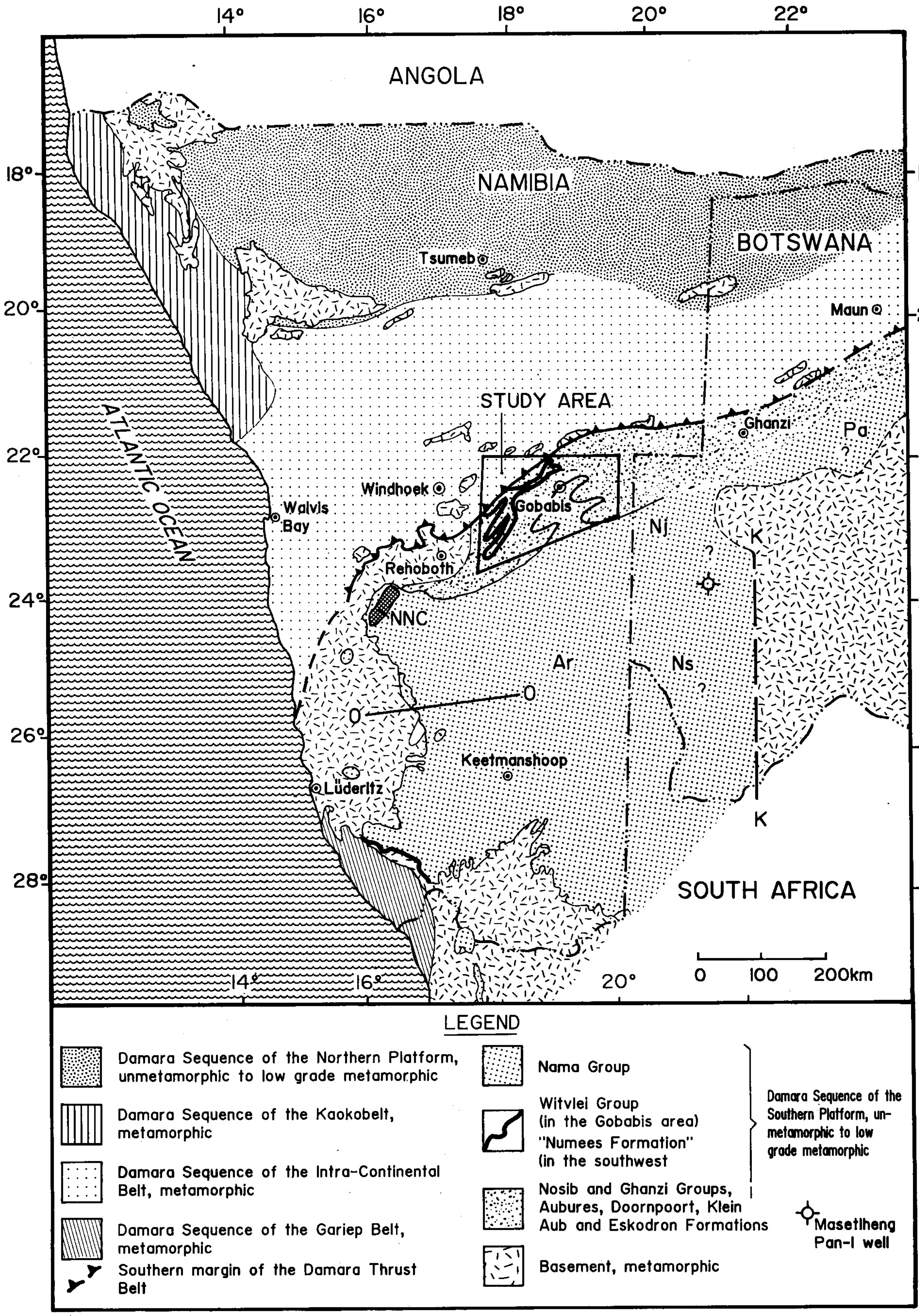


Figure 1.1: Generalised geological map of Namibia, western Botswana and northwestern South Africa, showing distribution of the Witvlei and Nama Groups and the location of the study area (post-Damara formations omitted). Ar-Aranos Syncline, Nj-Ncojane Syncline, Ns-Nosop Syncline, Pa-Passarge Syncline, NNC-Naukluft Nappe Complex, K--K Kalahari Line, O--O Axis of the Osis Ridge. (After Geological map of SWA/Namibia, 1980; Geological map of South Africa, 1984; Miller, 1983; Reeves, 1979)

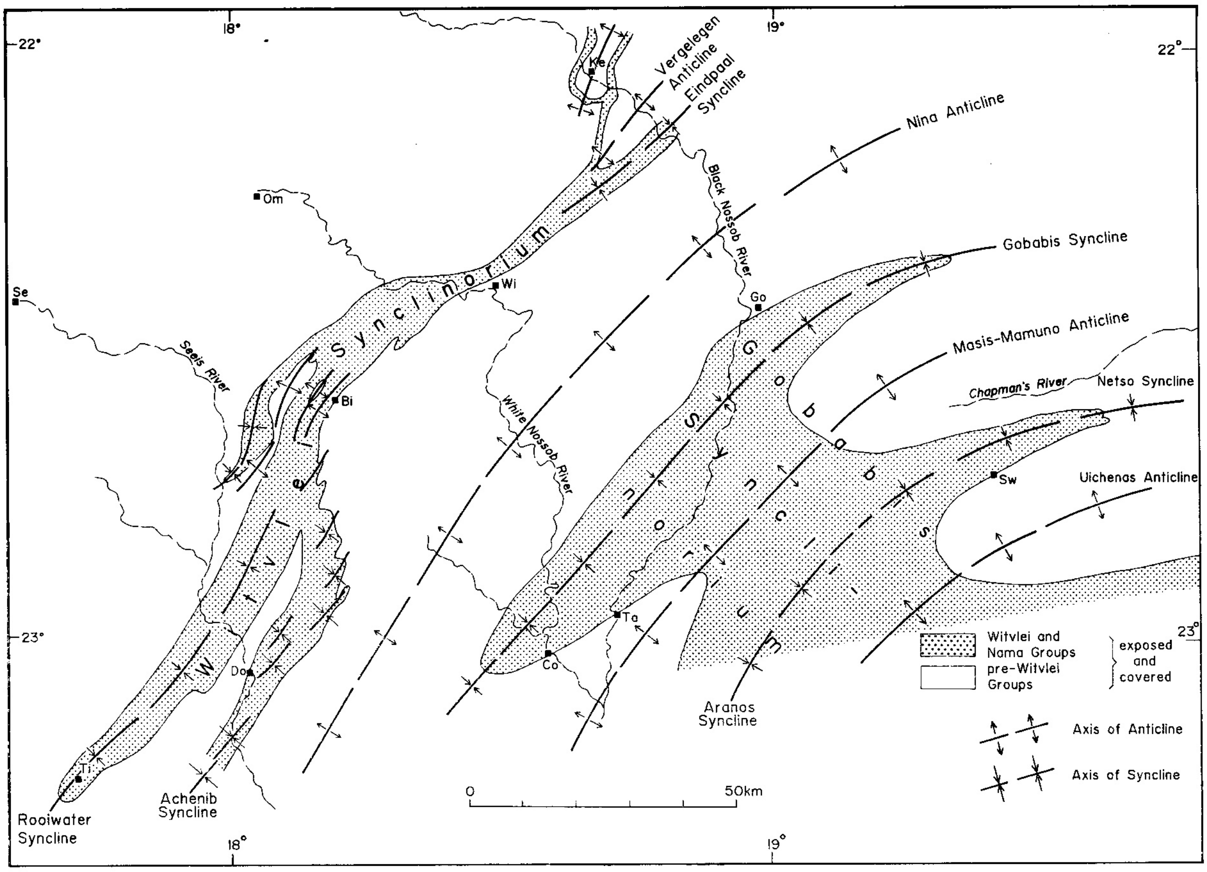


Figure 1.2: Tectonic map of the study area. Localities: Bi-Bildah, Co-Court, Do-Doornpoort, Go-Gobabis, Ke-Kehoro, Om-Omitara, Se-Seeis, Sw-Schwarzwald, Ta-Tahiti, Ti-Tigerpforte, Wi-Witvlei

Note: In the following description, areal numbers (like 2218 AB) refer to the topographic sheets at a scale of 1:50 000 on which the respective locality occurs. The Geological Survey of Namibia, Windhoek, is in possession of a number of unpublished geological maps at a scale of 1:50 000 covering the study area. Sheet 2218 Gobabis was compiled by means of reducing these maps to a scale of 1:250 000.

2. HISTORY OF INVESTIGATION

Rocks occurring in the Gobabis and Witvlei areas were first correlated with the Nama succession by Rimann (1913 and 1916) who described outcrops of "Kuibis" and "Schwarzkalk" beds and included several synclinal structures formed by these on his geological map. Influenced by Schenck's (1885a and b) and Range's (1910a and b) geological descriptions of "Namaland", Rimann also included very thick, fairly flat-lying quartzite underlying large tracts of the area, as "Basal Beds" in the Nama succession. De Kock (1934) and Gevers (1934) later gave geological descriptions of the Rehoboth and Dordabis areas, and the true stratigraphic position of the "Basal Beds", i.e. belonging to the Nosib Group, a succession that was slightly deformed and partly eroded prior to deposition of the "Kuibis" beds, was elucidated only much later by Schalk (1958, 1970), Schalk and Hälbich (1965), and Hegenberger and Seeger (1980).

Schalk (1958, 1961) described in some detail a succession of limestones and quartzites occurring in the eastern Windhoek District as the "Buschmannsklippe Formation", strongly resembling the lower Nama succession. Further knowledge was obtained during 1966 and 1967, when Schalk and Seeger compiled a preliminary geological map of the Witvlei area at a scale of 1: 100 000. The Buschmannsklippe Formation was included in the Nama Group by Martin (1965a) and Schalk and Hälbich (1965). The name "Buschmannsklippe" was used for a formation of the Kuibis Subgroup by SACS (1980) and now for a formation of the Witvlei Group.

Hegenberger and Seeger (1980) separated the Court Formation from the Buschmannsklippe Formation and both formations were placed at the base of the Kuibis Subgroup by SACS (1980). Correlations with similar strata in other parts of the country and the northern Cape Province have been attempted by Martin (1965a), Kaufmann *et al.* (1990) and Hoffmann (1989).

Patchy occurrences of mixtite that directly overlie the Nosib Group in the southern portion of the study area were called Buschmannsklippe tillite by Martin (1965a and b). Its stratigraphic position, however, remained uncertain (Martin, 1965a; SACS, 1980; Kaufmannn *et al.*, 1990). Schalk (1970) correlated it with the Blaubeker Formation tillite of the uppermost Nosib Group of the Rehoboth area. According to Hoffmann (1989) the mixtite is the lowest formation of the Witvlei Group.

The Court and Buschmannsklippe Formations are now separated from the Nama Group and form the new "Witvlei Group". The reason for this subdivision is that everywhere that the base of the Nama Group is exposed, with the exception of the Witputs area in the southwest and the northwestern Cape, it is formed by the Kanies Member of the Dabis Formation or, locally, by younger units of the Kuibis Subgroup. As the Kanies Member is correlated with the Weissberg Member of the study area, uniformity for the base of the Nama Group is achieved by the regrouping. For correlation of the Witvlei Group with the Witputs and Gariep Groups further southwest see Hoffmann (1989). Both Witvlei and Nama Groups are part of the Damara Sequence.

3. REGIONAL SETTING

The Witvlei and Nama Groups have been extensively eroded in the Windhoek and Gobabis Districts; their present occurrence is restricted to the Witvlei and Gobabis Synclinoria which are separated from each other by the Nina Anticline, a broad outcrop of the Kamtsas Formation arenites of the Nosib Group (Fig. 1.2).

In the western portion of the study area, the Witvlei Synclinorium extends for about 180 km in a southwest-northeast direction; the structure is up to 20 km wide. Its northwestern portion has been overridden by thrust sheets of pre-Damara, Nosib and Swakop Group rocks during Damaran orogenesis. The marginal thrust plane of the Damara Orogen forms the northwestern boundary of the Witvlei Synclinorium (Fig. 3.1).

Tightly folded Witvlei and Nama strata form steep-sided hills and ridges in the central part of the synclinorium. These consist largely of quartzite of the Weissberg Member, whereas the overlying shaly successions are deeply eroded and occupy the intervening valleys and plains. Once erosion has removed the crestal portions of the anticlines formed by Weissberg quartzite, the underlying softer rocks of the Buschmannsklippe Formation weather away more quickly than the fringing quartzite, and eroded centres of anticlines tend to form oblong valleys, leading to relief inversion (Plates 3.1 and 3.2).

Except for two oblong outcrops in the Kehoro area (2218 BA, Fig. 3.1), the northeastern extension of the Witvlei Synclinorium is covered by Kalahari deposits. The southwestern end of the Witvlei Synclinorium, now also mainly covered by Kalahari sand, is underlain by rocks of the Doornpoort and Kamtsas Formations which are exposed at the southwestern tip of the Rooiwater Syncline (Figs 1.2 and 3.1). Rocks resembling Witvlei and Nama strata have been intersected in boreholes drilled in the southwestern continuation of the exposed portion of the Achenib Syncline on the farms Albano 264, Kiripotib 262 and Tsams 75 (2318 AC) (K.E.L. Schalk, pers. comm., 1980), some 40 km southwest of the nearest surface outcrops of these rocks. Because of the continuous thick cover of Kalahari beds and basalt of the Kalkrand Formation (Karoo Sequence) the extent of the Achenib branch of the Witvlei Synclinorium to the southwest is not known. South of Rehoboth the sediments of the lower Nama Group originally extended far to the north of their present limits as is shown by the occurrence of an outlier of dark carbonate, most probably of the Zaris Formation, on the farm Oagoub 385 in area 2317 CB, and another of gritty, feldspathic quartzite of the basal Dabis Formation on Diergaards Aub 454 in area 2317 CA (Schalk, 1965a and b).

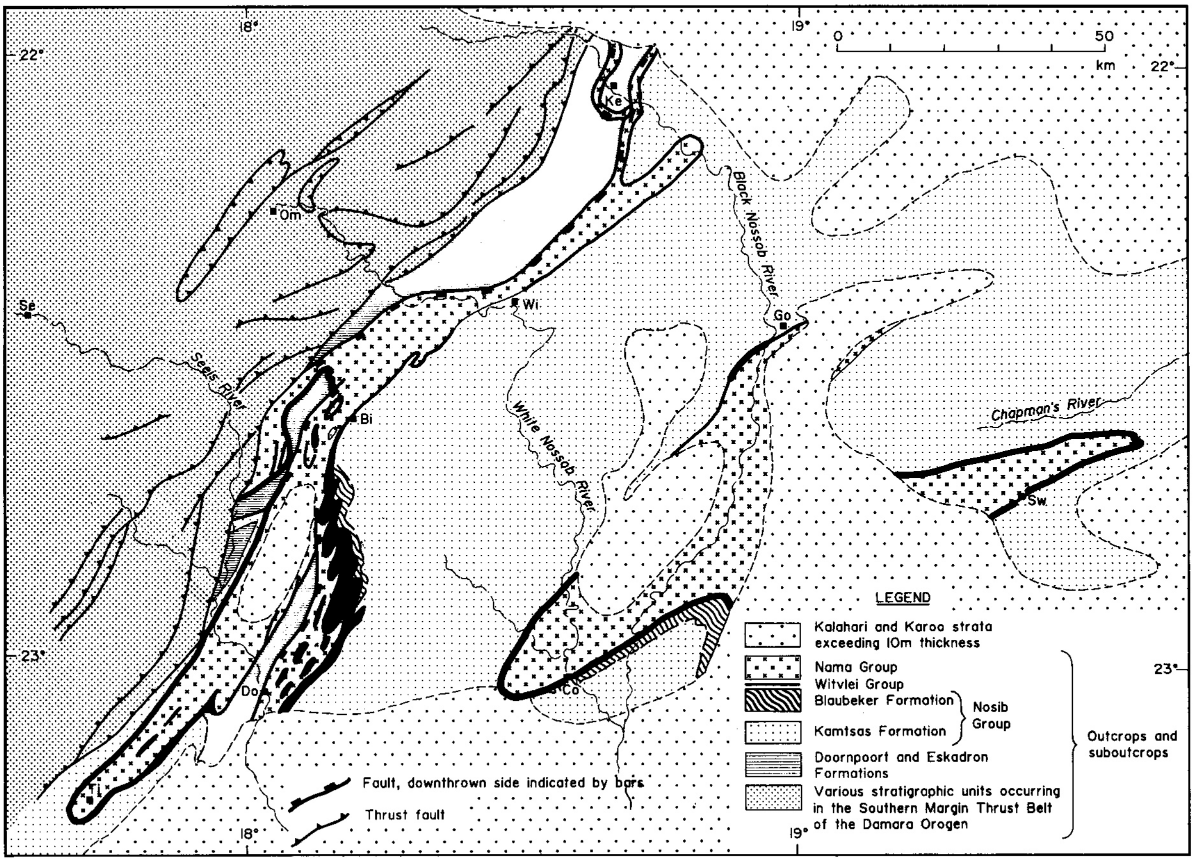


Figure 3.1: Generalised geological map of the study area. (After Geological map of SWA/Namibia, 1980; Geological map, Gobabis, 1981). For localities see Figure 1.2

Plate 3.1: For explanation see facing page

The Gobabis Synclinorium comprises three major substructures (Fig. 1.2), viz. the Gobabis Syncline, the Netso Syncline (Schalk, 1970) and the intervening Masis-Mamuno Anticline (Litherland, 1976) which, further northeast, forms the Ghanzi Ridge, a prominent positive landform of pre-Nama rocks that can be followed over a distance of about 400 km into northern Botswana. Although the Witvlei and Nama strata have been eroded on the southeastern flank of the Ghanzi Ridge, they could have been preserved in deep depressions like the Passarge Basin (Reeves, 1979) in continuation of the Gobabis Synclinorium.

Most of the strata under consideration are covered by Karoo and Kalahari sediments in the Gobabis Synclinorium. Exposures are found along the valleys of the Black and White Nossob Rivers and at the eastern tip of the Netso Syncline. East of the Black Nossob River the Gobabis Synclinorium opens southward into the Aranos Synclinorium. Nama rocks occur below a 250-m-thick cover of Kalahari and Karoo beds in a borehole drilled at Aminuis (area 2319 AB) (Hegenberger, 1985).

Plate 3.1 (left): Central Witvlei Synclinorium. Aerial photograph covers approximately the area of sheet 2218 CA, extending from the farm Weshof 585 in the north to the Buschmannsklippe mountains in the southwest and Josephine 226 in the south. The mountain ranges rise 200 to 300 m above the surrounding plain. The synclinorium consists of synclines and anticlines that strike approximately southwest-northeast (for details see Plate 3.2). The area is divided by a northeast-trending fault into two nearly equal parts. This fault cuts off the eastern portion of the Okombuka Anticline. On the eastern side of the fault, the Witvlei and Nama strata (bu, da, z) are underlain by Kamtsas Formation (ka) (Nosib Group), on the western side by Eskadron Formation (es) (Tsumis Group), suggesting early Damaran, post-Kamtsas - pre-Witvlei movement on the fault (rifting).

The structural form lines in the Okombuka Anticline indicate tight folding of the Eskadron rocks prior to deposition of the Witvlei strata, while the form lines in the east, on Frank 221, show local open folding of the Kamtsas Formation prior to deposition of the Witvlei Group. During the Damaran orogeny the fault was rejuvenated and the Okombuka Anticline upthrown against the Okambara Anticline. During the same event Kamtsas rocks of the Southern Margin Zone were thrusted eastward over the Nama rocks. The boundary thrust between the Southern Margin Zone and the Foreland of the Damara Orogen runs about north-south on the western margin of the photograph.

Legend: z-Zaris Formation, da-Dabis Formation, bu-Buschmannsklippe Formation, co-Court Formation, ka-Kamtsas Formation, es-Eskadron Formation. Aerial photograph Job 827, Strip 3, Photo 662

Plate 3.2 (overleaf): Eastern central portion of Plate 3.1. The fault in the northeastern corner separates areas in which the Witvlei and Nama rocks are underlain by Eskadron Formation from those in which they rest on Kamtsas Formation (see Plate 3.1). Structural form lines of the Kamtsas rocks in the southeast indicate a paraconformable contact with the overlying Witvlei rocks. Deformation of the Witvlei and Nama strata resulted in doubly plunging, open to tight folds most of which are symmetrical; some are overturned to the east. Interference patterns of the folds, especially obvious in the northwestern portion of the photo, are the effect of two phases of deformation with rotation of the older structures.

The different resistance to weathering of the rock formations resulted in various landforms. Kamtsas and Zaris Formations underlie flat country which is covered by sand and loam. Steeply rising Dabis quartzite (Weissberg Member) mantles the antiforms; its light colour contrasts with the darker one of the Okambara Member, with red-brown quartzite intercalations. The quartzitic middle unit of the Okambara Member forms a ridge (scarp), visible as a dark line on the photo, and is overlain by the slightly lighter-coloured upper unit which gives rise to flat topography. The La Fraque Member is partly covered by calcrete and stands out as a lighter tint, and is also less steep than the Okambara Member. The light-grey Bildah carbonate is speckled with bushes; steep rises of the Bildah Member on the plain of the Kamtsas quartzite are shaded. Alluvial fans formed where the courses of rivulets pass from mountainous areas to the flat foreland.

Legend: af-alluvial fan, s,l-sand and/or loam, scree, z-Zaris Formation, da-Dabis Formation, ok-Okambara Member, lf-La Fraque Member, bi-Bildah Member, ka-Kamtsas Formation, es-Eskadron Formation; symbol in brackets indicates a unit under cover. Aerial photograph Job 638, Strip 9, Photo 290

Plate 3.2: For description see page 9

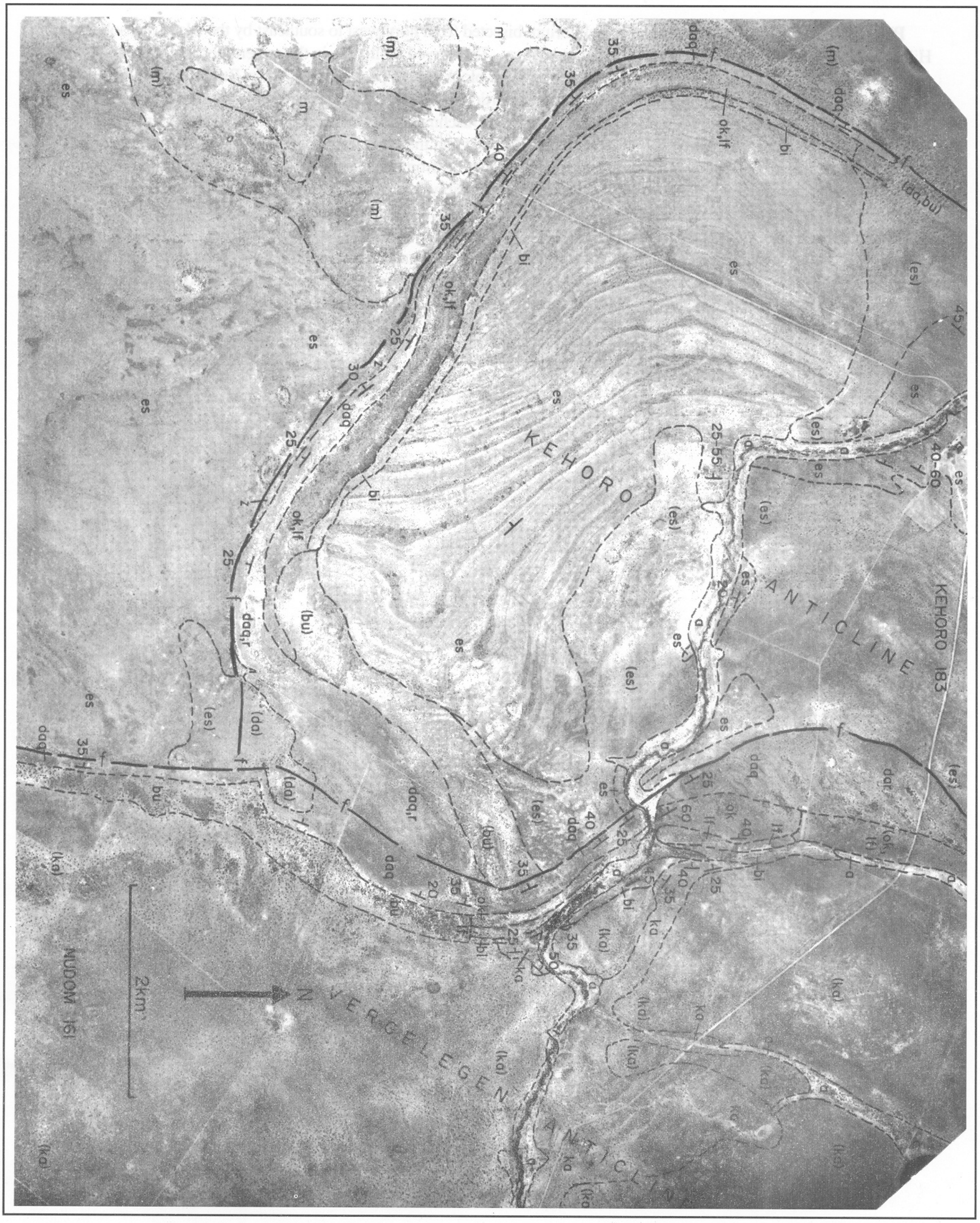


Plate 3.3: For description see page 12

Plate 3.3 (left): Northeastern Witvlei Synclinorium bisected from northwest to southeast by the Black Nossob River. Hills rise less than 20 m-above the surrounding flat country. The curved fault zone in the eastern portion of the photo has a complex history. It started with an upthrow of the western block during post-Kamtsas - pre-Buschmannsklippe times. Subsequent erosion removed the Kamtsas quartzite and a portion of the upper Eskadron succession from the western fault block and resulted in planation, allowing transgression of the Buschmannsklippe sea across the fault and deposition of its sediments on both sides. A post-Nama rejuvenation of the fault had the same sense of throw as the older one. The fault delimitating the Kehoro Anticline in the west and south branches off from the north-south fault. Eskadron Formation occurs on both sides of the fault, indicating that it was most probably only active during the second, post-Nama, phase of faulting. It was folded together with the Kehoro Anticline. The Buschmannsklippe and Nama rocks overlie the Eskadron Formation with an angular unconformity.

Legend: a-alluvium, z-Zaris Formation, da-Dabis Formation (daq-quartzite, dar-conglomerate), bu-Buschmannsklippe Formation undifferentiated, ok-Okambara Member, lf-La Fraque Member, bi-Bildah Member, ka-Kamtsas Formation, es-Eskadron Formation, m-Marienhof Formation; symbol in brackets indicates a unit under cover. Aerial photograph Job 666, Strip 2, Photo 150

4. STRATIGRAPHY

The lithostratigraphic subdivision of the Witvlei and Nama Groups is given in Table 4.1, the generalised geology of the study area in Figure 3.1 and geological sections in Figures 4.1 and 4.2.

4.1 FLOOR ROCKS

As the rocks along the western margin of the Kalahari Craton formed the source for the clastic sediments of the Witvlei and lower Nama Groups, their lithologies are briefly described.

Basic lava and metasediments of the **Marienhof Formation** (Rehoboth Sequence) (SACS, 1980) occur in horsts along the Southern Margin Zone of the Damara Orogen and one locality contributed sediments for the Witvlei Group.

Parts of the Witvlei Synclinorium are underlain by the **Eskadron** and **Doornpoort Formations** of the Tsumis Group (Hoffmann, 1989). The Eskadron Formation is an arenaceous succession of argillite and conglomerate with intercalated carbonate lenses and stringers. North of Witvlei the succession apparently has a thickness of about 10 000 m (Hegenberger and Seeger, 1980; Ruxton, 1981; Ruxton and Clemmey, 1986). Southwest of the farm Okombuka 218 (2218 CA) the Witvlei Group rests upon fine-grained, arkosic quartzite, including two layers of basaltic lava; these are assigned to the Doornpoort Formation (Schalk, 1970) of the Tsumis Group. The age of this Group is estimated to be between less than 1000 Ma and about 800 Ma (Hoffmann, 1989). The Eskadron and Doornpoort formations had been faulted, uplifted and tilted prior to the deposition of the Witvlei sediments, therefore the contact between them forms an angular unconformity, seen best on aerial photographs of Kehoro S 939 in area 2218 BA (Plate 3.3).

The **Kamtsas Formation (Nosib Group)** rests on the Doornpoort Formation with an angular unconformity (Schalk, 1970) and is about 5000 m thick. In the Gobabis area the mainly quartzose sequence has been subdivided into three units (Killick, 1983), the lower, middle and upper unit.

The lower unit, some 5000 m thick, occurs throughout the area between the Witvlei and Gobabis Synclinoria.

The middle unit forms a prominent range of hills which flank parts of the Witvlei and Gobabis Synclinoria. This unit is the main floor rock of the Witvlei and Nama Groups in the study area. It can be followed along the eastern margin of the Witvlei Synclinorium where it is most prominent around Witvlei. It consists of coarse- to medium-grained, feldspathic quartzite including scattered well-rounded pebbles of quartz, quartzite and subordinate igneous rocks. Near Witvlei, an about 5-m-thick conglomerate layer is directly overlain by Witvlei sediments. A similar range of hills, where a thin conglomerate layer also occurs, extend along the western portion of the Gobabis Synclinorium, from the Nikodemusberg north of Gobabis, along the Spitzkopf and Langer Forst, to the Nolteberg in the southwest. On the northern side of the Netso Syncline the “Groot Duin” hill consists of the same quartzite succession; here, however, the conglomerate is absent. The middle unit is about 1200 m thick (Killick, 1983). Near Gobabis and along both flanks of the Netso Syncline the upper unit is intercalated between the middle unit and the Witvlei Group sediments, comprising reddish-brown, fine-grained quartzite, siltstone and silty shale. Massive, medium- to coarse-grained arkosic quartzite in places forms the uppermost part of the upper unit. From a maximum thickness of about 1600 m near Gobabis and along the northern margin of the Netso Syncline, this unit thins out to the southwest, and the Witvlei Group oversteps onto the middle unit.

In the southern portion of the study area, the middle unit of the Kamtsas Formation is overlain by the **Blaubeker Formation** of the Nosib Group (Schalk, 1970), a diamictite comprising unsorted boulders and pebbles in an unstratified fine-sandy to shaly matrix. The main outcrops lie in the valleys of the White and Black Nossob Rivers in the southern Gobabis Synclinorium (Fig. 3.1). Here the clasts measure up to 50 cm across and consist mainly of Doornpoort and Kamtsas quartzite, with subordinate gneiss, rhyolite, basalt, diorite, granite, vein-quartz and mica schist. Toward the top of the mixtite, the proportion and size of the clasts decrease. Faceted and striated pebbles indicate a glacial origin (Martin, 1965a and b; Kröner and Rankama, 1973). The thickness is estimated to be about 450 m on the farm Court 32. Intermittent outcrops occur along the eastern margin of the Witvlei Synclinorium between the farms Constance 230 and Josephine 226 (areas 2218 CA and CC; Fig. 3.1). The mixtite is either immediately overlain by Witvlei rocks or separated from these (in the Witvlei Synclinorium) by up to 100 m of quartzite similar to that of the Kamtsas Formation from which it has probably been derived by reworking.

According to Martin (1965a) the diamictite forms the base of what he called the Buschmannsklippe Formation (now Witvlei Group). Schalk (1970) correlated it with the Blaubeker Formation of the Nosib Group in the Rehoboth area. Lately Hoffmann (1989) has revised its stratigraphic position and placed it at the base of the Witvlei Group. This is corroborated by palaeomagnetic investigations (Henthorn, 1976) which indicate that the Kamtsas Formation is considerably older (approximately 1100 Ma on the polar wander curve) than the Blaubeker Formation (about 650 Ma). The pole of the latter has a similar position on the polar wander path to that of the Kuibis Subgroup. The patchy occurrence of the mixtite indicates major unconformities between the Kamtsas and the Blaubeker Formations on the one hand and the Blaubeker and the Court Formations on the other.

This memoir does not deal with the Blaubeker Formation and still follows Schalk’s (1970) stratigraphic subdivision.

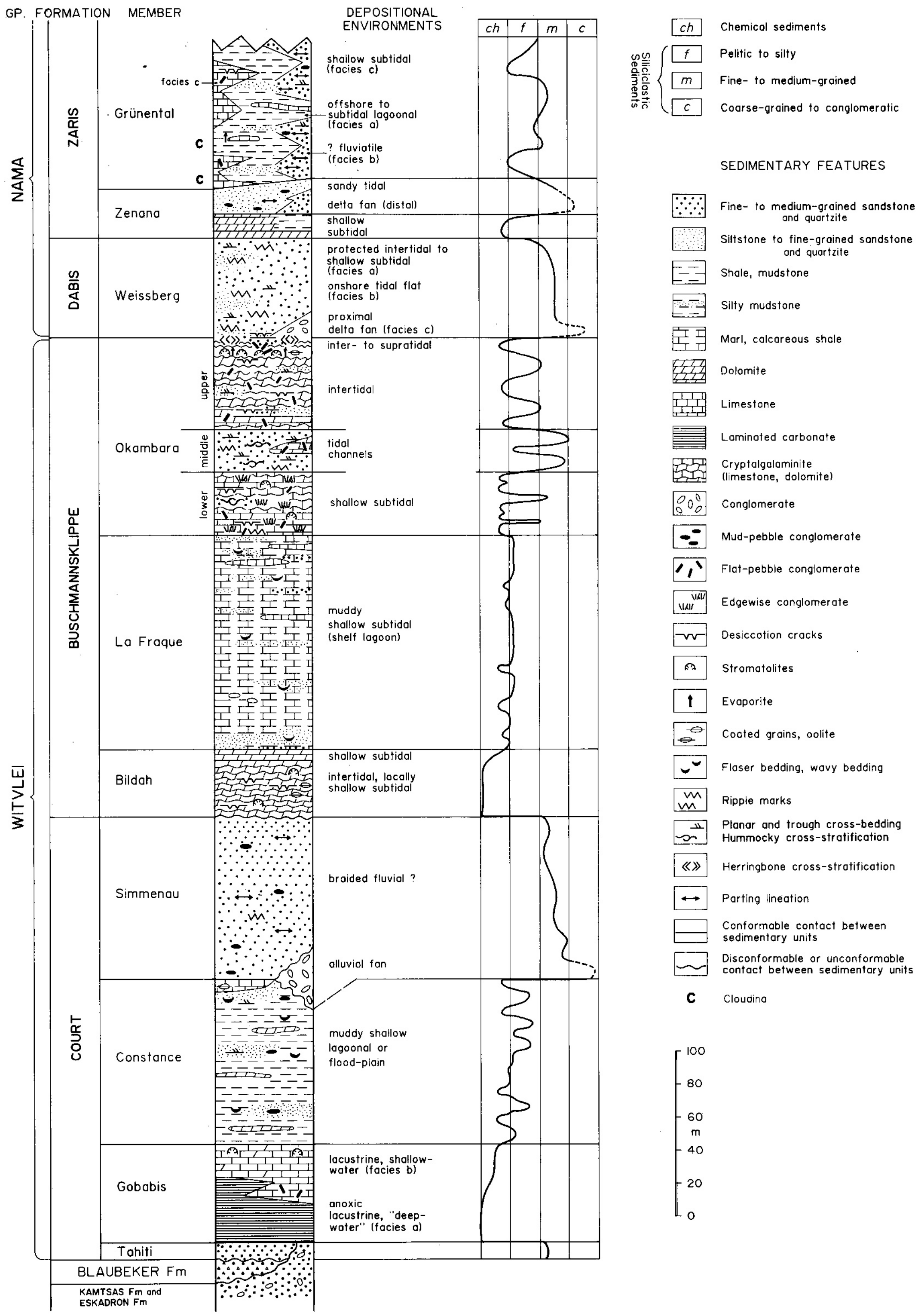
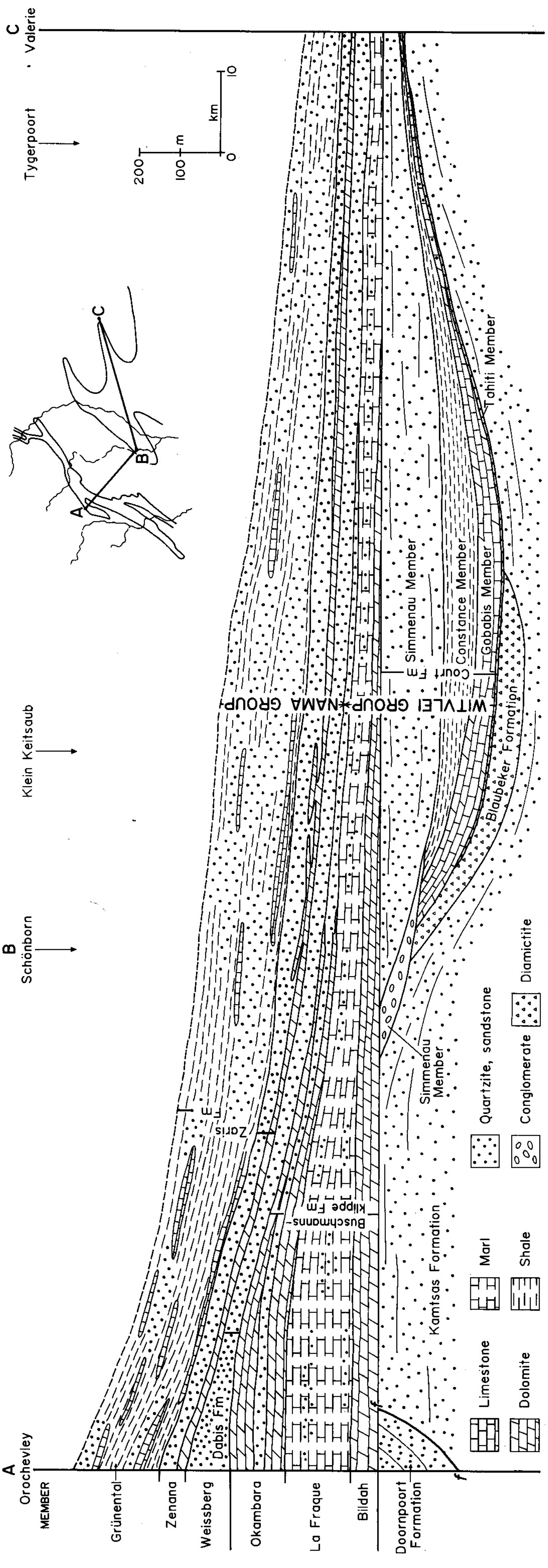


Figure 4.1: Idealised vertical geological section of the Witvlei and Nama Groups in the study area. Small occurrences of beach facies are omitted

Figure 4.2 (right): West—east section (A - B - C) through portions of the depositional basin of the Witvlei and Nama Groups east of Windhoek. Take note of: local discordance between Constance and Simmenau Members of the Court Formation; transgression of the Bildah Member over Doornpoort Formation (which had been thrusted southwestwards upon Kamtsas rocks), Kamtsas Formation and Court Formation, leading to an unconformable contact with the former and disconformable contacts with the latter two; paraconformable and concordant contacts between all members overlying the Court Formation - some are gradational; increasing thickness of all members overlying the Court Formation from east to west, i.e. away from the Kalahari Craton in direction of the Damara basin


A
Orochevley
B
Schönborn
Klein Keitsaub
Tygerpoort
C
Valerie
200
100
m
0
km
10
A
B
C
MEMBER
Grünental
Zenana
Weissberg
Okambara
La Fraque
Bildah
Doornpoort Formation
Dabis Fm
Buschmannsklippe Fm
Zaris Fm
Simmenau Member
Kamtsas Formation
WITVLEI GROUP
NAMA GROUP
Simmenau Member
Court Fm
Constance Member
Gobabis Member
Tahiti Member
Blaubeker Formation
Limestone
Dolomite
Marl
Shale
Quartzite, sandstone
Conglomerate
Diamictite

Group	Formation	Member	Main lithology	Correlates in the Aranos Basin
NAMA	Zaris	Grünental	sandstone (facies b)	absent
			shale, dark limestone (facies c)	Urikos
			dark limestone (facies c)	Hoogland, Omkyk
		Zenana	quartzite (upper unit)	Kliphoek
			dark dolomite (lower unit)	Mara
	Dabis	Weissberg	orthoquartzite (facies a) submature quartzite (facies b) conglomerate (facies c)	Kanies
- - - - - - - - -	- - unconformity,	in places	paraconformity or conformity	- - - - - - - - - - - - - -
WITVLEI	Buschmanns-klippe	Okambara	light-grey dolomite (upper unit) quartzite (middle unit) light-grey limestone (lower unit)	absent
		La Fraque	shale, marl, subordinate carb.	absent
		Bildah	light-grey and pink dolomite	absent
	- - - - - - - - - - - - -	- - - - - - - - - -	- - - - - - - - unconformity - - - - - - -	- - - - - - - - - - - - - -
	Court	Simmenau	quartzite (main facies) conglomerate (subordinate)	absent
		Constance	shale, siltstone, subordinate carb.	absent
		Gobabis	dark carb. (laminated or massive)	absent
		Tahiti	quartzite	absent
?	- - - - - - - - - - - - -	- - - - - - - - - -	- - - - - - - - - unconformity - - - - - - -	- - - - - - - - - - - - - -
	Blaubeker		mixtite	locally present
?	- - - - - - - - - - - - -	- - - - - - - - - -	- - - - - - - - - unconformity - - - - - - -	- - - - - - - - - - - - - -
NOSIB	Kamtsas		quartzite, conglomerate	present in the north
- - - - - - - - -	- - - - - - - - - - - - -	- - - - - - - - - -	- - - - - - - - - unconformity - - - - - - -	- - - - - - - - - - - - - -
TSUMIS	Eskadron		quartzite, shale, carbonate	
	Doornpoort		quartzite, conglomerate	Aubures

Table 4.1: Lithostratigraphic units of the Damara Sequence in the study area and their correlates in the Aranos Basin

4.2 WITVLEI GROUP

Outcrops of the Witvlei Group are known from the study area only. However, it extends from this region further southwards into the Aranos Basin, where it is covered by Nama Group and younger sediments. It is conceivable that the depositional basin of the Witvlei Group is bounded by the Osis Ridge to the south (Fig. 1.1), an east-northeast-striking basement high that featured prominently at the onset of Nama sedimentation (Germs, 1972a) and therefore probably existed during Witvlei times.

4.2.1 Court Formation

This unit outcrops only in the southern and eastern portions of the study area. It is probably absent from the south-western portions of the Witvlei Synclinorium.

4.2.1.1 Tahiti Member

The member comprises massive, fine-grained, medium to brownish-grey arkosic quartzite. The weathered surface of the rock is pitted by cavities 3 to 5 mm across, and is dotted by abundant brownish speckles. In the valley of the Black

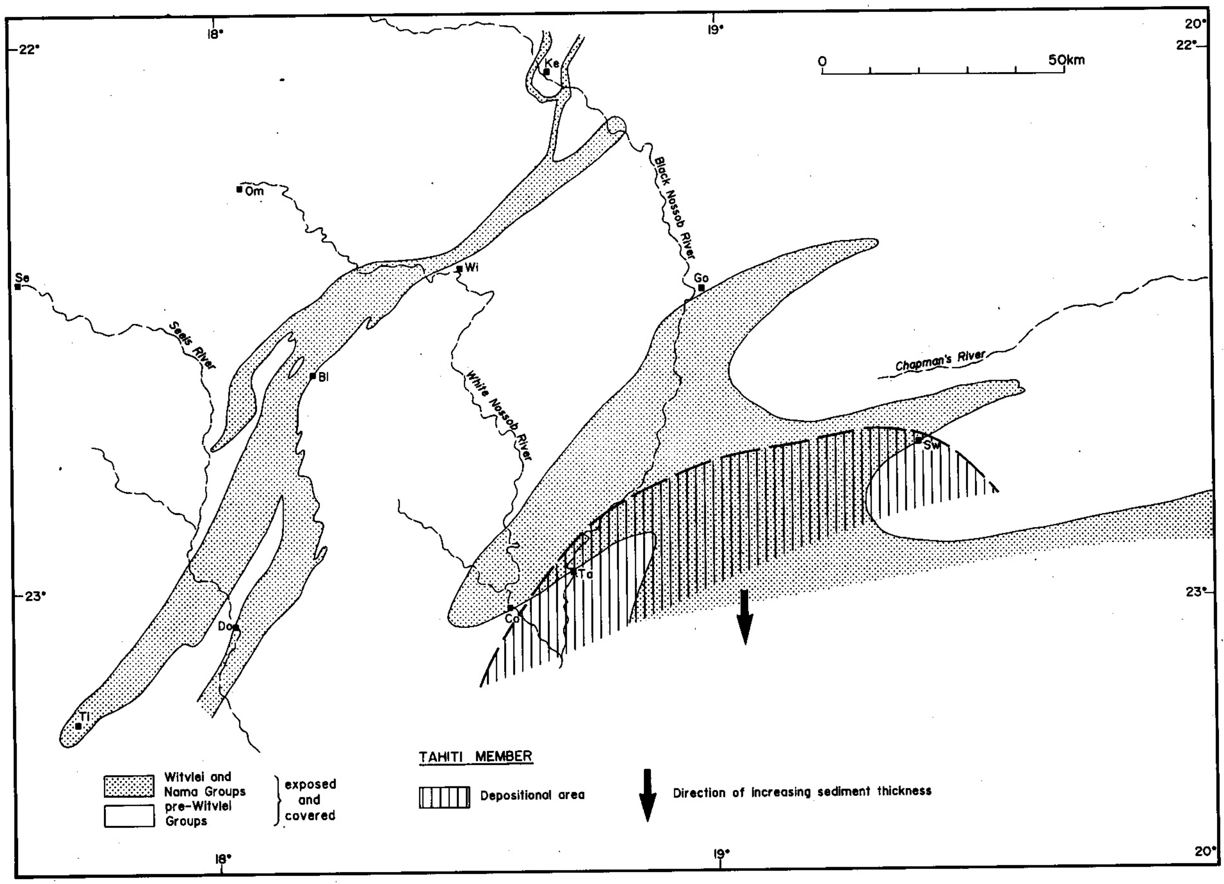


Figure 4.3: Depositional area of the Tahiti Member. For localities see Figure 1.2

Nossob River on the farm Tahiti 61 (2218 DC) this quartzite overlies pebbly siltstone of the uppermost Blaubeker Formation. The thickness is estimated to be between 5 and 10 m since the upper contact of the quartzite is not exposed. A similar quartzite succession was en-countered in the eastern portion of Schwarzwald 344 (2219 CB), where it is well exposed along the path leading from the homestead northwards to the cattle post. About 1 km further west, the Gobabis Member is directly underlain by 1- to 2-m-thick greyish quartzite speckled with reddish-brown streaks.

The inferred initial distribution of the member is shown in Figure 4.3.

4.2.1.2 Gobabis Member

With the exception of a narrow strip along the southern edge of the study area where the Tahiti Member is developed, the Gobabis Member forms the base of the Witvlei Group in most of the Gobabis Synclinorium and in the southeastern portion of the Witvlei Synclinorium (Fig. 4.4). It consists of three distinct facies of medium-grey to black carbonate rocks:

Facies (a) are **laminites** consisting of alternating light-grey to pink and dark carbonate laminae, from a fraction of a millimetre to more than 1 mm thick. This facies, which on the farm Tahiti 61 is up to 25 m thick, seems to have developed only in areas where the entire thickness of the member exceeds 50 m. It is best exposed 1 km south of Gobabis on the eastern side of the Black Nossob River valley, (2218 BD), where it makes up the lower 40 to 45 m of the here 60-m-thick Gobabis Member.

Next to the homestead on the farm Court 32 (2218 BA) the upper 18 m of the member (which here has an overall thickness of 65 m) is only indistinctly laminated and thus transitional to facies (b).

Facies (b) is comprised of **well-bedded to massive carbonate** with scattered intraclasts in some places. On Court 32, facies (b) is 47 m thick. Near Gobabis, this facies overlies facies (a); its uppermost portion includes stromatolitic structures (Plate 4.1). North of the homestead on Josephine 226 (2218 CA) the member is about 40 m thick and consists of massive, dark, algal-laminated carbonate layers which include dome-shaped structures.

Facies (c) is **sandy carbonate to calcareous sandstone** that is exposed along the southern margin of the Netso Syncline (2219 CB and DA) and locally on the southeastern margin of the Witvlei Synclinorium (2218 CC). On the farm Schwarzwald 344 (2219 CB) the sandy carbonate contains coated grains and clasts of pure and sandy carbonate (Plate 4.2). On portion Klippiespan of the farm Marguerite 238 (2218 CC), along the road to Julia 239, the lower portion of the Gobabis Member consists of a sedimentary breccia with clasts of well-layered carbonate (Plate 4.2), pebbles of pre-Witvlei quartzite and ooid grainstone. This can be followed for about 100 m along strike.

The stratigraphic position of a succession of light-grey,

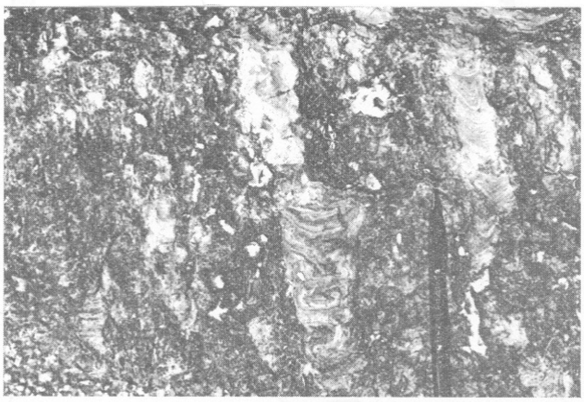


Plate 4.1: Longitudinal section of stromatolites in the uppermost part of the Gobabis Member, facies (b). Columns of light-coloured concave laminae alternate with unlaminated dark columnar zones. Gobabis Townlands. Pen is 12 cm long

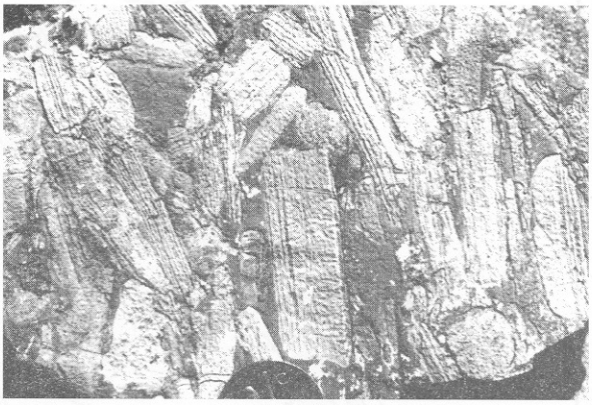


Plate 4.2: Sedimentary breccia in the lower part of the Gobabis Member, facies (b). Clasts consist of indistinctly laminated, light-coloured carbonate and are included in a dark carbonate matrix. Several of the clasts are imbricated and slightly bent to allow closer spacing. Road from Klippiespan (portion of Marguerite 238) to Julia 239. Coin is 1.9 cm across

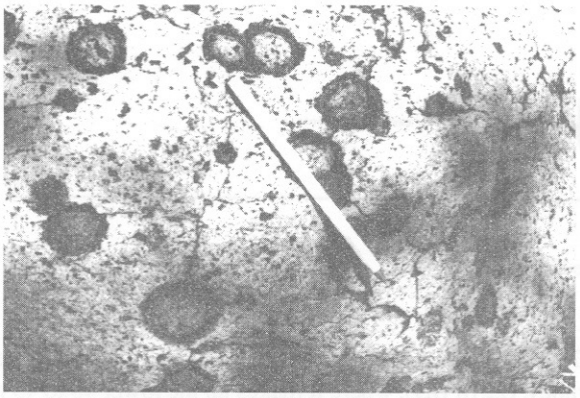


Plate 4.3: Pitted weathering of the Simmenau quartzite; in most cases a sandstone sphere remains inside the cavity. The pitting may be caused by the preferential weathering of concretionary carbonate cement. Poortje 287. Pen is 14 cm long

Plate 4.4: Canyon of the White Nossob River cutting through flat-lying, thickly to massively bedded quartzite of the Simmenau Member. Eilenriede 53—Aais I 31

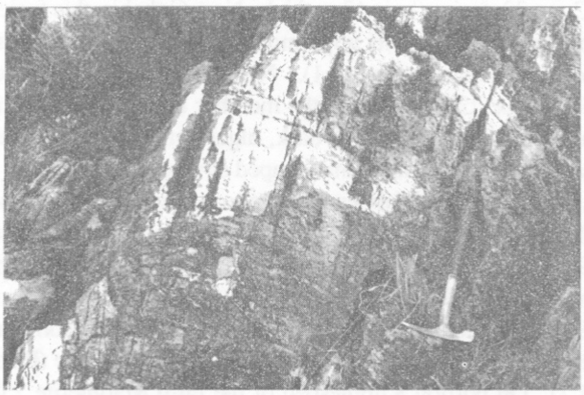


Plate 4.5: Massive, light-grey, cryptalgal laminated dolomite of the Bildah Member. Chert seams indicate layering. Strata form a dome-shaped mound, a few metres across. Southwestern portion of Held 84. Pickhandle is 28 cm long

Plate 4.6: Normally packed oolite in the Bildah Member consisting mainly of multiply coated ooids and subordinately of superficially coated grains. Northern portion of Orochevley 216. Uppermost tabular clast of Eskadron quartzite is 6 cm long

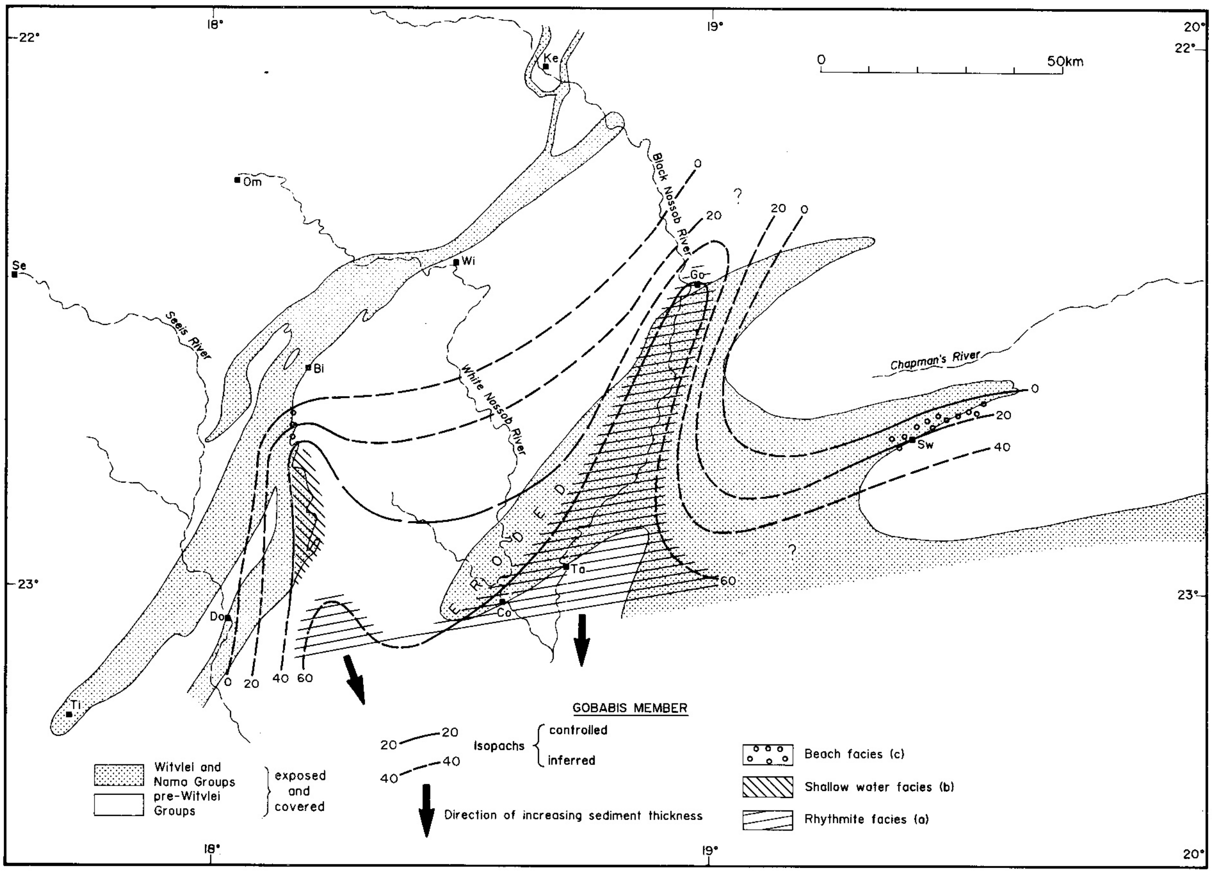


Figure 4.4: Facies distribution and isopachs of the Gobabis Member. Basin opens southward. For localities see Figure 1.2

calcareous sandstone, up to 30 m thick, exposed along the boundary between Josephine 226 and Frank 221 (2218 CA), remains doubtful. It rests on Kamtsas quartzite and is overlain by sandy carbonate of the Bildah Member. It may either belong to the Bildah Member or it could represent facies (c) of the Gobabis Member. Correlation with the Gobabis Member is favoured as it would represent the nearshore facies in a region where the margin of the depositional basin is assumed (Fig. 4.4).

Where the Gobabis Member is overlain by the Constance Member, the boundary between both is poorly defined due to a facies transition. When the latter member is absent, the Gobabis Member is overlain by the Simmenau Member along a sharply defined contact. On the western side of the Gobabis Synclinorium the Gobabis Member was reworked and eroded either entirely (in the southwestern portion) or partly (near Gobabis) during Simmenau times (see Chapter 4.2.1.4).

4.2.1.3 Constance Member

The rocks of the Constance Member occur in an area similar to, but somewhat smaller than that of the underlying Gobabis Member (Fig. 4.5). The best exposures are found along the western bank of the White Nossob River on Court 32 (2318 BA), between the homestead and the northern boundary of that farm. Here the member consists of grey, greenish-grey and yellowish to reddish shale, mudstone and siltstone. These rocks grade into fine-grained sandstone which is most abundant in the upper part of the section. Good outcrops also occur on the eastern bank of this river on Aais 131 (2318 BA). Carbonate layers up to 40 cm thick are present in the lower portion, while in the upper portion sandstone and carbonate is interbedded in the pelitic succession.

A gravel pit on the western bank of the Black Nossob River on the farm Kanabis 54 (2218 DC) exposes 3.5 m of shale and mudstone with interbedded siltstone layers up to 15 cm thick. The succession has a light yellowish-grey colour and good horizontal layering. On the eastern side of that river, near a cattle post on Tahiti 61 (2218 DC), quartzite with several layers of greyish to reddish, locally oolitic, carbonate is interpreted as a transition to the immediately overlying quartzite of the Simmenau Member.

Along the southeastern margin of the Witvlei Synclinorium, between the farms Constance 230 and Doreen 227 (2218 CC), the upper portion of the member consists of varicoloured shale with siltstone and carbonate interbeds. At several localities (e.g. south of the homestead on Constance 230) the top of the member is formed by an indistinctly bedded carbonate zone, about 25 m thick.

Since the fine clastic succession weathers easily, it forms flat relief with only rare outcrops along watercourses. On portion Klippiespan of Marguerite 238 (2218 CC) the total thickness of the member is estimated to be between 100 and 200 m.

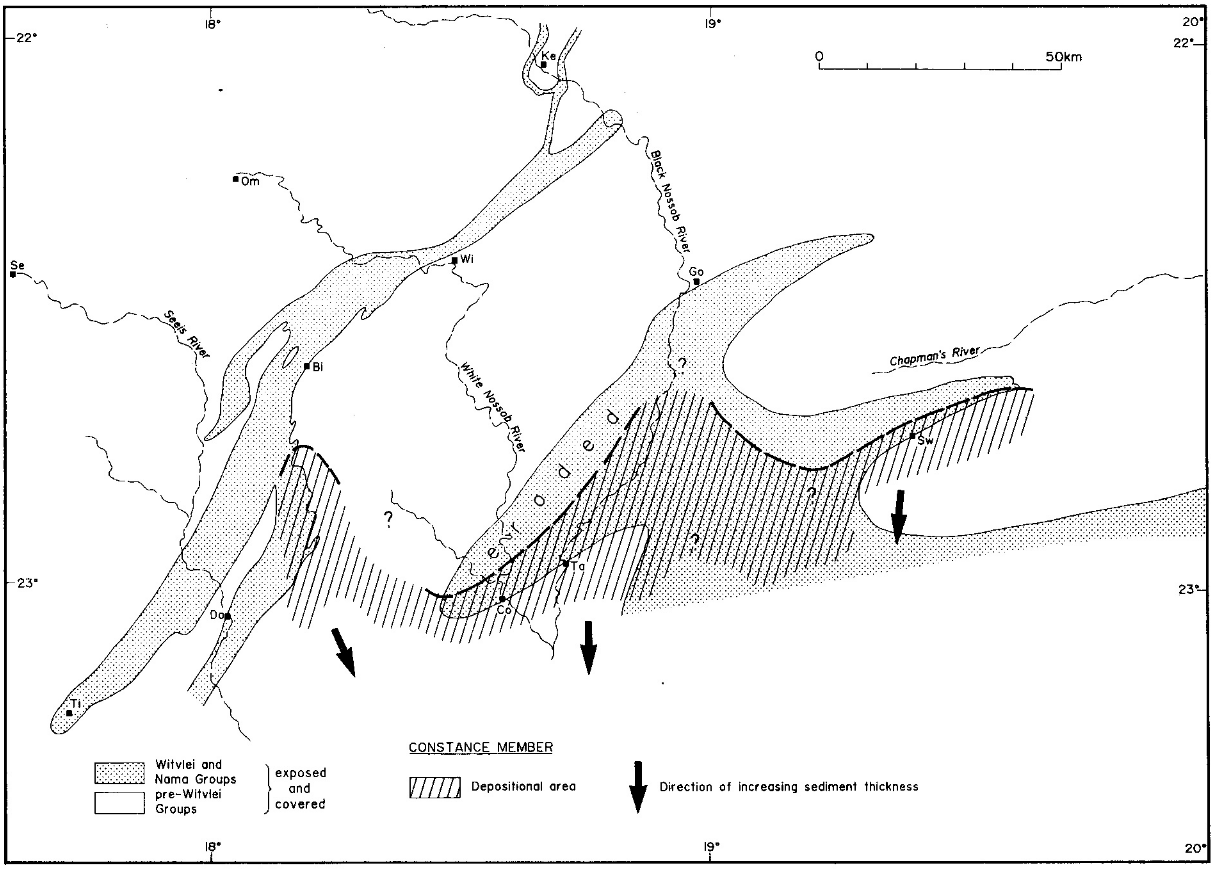


Figure 4.5: Depositional area of the Constance Member. Basin extends southward. For localities see Figure 1.2

4.2.1.4 Simmenau Member

This unit consists mainly of medium- to fine-grained, thickly to massively bedded, feldspathic quartzite. The fresh rock is light to medium grey and weathers brownish-grey. The massive layers are generally several metres thick and a weak stratification occurs locally. Weathering may produce a characteristic surface pitted with abundant cavities and studded with protruding sandstone spheres up to 2 cm across (Plate 4.3). Similar features have also been found in the Weissberg quartzite (Nama Group) on portion Groenrivier of the farm Klein Keizaub 59 (2218 DD) and at a few localities in the pre-Witvlei Kamtsas quartzite to which the Simmenau quartzite may appear strikingly similar.

The member is best exposed between Eilenriede 53 and Kanabis 54 (2218 DC, 2318 BA) in a 2-km-long canyon of the White Nossob River (Plate 4.4). Here the quartzite attains its maximum thickness of 150 to 160 m (Fig. 4.6). From top to bottom the succession consists of:

- 10 m of thinly bedded quartzite with current lineations. The uppermost 0.5 m consists of calcareous, very thinly bedded sandstone, the calcareous content probably being derived from the overlying Bildah Member of the Buschmannsklippe Formation;
- 10 m of thickly bedded (0.5 to 1.5 m) quartzite;
- 50 m of unstratified, massive quartzite;
- 80 m of thickly bedded (0.5 to 2 m) quartzite with interbedded coarse-grained to gritty layers. In some places the bedding planes are densely covered by clay galls which, when weathered away, leave the planes grooved and pitted.

The Simmenau Member forms the outer and stratigraphically lower most of three parallel zones of light-grey quartzite surrounding the Netso Syncline (2219 CB and DA). These three zones give rise to a range of hills which are modified by two deflections developed on the less resistant strata (see below). The Simmenau quartzite contains less silica cement and is slightly darker coloured and more thickly bedded than the other two quartzites (Weissberg and Zenana Members of the Nama Group).

In the western portion of the farm Styria 52 (2218 DC) massive, feldspathic, grey-brown quartzite underlies Bildah rocks; its appearance is very similar to Kamtsas quartzite. Along strike to the northeast, at the Buschmannsklippe hill on Breitenberg 51, the rock contains pebbles of carbonate resembling the Gobabis Member, proving its correlation with the Simmenau Member, in addition to pebbles of vein quartz, quartzite and siltstone.

Further northeast along strike, near the southern boundary of Schönborn (Kaukurus 79F, 2218 DC), the upper part of the member, here about 80 m thick, is composed of typical Simmenau quartzite, while the lower abounds with pebbles and cobbles forming an indistinctly bedded conglomerate. The rounded to subangular clasts, measuring up to 6 cm, consist mainly of quartzite and dark carbonate and are set in a sandy to calcareous matrix. Locally all clasts consist of carbonate rocks and even the matrix is quite pure carbonate. Of special interest are granite clasts (the largest measuring 20 cm across) derived from erosion of the Blaubeker mixtite. Quartzitic clasts are of Doornpoort, Kamtsas and Simmenau type. Northeastward of the southern boundary of Schönborn 79F, the Simmenau Member consists local-

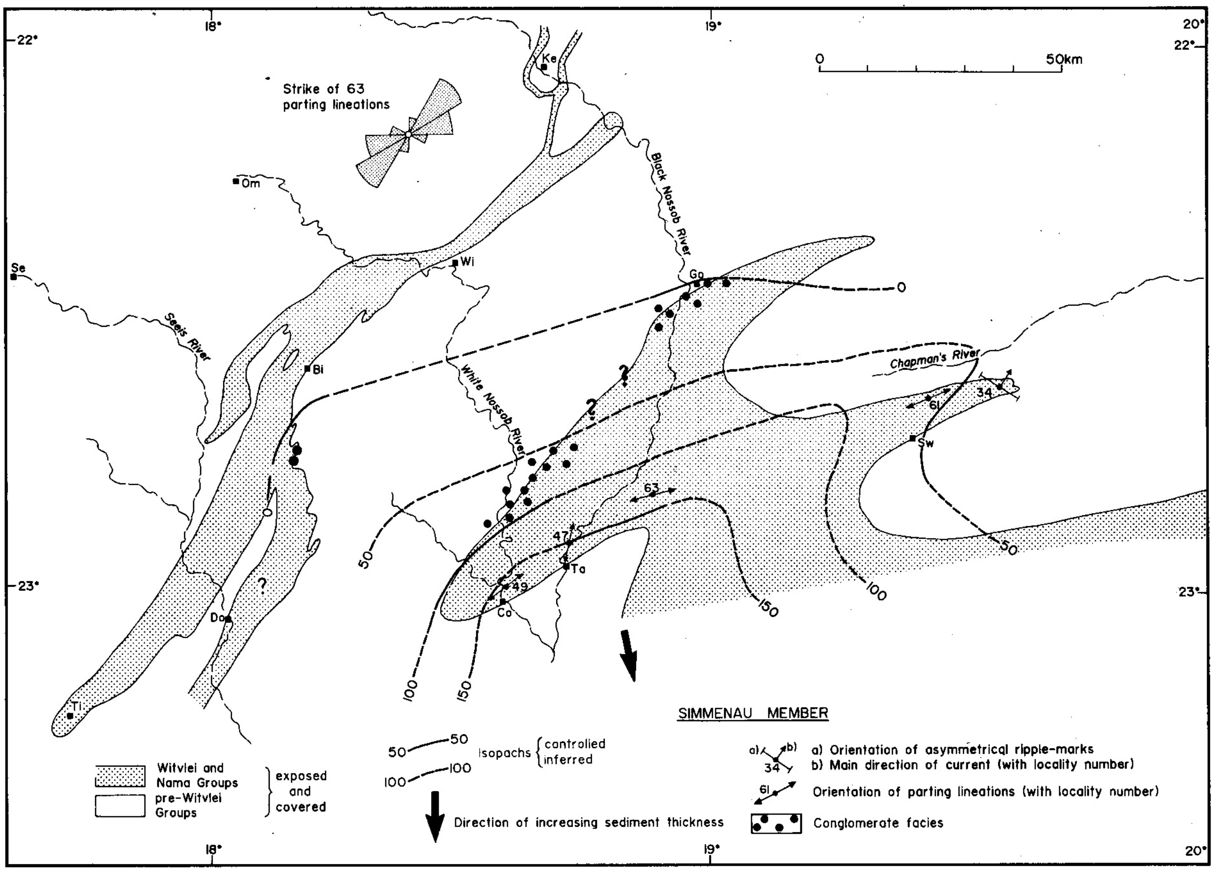


Figure 4.6: Depositional area of the Simmenau Member. Thickness and orientation of sedimentary structures indicate a basin extending to the south. For localities see Figure 1.2. For localities of ripple marks and parting lineations refer to corresponding numbers in Table 6.1

ly of massive dark carbonate, free of clasts but including domal stromatolite structures.

The Simmenau Member appears again 50 km further northeast, at Gobabis, along the northwestern margin of the Gobabis Synclinorium. Here it consists of conglomerate with clasts up to boulder size of dark carbonate and Simmenau-type quartzite (one carbonate boulder measured 70 x 70 x 50 cm) set in a sandy to calcareous matrix. The conglomerate is about 4 m thick, rests on facies (b) of the Gobabis Member (from which all the carbonate clasts were derived), and is overlain by up to 4 m of typical Simmenau quartzite. Locally, however, quartzite forms the base of the member and this is overlain by conglomerate.

From the appearance of the conglomerate on Schönborn 79F and at Gobabis, two conclusions can be drawn, viz:

a) The conglomerate was deposited during upper Simmenau times since it contains clasts of older Simmenau rocks, and is underlain by Simmenau quartzite in places.

b) The carbonate locally included in the Simmenau Member was derived from finely reworked Gobabis Member carbonate.

In the southwestern corner of Frank 221 (2218 CC), 5 to 10 m of calcareous sandstone and sandy carbonate (intercalated between the Kamtsas Formation and the Bildah Member) pass laterally into a conglomerate with pebbles and boulders of dark, sandy carbonate and calcareous sandstone in carbonate matrix. This sequence could therefore be considered a marginal facies of the Gobabis Member or a correlate of the Simmenau conglomerate.

4.2.2 Buschmannsklippe Formation

The depositional area of this formation was considerably larger than that of the Court Formation, and in the north, west and south extended beyond the limits of the study area (Fig. 4.2), transgressing over the rocks of the Court Formation onto the pre-Witvlei basement. To the east, however, the occurrence of the Buschmannsklippe succession is more restricted than that of the Court Formation.

The Buschmannsklippe Formation is most typically developed south of Witvlei in area 2218 CA and in the northern portion of area 2218 CC, the best section being found in a ravine immediately west of the homestead on Okambara 219. Due to insufficient data, no isopachs can be constructed in the southwestern portion of the Witvlei Synclinorium, i.e. the Rooiwater Syncline, but all members of the formation occur there with the exception of the Bildah Member which, however, might be covered.

4.2.2.1 Bildah Member

Having been deposited during a major transgression, this unit forms the base of the Witvlei Group in areas outside the depositional area of the Court Formation, i.e. in the central and northern portions of the Witvlei Synclinorium. Although sedimentation took place in an expanding basin, the Bildah Member generally does not start with a basal conglomerate or scree. Where typically developed, the member consists of light-grey and pink dolomite that was

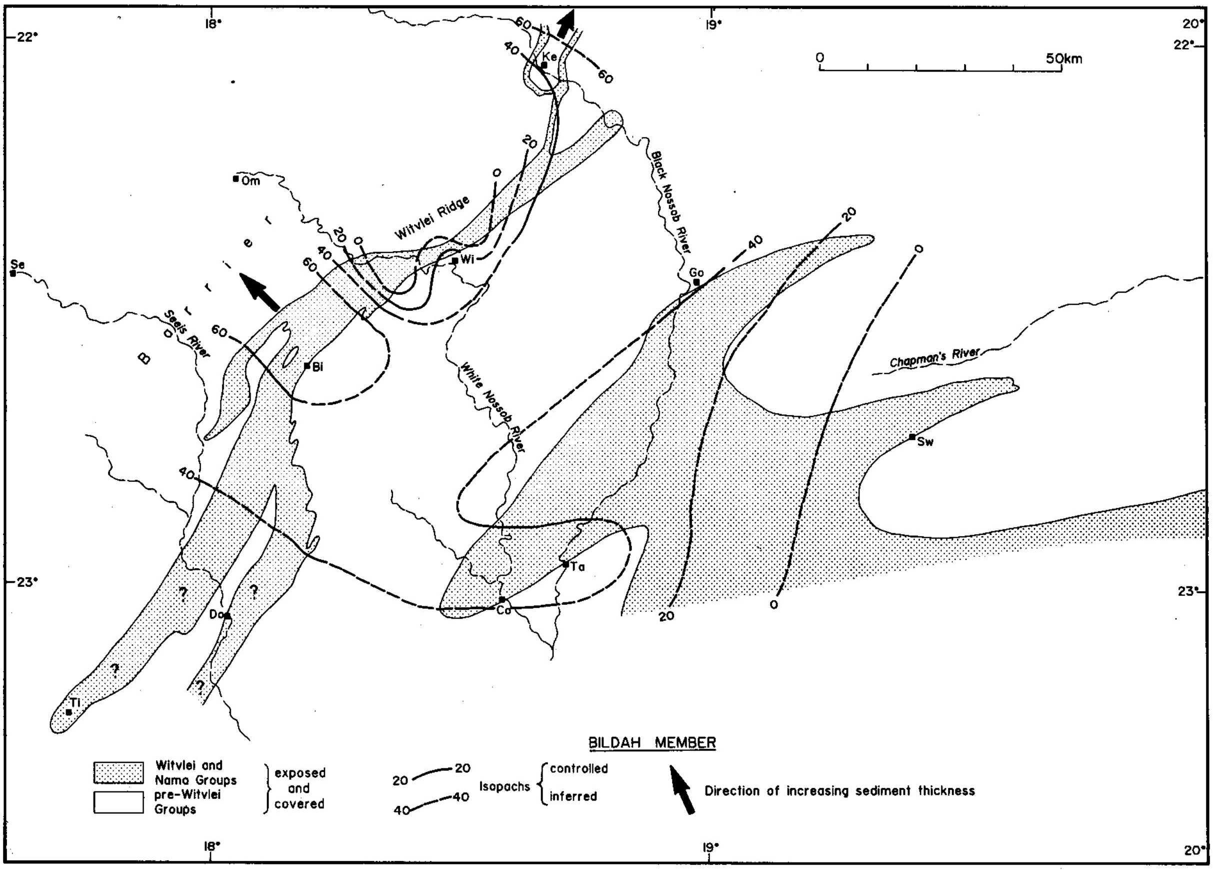


Figure 4.7: Bildah Member. Basin extends to the west and north. For localities see Figure 1.2

deposited on an even floor under very shallow water with sporadic subaerial exposure.

The maximum thickness of about 60 m has been measured between the farms Bildah 220 and Okombuka 218 (2218 CA), but it might have been thicker further west (Fig. 4.7). A similar thickness has been found on Kehoro N 185 (2218 BA). Between the above two regions the thickness is reduced on the Witvlei Ridge and locally the member pinches out (Fig. 4.7).

The dolomite is thickly bedded to massive, the layering often indicated by the position of thin chert seams. Only in the upper 10 m of the succession is bedding distinct and individual layers are thin, while the pink colour deepens and the rock forms a transition into the overlying La Fraque Member. A characteristic feature of the Bildah dolomite is its cryptalgal lamination, manifested by slight colour differences between adjoining laminae, and selective weathering of certain laminae, leading to undulating algal layers and domal stromatolites up to 3 m across and 1 m high (Plate 4.5).

Common features in the algal-laminated Bildah carbonate are tubular, approximately vertically disposed structures consisting of coarse crystalline calcite or chert and measuring 0.5 to 2 cm across. They either form small depressions on bedding planes or stand out crater-like. These structures are dealt with in more detail in Chapter 6.2.1.

Two kilometres south of the northern corner beacon on the farm Orochevley 216 (2218 CA) a synsedimentary erosion channel is cut into Bildah carbonate and locally even into the underlying pre-Witvlei quartzite. It was filled by unbedded oolite including small fragments of quartzite, slate and carbonate. The ooids are up to 2 mm across, or if elongated, up to 4 mm long (Plate 4.6). The channel can be traced for about 150 m. Its oolite filling is up to 10 m thick and is overlain by the common dolomite of this member.

Approaching the coastline or shoals, the cryptalagal laminite interfingers with and is progressively replaced by clastic deposits derived from the mainland and the underlying strata as the following examples indicate:

1. Southwest of the homestead on Okombuka 218 (2218 CA) a small tectonic sliver, occurring between the two (tectonically duplicated) outcrops of algal laminated Bildah Member, consists in the lower few metres of indistinctly bedded layers, 0.2 to 0.8 m thick, of light-grey to brownish sandy dolomite, which contains well-rounded and subrounded tabular pebbles of Eskadron (pre-Witvlei) quartzite. Desiccation polygons and spelaeothems are common. The succession rests on Eskadron conglomerate and quartzite. Higher up the dolomite is either devoid of pebbles or has a few pebbly layers that consist of small, rounded to angular clasts, up to 5 mm across, most of which are coated by carbonate and set in a carbonatic to sandy matrix. The minimum thickness of this zone is 10 m.

2. Northeast of Weshof 585 (2218 CB) the thickness of the member decreases and intercalated sandy material becomes progressively more abundant. On the farm Zenana 100 (2218 CB) arenaceous carbonate, including intraclasts, is intercalated. On the western portion of Held 84 (2218 AD) the middle Bildah Member consists of alternating pure and

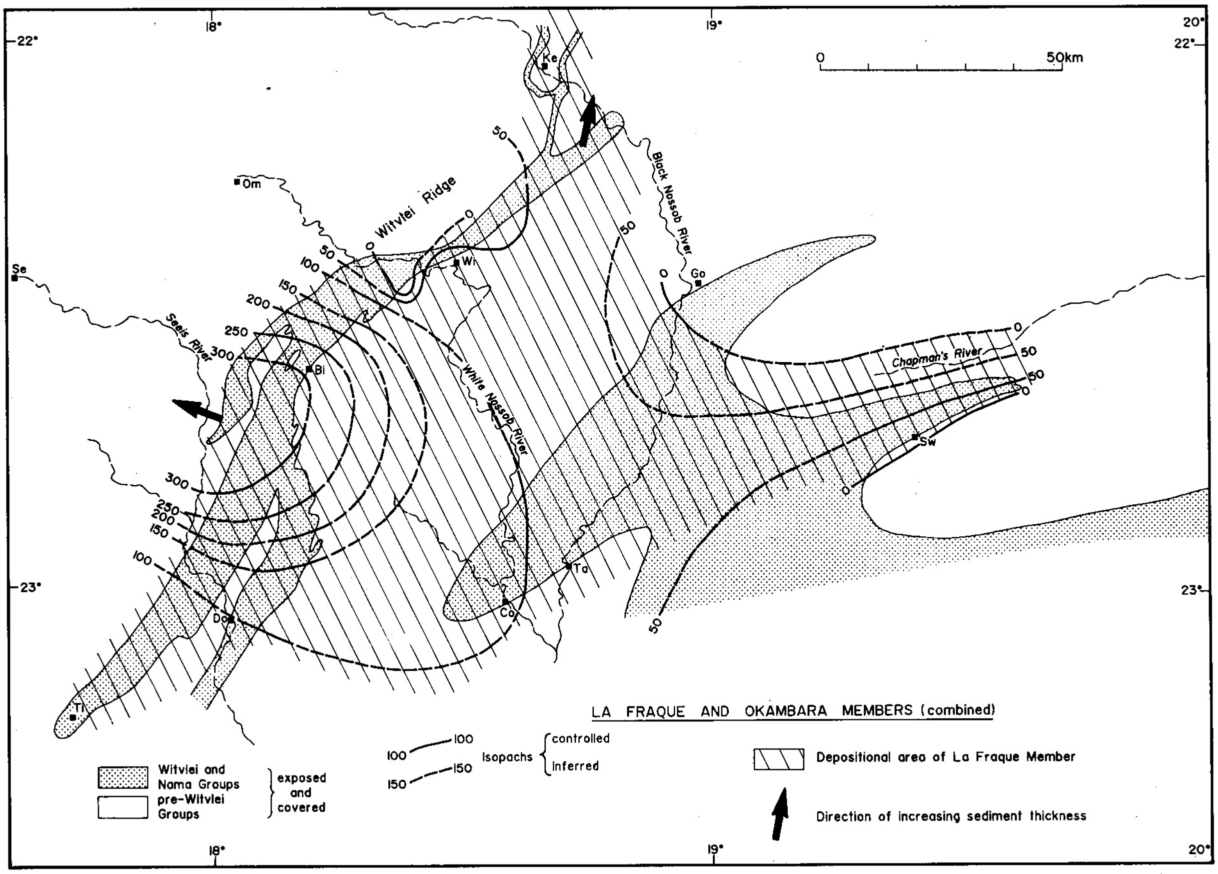


Figure 4.8: Depositional area of the La Fraque Member (indicated by hatching). Isopachs show combined thickness of the La Fraque and Okambara Members. Basin extends to the west and north. For localities see Figure 1.2

sandy carbonate. Further northeast the carbonate content decreases and the member is represented by some 5 m of medium-grained, pebbly sandstone which peters out toward the northeast, in the central portion of the farm. Clastic and rudaceous intercalations attesting to the vicinity of former coastlines extend up to 4 km into the basin.

3. At Witvlei (2218 AD), outcrops of pure as well as sandy carbonate rocks reappear, but in the northeastern portion of Okatjirute 155 (2218 AD), the Bildah Member is absent again and from here on northeastwards, over a distance of some 40 km, only sporadic outcrops of Bildah rocks occur. Their presence below a thin cover of Kalahari deposits is indicated by a line of large *Acacia eriloba* trees, a species that grows profusely on the weathered contact zones of the member.

4. In the Kehoro area (2218 BA) scattered outcrops of typical Bildah dolomite occur (plate 3.3). Nearshore deposits similar to those on Held 84, described above, form an isolated outcrop on the southern bank of the Black Nossob River on the farm Sachsenwald 940 (2218 BA). Here the carbonate rocks contain fine fragments of red Eskadron quartzite, many of them coated, and of basic lava.

5. Calcareous sandstone in the region of the boundary between the farms Josephine 226 and Frank 221 (2218 CA) passes upward into algal laminated carbonate of the Bildah Member, the lower portion of which is slightly sandy and medium grey in colour. Those rocks can be correlated either with the Gobabis and/or Simmenau Members or with the Bildah Member. The correlation with the Gobabis Member, however, is favoured (see Chapter 4.2.1.2).

In the western portion of the Gobabis Synclinorium the Bildah Member typically is developed, but due to the flatness of the landscape it is largely covered by surface limestone and exposures occur only in the valleys of the Black and White Nossob Rivers.

A few boulders of white marble occur scattered within a five-square-metre area about 200 m northeast of 'he homestead on Okasewa NW 120 (2218 AD), within the fault zone separating Nama rocks (Zaris Formation) from Eskadron quartzite. The marble is interpreted as a tectonic sliver which, because of its light colour, is probably recrystallised Bildah dolomite.

4.2.2.2 La Fraque Member

This succession mainly comprises thinly bedded to laminated calcareous siltstone and fine-grained sandstone with a number of intercalated carbonate layers (plate 4.7). The silty rocks are reddish brown, becoming greenish and greenish brown upwards. The La Fraque Member is best developed in the central portion of the Witvlei Synclinorium (2218 CA, Fig. 4.8). As the strata are less weather-resistant than the rocks of the overlying and underlying units, they form a gentle slope between two scarps.

The uppermost portion of the Bildah Member becomes increasingly pelitic towards the top; the lower boundary of the overlying La Fraque succession is drawn where shaly siltstone predominates over calcareous rocks. This boundary is quite distinct in outcrop. The upper boundary of the La Fraque Member, however, is indistinct, the main difference between the La Fraque and the overlying Okambara successions being the" greater thickness and compactness of the carbonate beds in the Okambara Member. While this change is not confined to the same stratigraphic level throughout the area, recognition of the upper boundary of the La Fraque Member is aggravated by a cover of scree derived from the overlying Okambara rocks.

In an anticline extending over portions of the farms Okombuka 218 and Okambara 219 (2218 CA, Plate 3.2) massive, light-grey to reddish-grey carbonate is intercalated in the lower portion of the La Fraque Member. Another outcrop of the La Fraque succession forms a cliff southwest of the homestead on Bildah 220 in which a 20-m-thick zone of fine-grained, thin-layered quartzite in its upper portion contains several 20-cm-thick carbonate beds with lenses of flat-pebble conglomerate.

Near the southern boundary of Doreen 227 (2218 CC), west of the main road, a 15-m-thick zone of conglomerate and sedimentary breccia, beginning 36 m above the base of the La Fraque Member, can be followed for more than a kilometre along strike in several hillocks. Clasts consist of carbonate, siltstone and fine-grained sandstone. These are set in a fine-sandy carbonate matrix. Associated with this breccia are boulders derived from a 15-cm-thick carbonate layer which contain large carbonate crystals (calcite or aragonite) in clusters and palisade-like arrangements up to 6 cm high.

A borehole near the eastern boundary of portion Fraaihoek of the farm Graswêreld 272 (2317 BB) penetrated red-brown, slightly calcareous siltstone attributed to the La Fraque Member, which is not exposed in the southern portion of the Witvlei Synclinorium, but must exceed 50 m in thickness here.

In the area between Bildah 220 and Okombuka 218 (2218 CA), the La Fraque Member reaches its maximum thickness (between 120 and 160 m). From there northeastward it becomes thinner. On Weshof 585 (2218 CB) it measures 70 m, in the western portion of Held 84 (2218 AD) 50 m, and eventually it pinches out against the Witvlei Ridge (Fig. 4.8). Simultaneously the proportion of the calcareous component gradually decreases and finally the member consists of fine detrital material only; it cannot be distinguished from the Okambara Member at this locality. Near Witvlei the member reappears, but is again absent in the eastern half of Okatjerute 155 (2218 BD) where the Weissberg Member directly overlies Kamtsas Formation of pre-Witvlei age. Further to the northeast the La Fraque Member is not exposed for about 40 km as far as the closure of the Eindpaal Syncline (Fig. 1.2). However, boreholes on Lee Enfield 303 (2218 BC) and Mahagi 302 (2218 BB) penetrated red shale and siltstone, and it is believed that the La Fraque Member underlies a depression between the Weissberg quartzite and a strip of large camelthorn trees following the Bildah Member. In the Kehoro area (2218 BA) a few outcrops of reddish marl and siltstone occur, but most of the member is covered by calcrete and sand.

In the Gobabis Synclinorium the La Fraque Member is absent in the north at Gobabis, but occurs along the southwestern margin where it reaches about 50 m in thickness on Schönborn 79F (2218 DC) and is locally exposed between Eilenriede 53 (2218 DC) and Klein Keitsaub 59 (2218 DD). Here only flakes of reddish mudstone, siltstone and subordinate marl are found in calcrete and burrows. On KudÜberg 60 (2218 DD), the thickness of the member is estimated to be about 60 m. A separation from the overlying Okambara Member is difficult in the Gobabis Synclinorium. Around the Netso Syncline (2219 CB and DA) a prominent bench (mostly covered by calcrete and loam) lies between the quartzite ridges of the Simmenau and Weissberg Members. Two borrow pits, one on the farm Leeupoort 595 on the eastern side of the main road, and the other near the homestead on Poortjie 287, reveal reddish, slightly calcareous mudstone. The whole succession between the two quartzitic sequences is about 70 m thick on the northern side of the Netso Syncline, but only 10 to 20 m thick on the southern limb (Fig. 4.8). This succession is considered to represent the La Fraque Member which is the only unit of the Buschmannsklippe Formation developed along the Netso Syncline.

4.2.2.3 Okambara Member

Like the underlying La Fraque Member, this member is most typically developed and best exposed in the central portion of the Witvlei Synclinorium (2218 CA) where it consists of dolomite and subordinate limestone that form layers generally between 0.1 and 1.0 m thick. Fine-sandy to silty intercalations, with rapidly changing ratios of carbonate and arenaceous components, occur throughout the member, but abound in the middle portion. In area 2218 CA the thickness of the member is 125 m (Fig. 4.9). The following section, from top to bottom, was compiled from several outcrops in this area:

Top: Quartzite (lower most portion of the Weissberg Member);

11) 4 m of alternating strongly crossbedded quartzite and dolomite layers (Plate 4.7) with lenses of flat-pebble conglomerate;

10) 6 to 12 m of dolomite with one or more stromatolite layers (Plate 4.8), intercalated flat-pebble conglomerate; chert present in layers or as spherical nodules (originally displacive evaporite mineral); light-grey carbonate nodules;

9) 7 m of light-grey to brownish dolomite, generally thick-bedded, local flat-pebble conglomerate;

8) 8 m of quartzite and crossbedded sandy dolomite; alternating quartzite and carbonate in lower portion;

7) 30 m of light-coloured, thick-bedded dolomite with brown quartzite interbeds, flat-pebble conglomerate with dark-weathering arenaceous dolomite matrix and pebbles of pure, light-coloured dolomite; large stromatolites are present (Plate 4.9);

6) 20 to 28 m of fine-grained, red-brown, quartzite, thinly bedded, parallel and cross-laminated, with hummocky cross-stratification, mud galls; subordinate sandy carbonate with flat-pebble conglomerate;

5) 25 to 30 m of greyish-pink, thick-bedded to laminated limestone, which consists to a large extent of crinkled algal mats (Plate 4.10); characteristic zones of edgewise conglomerate which laterally may pass into flat-pebble

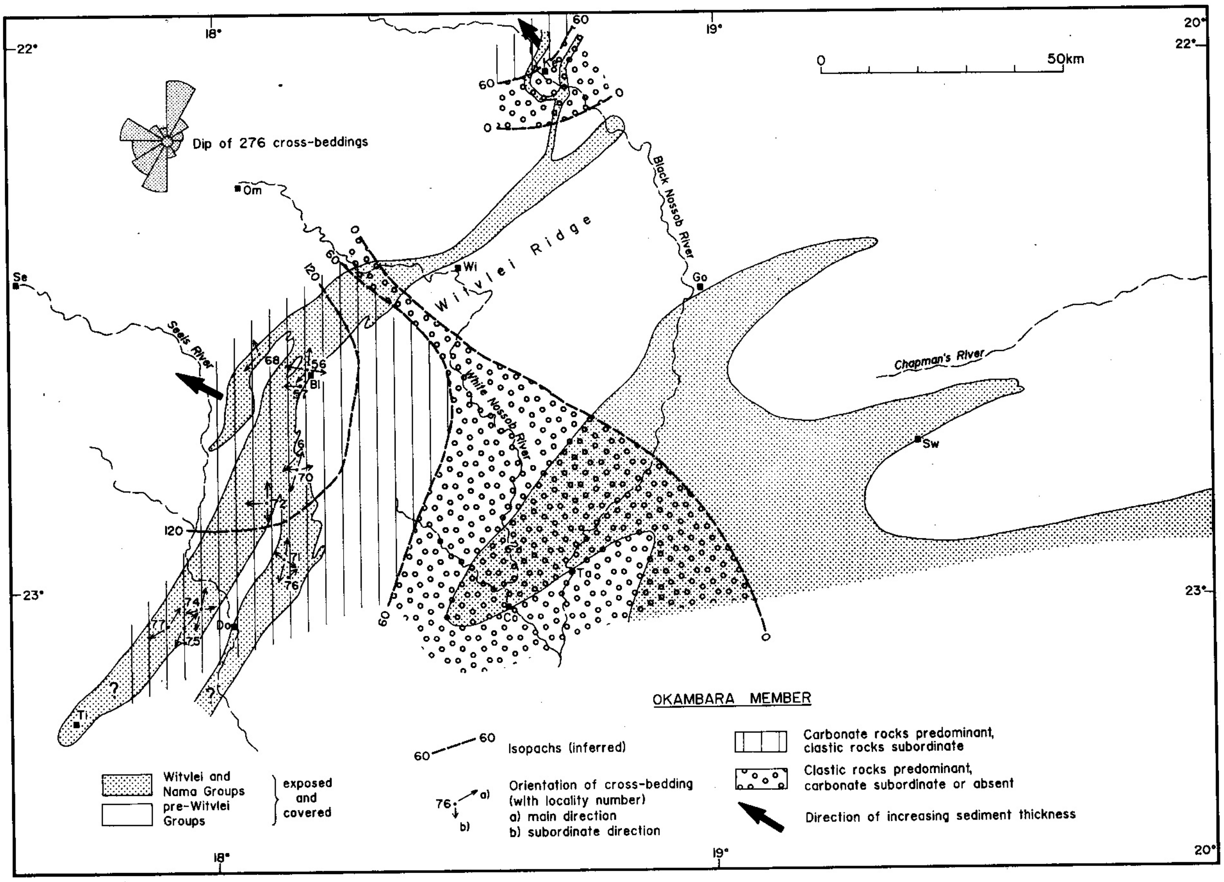


Figure 4.9: Depositional area of the Okambara Member showing facies distribution, approximate thickness and orientation of crossbedding. For localities see Figure 1.2; for localities of crossbedding refer to corresponding numbers in Table 6.1

conglomerate. Subordinate fine-grained quartzite layers are intercalated; erosion troughs in the quartzite are filled with carbonate rock. At some localities ripple marks occur on the bedding planes of the quartzite which has hummocky cross-stratification;

4) 1.3 m of fine- to medium-grained, feldspathic quartzite; beds in upper portion about 25 cm thick, those in lower part 1 to 5 cm, laminated and hummocky cross-stratified; local sandy carbonate with flat-pebble conglomerate;

3) 2 m of dark red-brown shale and siltstone with shallow ripple marks;

2) 8 m of reddish-grey to red-brown, laminated limestone containing edgewise conglomerate with remarkably thin clasts (Plate 4.11); subordinate layers of shale and siltstone;

1) Approximately 6 m of red-grey limestone layers between 10 and 20 cm thick; intercalated beds of red-brown marl, shale and siltstone from 5 to 30 cm thick.

The succession can be subdivided into three units:

1) a lower limestone-dominated unit, about 45 m thick (layers 1 to 5);

2) a middle, quartzitic unit, approximately 25 m thick (layer 6);

3) an upper dolomite-dominated unit, its arenaceous content higher than that of the lower unit; 60 m thick (layers 7 to 11).

The lower end of the section seems to coincide with the boundary between the Okambara and La Fraque Members; it is best exposed on the northern tip of a hill 4 km southsouthwest of the homestead on Bildah 220 (2218 CA). The thick middle quartzite unit (no. 6 of section) is a weather-resistant marker band which forms a ledge and is recognised on aerial photos by its dark colour (Plate 3.2). This quartzite becomes very prominent in areas 2218 CC and 2318 AA, whereas the thickness of both underlying and overlying carbonate units decreases. In the region of the farms Constance 230 and Doreen 227 (2218 CC) the quartzite is 50 m thick, the lower 38 m of which are typically red-brown, whereas the upper 12 m are light coloured if fresh and resemble the Weissberg quartzite, but they weather to a brownish colour. In this region the carbonate succession overlying the quartzite is 25 m thick and is devoid of stromatolites. Further south, on portion Keerweeder of Achenib 247 (2318 AA) the quartzite unit reaches 40 m in thickness and is separated by 25 to 30 m of poorly exposed carbonate from the Weissberg quartzite. A comparison of this region with the Rooiwater Syncline in area 2317 BB indicates that the carbonate thickens and the quartzite becomes thinner in a northwesterly direction.

The stromatolite-bearing layer in the upper unit (layer 10 of section) forms a narrow, southwest-northeast-striking zone which is continuous over a distance of at least 35 km, between the Weissberg corner beacon (2218 CA) in the northeast and the southeastern corner of Otjimbondona 225 (2218 CC). Its extent coincides with the optimal development of carbonate facies of the upper unit of the Okambara Member. The southernmost occurrence of the stromatolitic marker is an isolated outcrop near the north-

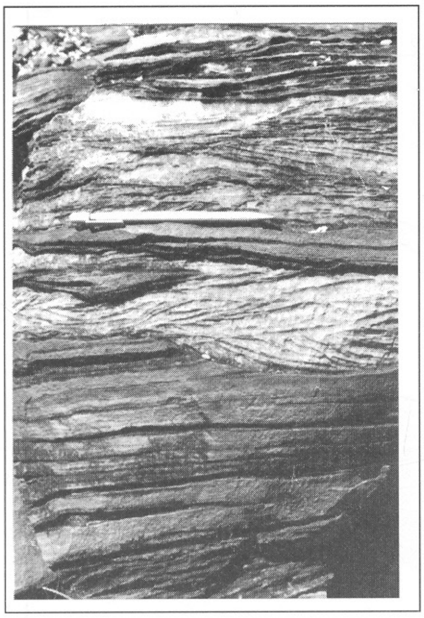


Plate 4.7 (left): Dolomite and sandy dolomite in the uppermost part of the Okambara Member; strongly reworked, channel-fills, crossbedded. Okambara 219. Pencil is 14 cm long. *Photograph:* K.H. Hoffmann

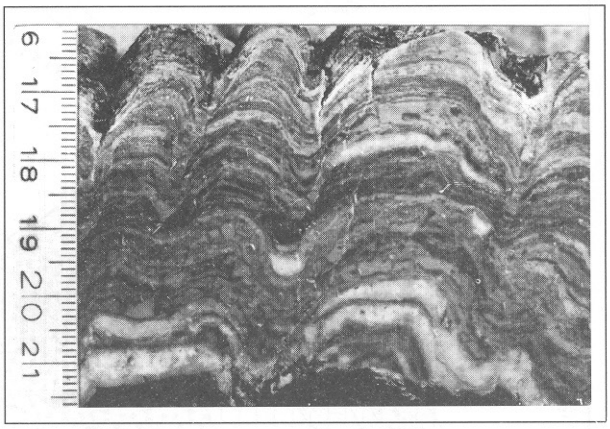


Plate 4.8: Polished section through a stromatolite (LLH-C) from the upper unit of the Okambara Member. Dolomite laminae are light brown and chert layers are white; desiccation and incipient reworking in lower half. Okambara 219. Scale in centimetres

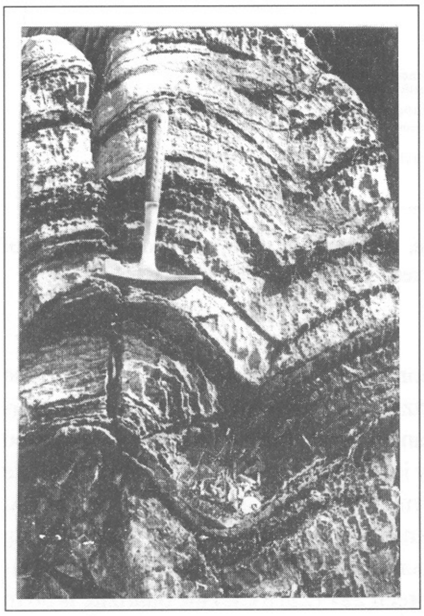


Plate 4.9: Large algal domes in the upper unit of the Okambara Member. They are separated from each other by an abrupt change in the inclination of algal layers. Doreen 227, near the boundary with Otjombondona 225. Pickhandle is 28 cm long

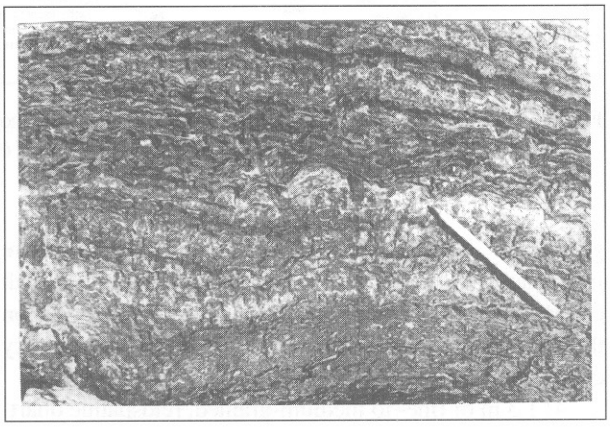


Plate 4.10: Crinkled to corrugated algal mats in the lower unit of the Okambara Member; laminae of light-coloured pure limestone and darker fine-sandy limestone; some chert layers in upper portion. Algal mats locally form domal stromatolites (LLH-C). Okambara 219. Pen is 14 cm long

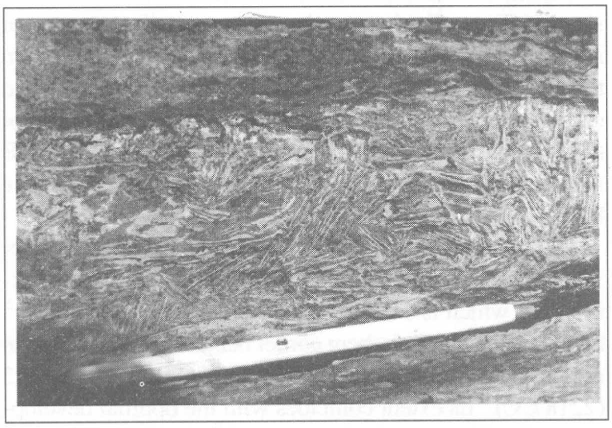


Plate 4.11: Layer of edgwise conglomerate with scoured lower surface in the lower unit of the Okambara Member. Okambara 219. Pen is 14 cm long

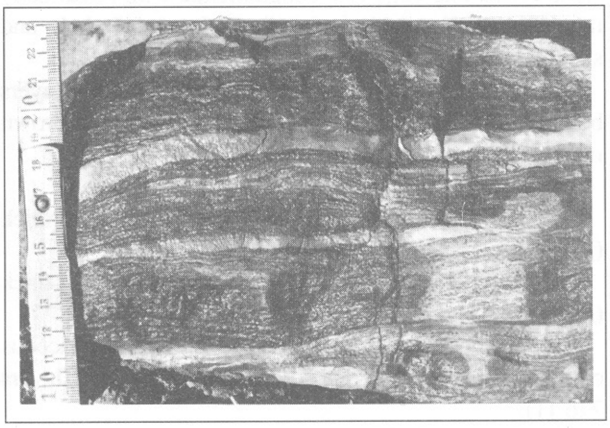


Plate 4.12: Dark carbonate of the Omkyk Member with thin light-coloured layers and laminae. Rooiwater S 274. Scale in centimetres

eastern corner of Kowas 233 (2218 CC), another 18 km to the southwest. Stromatolites are well exposed west of the homestead on Okambara 219 and on the northern side of the main road on the boundary between the farms Doreen 227 and Otjimbondona 225 (2218 CC).

At the latter locality large dome-shaped algal laminated carbonate build-ups, up to 10 m across and about 1.5 m high, occur in the lower 15 m of the upper unit (layer 7 of section, Plate 4.9). Similar structures, however, were encountered in the lower unit of the member west of the homestead on Josephine 226 and on Bildah 220 (2218 CA).

The uppermost part of the Okambara Member (layer 11) and the lower most few metres of the Weissberg Member are characterised by quartzite with herringbone cross-stratification, oscillation ripples, mud-pebble conglomerate and desiccation cracks. This zone, which is mostly covered by scree, is best exposed north of the main road on the boundary between Doreen 227 and Otjimbondona 225 (2218 CC). It seems to form a transition zone between the two members.

Approaching the Witvlei Ridge, the carbonate content of the Okambara Member decreases from southwest to northeast in the western portion of Held 84 (2218 AD), just as is the case with other members of the Buschmannsklippe Formation. Like these, the Okambara Member is finally represented only by sandstone which retains its characteristic red-brown colour, fine grain size, good layering and crossbedding. It contains some ripple marks and mud-pebble conglomerates. Conglomerate in the upper portion consists of pebbles and boulders up to 25 x 15 x 10 cm all set in a sandy matrix. Finally the member becomes indistinguishable from the La Fraque Member and both pinch out towards the Witvlei Ridge. Here an angular inconformity appears to separate these units from the overlying Weissberg Member.

Northeast of Witvlei, the Okambara Member is either absent or consists of a thin arenaceous succession which, together with the La Fraque Member, underlies a depression bordering the Weissberg quartzite. In the region of the common corner of Nudom 161, Kehoro S 939 and Kehoro 183 (2218 BA, Plate 3.3) it consists of red-brown, fine-grained quartzite. Three kilometres northeast of the homestead on Kehoro N 185 the member is represented by scattered outcrops of reddish to grey-red, algal-laminated, often sandy dolomite with intercalated red-brown, crossbedded sandstone that forms layers up to 50 cm thick. Poorly developed domal stromatolites have been observed.

In the Gobabis Synclinorium the Okambara Member is restricted to the southwestern portion. Here a predominantly arenaceous facies, devoid of or very poor in carbonate intercalations, is exposed. The thickness of the Okambara succession is greatly reduced, compared with that at the type locality (Fig. 4.9). It is about 60 m on KudÜberg 60 (2218 DD) and about 50 m on Schönborn 79F (2218 DC). The predominantly fine sediments of the Okambara and La Fraque Members are lithologically similar in the Gobabis Synclinorium.

4.3 NAMA GROUP

4.3.1 Dabis Formation

In the course of the expansion of the depositional basin the Witvlei Ridge was submerged and subsequently covered with sediments. The depocentre for the lower Nama Group is believed to have been situated to the west of the study area.

4.3.1.1 Weissberg Member

This member consists of dense, light-grey quartzite, a very characteristic, weather-resistant rock that stands out from the surrounding country in the form of steep-sided hills (Plates 3.1 and 3.2).

Three lithofacies of the Weissberg Member can be distinguished:

Facies (a) is a fine-grained, light-grey orthoquartzite which is the prevalent lithotype of the member in the southern half of the Witvlei Synclinorium, south of the Witvlei Ridge (Fig. 4.10). The rock is distinctly stratified, individual layers being between 10 and 80 cm thick. Many bedding planes have symmetrical ripple marks.

Quartzite with sedimentary structures of the same type as that intercalated in the uppermost Okambara carbonates also forms the lowest 3 to 5 m of the Weissberg succession. It is intensely cross bedded, in many cases displaying herringbone textures, contains mud-gall conglomerates (which, when weathered away, leave a deeply pitted surface), and on bedding planes, desiccation cracks occur.

Grain size and maturity of this lower portion are the same as in the higher portions of the member, but the rock is more intensely weathered near the contact with the underlying carbonate strata of the Okambara Member and has a darker and pitted appearance, while the upper portion of the Weissberg Member is so intensely silicified that the outlines of the quartz grains are only distinguishable under the microscope.

A maximum thickness of 100 to 115 m was measured on the farm Scheidthof 293 (2218 CC). From there the thickness decreases toward the south and northeast (no rocks of this member are exposed west of Scheidthof), being 70 m on portion Keerweeder of Achenib 247 (2318 AA), 25 m in the northwestern corner of Bildah 220 and 16 m on Weshof 585 (2218 CB).

Facies (b), is a **subarkose**, distinguished from the former rock type mainly by its feldspar content and its generally coarser grain size. It is medium grained and contains patches of coarse to gritty sand. Grains are up to 5 mm across and consist of milky or clear quartz, feldspar and subordinate chert. The coarser material was concentrated on the surface or in the topmost portion of individual layers. Ripple marks are scarce and clay pellets occur locally. Grain size increases from the Witvlei Ridge, where the member is up to about 20 m in thickness, to the northeast

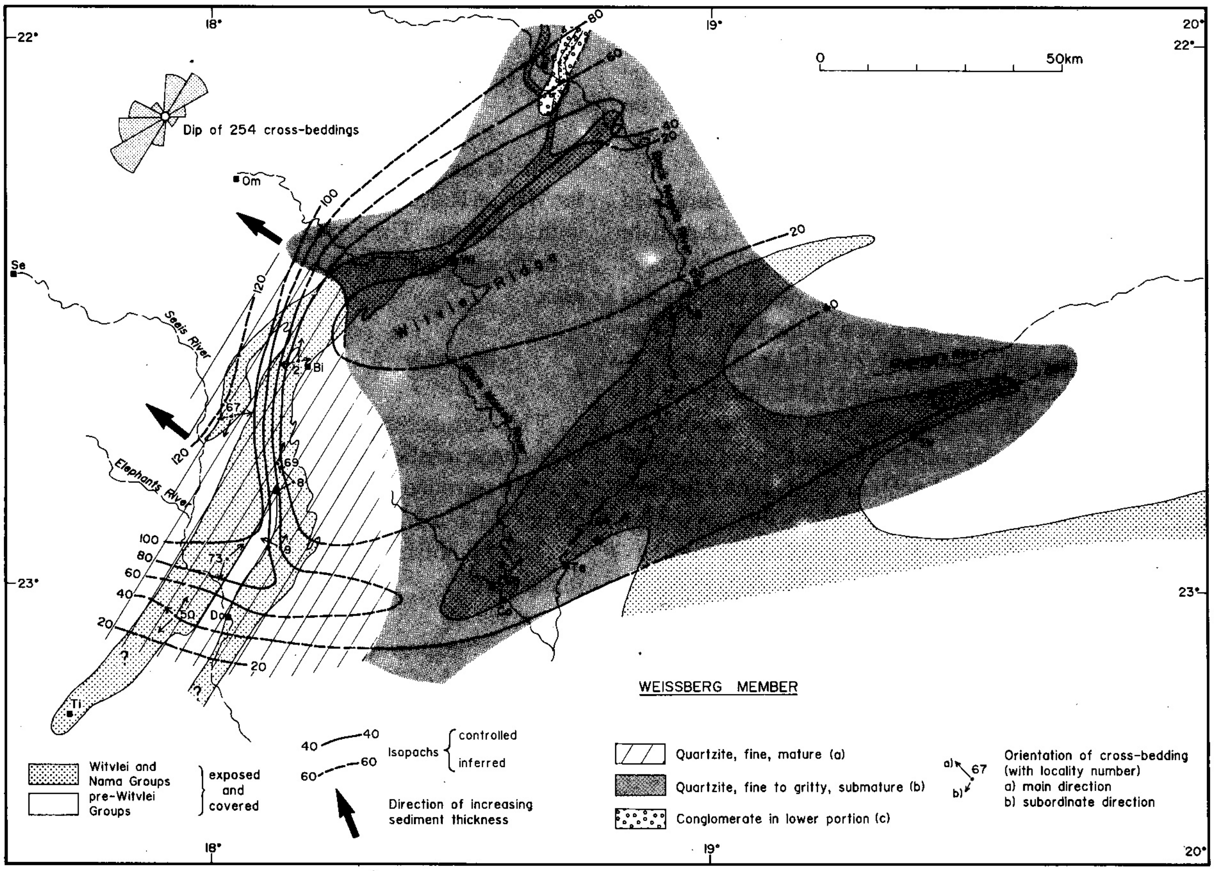


Figure 4.10: Weissberg member. Facies distribution, thickness and orientation of crossbedding are indicated. For localities see Figure 1.2; for localities of crossbedding refer to corresponding numbers in Table 6.1

and around Kehoro 183 (2218 BA) quartz pebbles measure up to 10 mm across, feldspar fragments up to 8 mm across. On the Witvlei Ridge the Weissberg Member transgressed rocks of the Buschmannsklippe Formation onto quartzite of the Kamtsas Formation. Northeast of the Witvlei Ridge the Weissberg Member is developed as facies (b).

The same coarse-grained lithotype forms the Weissberg Member in the Gobabis Synclinorium. On KudÜberg 60 (2218 DD) the maximum grain size is 10 mm; here the uppermost 10 to 15 m, however, are fine grained and even pass into siltstone. In the Netso Syncline (2219 CB and DA) the Weissberg quartzite forms the middle of three quartzite sequences.

Facies (c) occurs locally in the lower portion of the Weissberg Member in the Kehoro area (2218 BA) (Fig. 4.10; Plate 3.3). It consists of **coarse-grained to conglomeratic quartzite** containing densely packed subangular to subrounded clasts of light-grey, milky or brown quartz, quartzite, siltstone and red-brown chert, which measure up to 10 mm across; grains of hematite ore reach up to 4 mm in size. Fragments of Doornpoort-type quartzite and ore are subordinate. The grain-size range forms a continuum from the pebble fraction to the matrix which consists of hematite-cemented quartz grains, feldspar and disintegrated rock fragments. The conglomerate is dark red-brown in colour due to the hematite matrix and coating to the pebbles. This facies is easily distinguished from the light-coloured pebbly quartzite of facies (b). The conglomerate forms layers 30 to 60 cm thick, some of which are graded. Also included in the succession are reddish-brown pebble-free layers of medium-grained quartzite.

In the eastern portion of Kehoro S 939, the conglomerate succession of lithofacies (c) is 15 to 30 m thick and is overlain by 50 to 75 m of greyish, gritty quartzite of lithofacies (b). A pit situated in the northern portion of a large pan on Kehoro S 939 (dug while prospecting for gold around the turn of the century, Rimann, 1913, p. 41; 1916) exposes the lower contact of the conglomerate with carbonate of the Okambara Member.

Along the main road on Kehoro 183 only conglomerate of facies (c) is exposed (Plate 3.3); this represents either the entire Weissberg Member, or the overlying quartzite of lithofacies (b) is suppressed by faulting. Further to the south, on both sides of the Black Nossob River, only lithofacies (b) occurs.

The Weissberg Member is considered to be a correlate of the Kanies Member of the Dabis Formation of the central and southern portions of the Nama basin (SACS, 1980; Germs, 1983). However, whether the succession should be considered equivalent not only of the Kanies Member but rather of the entire Dabis Formation in the southern Nama terrain depends on the interpretation of the nature of the boundary between the Dabis and Zaris Formations in the study area and of the position of the Zenana Member of the Zaris Formation (see Chapter 4. 3 .2. 3; also Germs 1983, Figs 4.2 and 4.3).

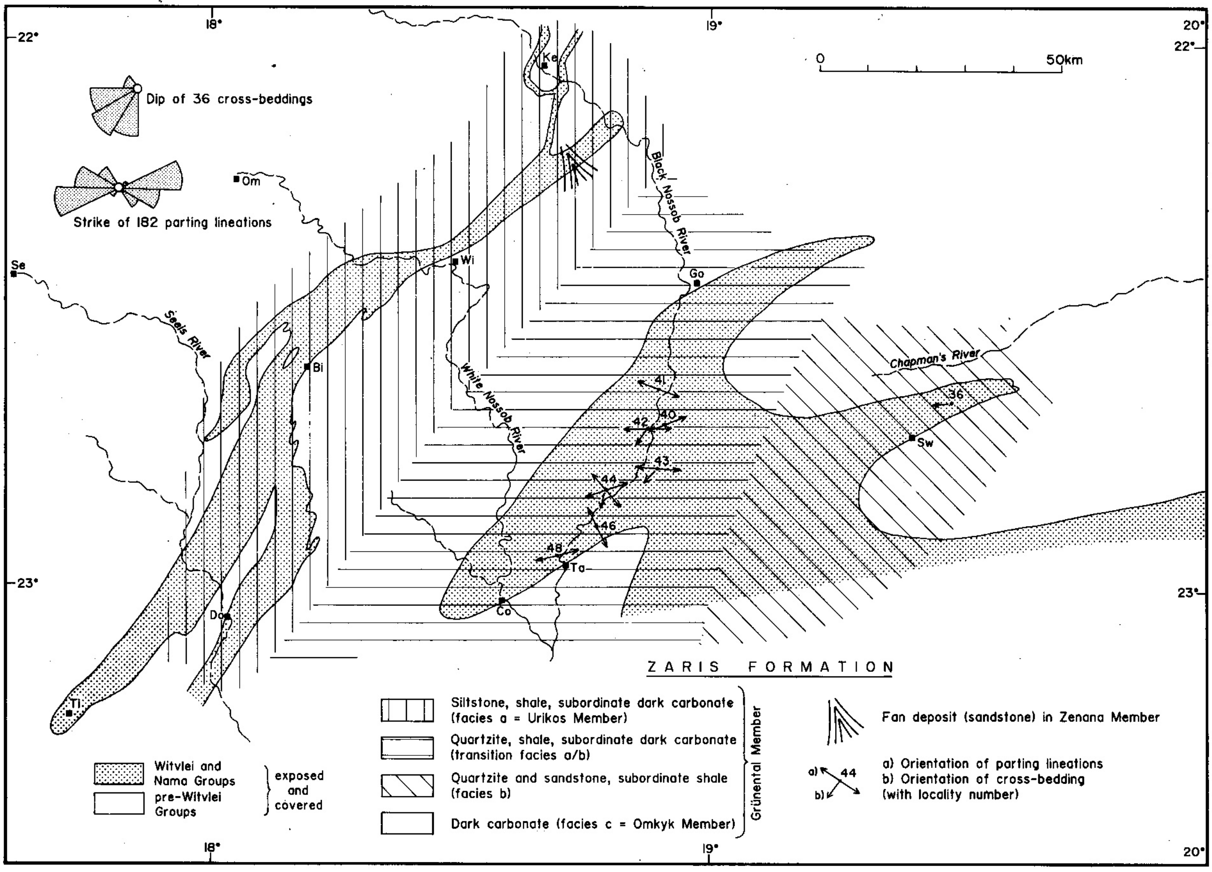


Figure 4.11: Zaris Formation. Facies distribution and orientation of sedimentary structures. For localities see Figure 1.2; for localities of parting lineations and crossbedding refer to corresponding numbers in Table 6.1

4.3.2 Zaris Formation

Rocks of this formation are preserved in the cores of synclines where, due to their soft nature, they underlie flat country bounded by ridges of hard Weissberg quartzite (Plates 3.1 and 3.2).

4.3.2.1 Zenana Member

This member can be subdivided in two lithological units:

(a) The **lower unit** consists of a 10- to 30-m-thick dolomite zone which is massively layered and mottled to banded in tones of dark grey, greenish grey, yellowish and brownish grey. In places white calcite veins cut the rock. Whereas in the Witvlei Synclinorium the zone consists of one carbonate succession only, in the southern Gobabis Synclinorium, between the farms Eilenriede 53 and Keitsaub 68 (2218 DC and DD), it splits up into three carbonate layers, separated by light-grey quartzite of the Weissberg type; this is well exposed on the southern bank of the Black Nossob River on KudÜberg 60 (2218 DC). The lowest of the dolomite bands is considered the basal layer of the Zenana Member. However, the development of the quartzitic intercalations seems to be a local feature since to the northeast of the occurrence, on KudÜberg 60, as well as to the southwest on Kanabis 55 (2218 DC), only one carbonate zone is developed. At the latter locality the unit forms a wide outcrop south of the trigonometrical beacon.

In the Netso Syncline (2219 CB, DA) the lower unit of the Zenana Member forms a depression between the quartzite ridges of the Weissberg Member and the upper unit of the Zenana Member. Here most of the lower unit consists of shale and mudstone; such rocks are the only ones exposed at a pan in the central portion of Soetblomspan 596 and have been intersected in a borehole on Poortje 287 (2219 DA).

(b) The basal carbonate zone is overlain by the **upper unit**, a quartzite resembling the Weissberg quartzite. It forms thinner beds, is slightly darker than the Weissberg rocks and is devoid of gritty inclusions. Ripple marks are scarce and usually indistinct. There are intercalations of greenish to medium-grey siltstone and shale and thin layers of dark-greyish carbonate resembling that at the base of the member. In the Witvlei Synclinorium this zone is probably less than 50 m thick, but this thickness increases locally. At the northern corner beacon on Mahagi 302 (2218 BA), on the northwestern side of the Eindpaal Syncline, the strata are between 80 and 100 m thick and consist of compact, medium-grey, medium-grained, slightly feldspathic quartzite. A thinly bedded siliciclastic succession of fine-grained sandstone and siltstone also occurring on Mahagi 302, on the southeastern side of the Eindpaal Syncline, is considered to belong to these strata as a more distal part of a fan (Fig. 4.11). In the southern portion of the Gobabis Synclinorium the quartzite of the upper unit is prominent, resembling that of the overlying Grünental Member. The boundary between the Zenana and Grünental Members is

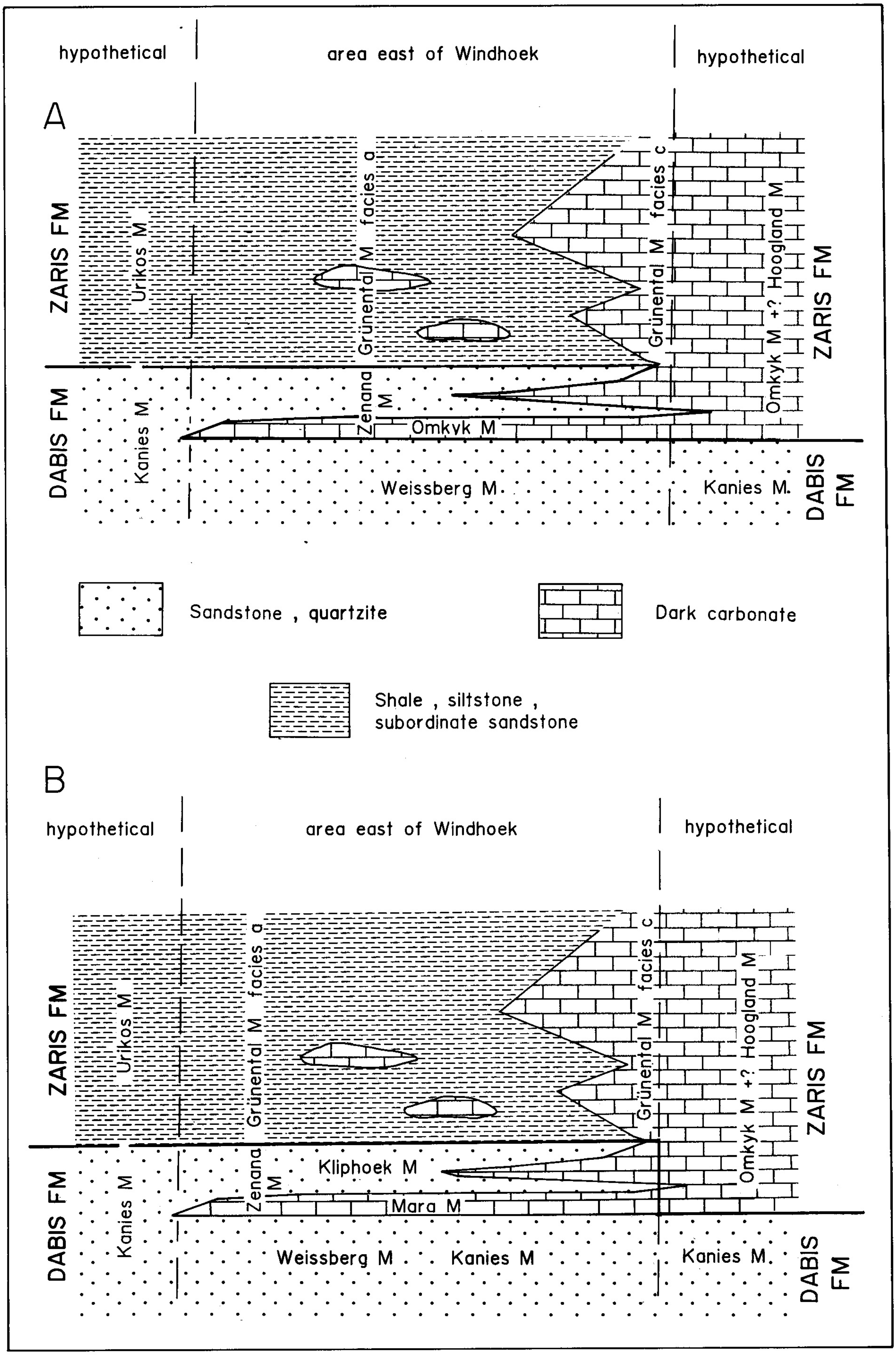

hypothetical
area east of Windhoek
hypothetical
A
ZARIS FM
Urikos M
Grünental M facies a
Grünental M facies c
Omkyk M +? Hoogland M
ZARIS FM
DABIS FM
Kanies M
Zenana M
Omkyk M
Weissberg M
Kanies M.
DABIS FM
Sandstone , quartzite
Dark carbonate
Shale , siltstone , subordinate sandstone
B
hypothetical
area east of Windhoek
hypothetical
ZARIS FM
Urikos M
Grünental M facies a
Grünental M facies c
Omkyk M +? Hoogland M
ZARIS FM
DABIS FM
Kanies M
Zenana M
Kliphoek M
Mara M
Weissberg M
Kanies M
Kanies M.
DABIS FM

Figure 4.12

therefore indistinct; it is placed where the grain size of the sediments changes from medium to fine, and the colour from grey to brown.

In the Netso Syncline the Zenana quartzite forms the innermost (upper) of the three quartzite bands which outline the margin of the synform. Here this quartzite and the Weissberg Member are very similar; both contain clay pellets. On Poortje 287 (2219 DA) it is 7 to 10 m thick, much thinner than both the underlying quartzitic sequences in this region.

At several localities, e.g. along the Black Nossob River on the farms Eilenriede 53 (2218 DC) and KudÜberg 60 (2218 DD), sedimentary structures like current lineations, flute casts and desiccation cracks are well preserved. Mud pellets abound in the thin- to medium-bedded quartzite; these are mostly weathered away and leave a fretted surface.

4.3.2.2 Grünental Member

This member has a widespread occurrence in the Witvlei and Gobabis Synclinoria where it is preserved in the cores of major synclines. Poor exposures and folding make estimation of the thickness of the succession difficult; it possibly exceeds 150 m. The member is devoid of marker bands or distinct changes in lithofacies.

Three laterally interfingering facies exist (Fig. 4.11):

Facies (a) prevails in the Witvlei Synclinorium and consists of a siltstone and shale succession with minor carbonate intercalations. It underlies flat terrane which is widely covered by loam derived from weathering of the fine-grained sediments (Plate 1.1). Hillocks a few metres high and some tens to a hundred metres across in such terrane, consist of dark, massive to thickly.bedded limestone lenses which are intercalated at different levels in the clastic sediment. They stand out from the surroundings and are mostly covered by surface limestone.

In the western portion of the Gobabis Synclinorium thick beds of immature red-brown to grey sandstone are intercalated in shaly to silty successions, as described above, and form a transition between facies (a) and (b) (Fig. 4.11). The sandstone is well bedded in 1- to 50-cm-thick layers; individual layers are finely laminated. Sedimentary features like current lineations, trough crossbedding and mud pellets are common.

Facies (b) prevails in the central region of the Netso Syncline (2219 CB); it is characterised by fine-grained, medium- to thinly layered crossbedded quartzite. Siltstone and dark shale which are mostly covered, may occur.

Facies (c) occurs only in the Rooiwater Syncline, the southwestern spur of the Witvlei Synclinorium (Fig. 4.11), between Graswêreld 272 and Tigerpforte 59 (areas 2317 BB and BC). Dark-grey, massive to medium-bedded, locally laminated limestone (plate 4.12) is exposed in the core of the synform; dark-grey shale, siltstone and thin quartzitic layers are intercalated. This succession is best exposed near the old homestead on Tigerpforte 59 (2317 BC), where a 15-m-thick layer of shale is interbedded in the carbonate succession. The thickness of the exposed section is estimated to be 200 m.

Boreholes on Klausgrund 344 (2218 CC) penetrated dark shale and mudstone only, indicating a decrease in clastic content from the north (facies a) to the south (facies c).

4.3.2.3 Subdivision of the Zaris Formation and correlation of its subunits

The Weissberg Member of the Dabis Formation in the study area corresponds lithologically to the Kanies Member in the Aranos Syncline (SACS, 1980; Germs, 1983), being the lower most unit of the Nama Group which was observed in both regions (see Table 4.1). It therefore forms a marker for the correlation of the overlying strata, which is complicated, however, by the absence of continuous outcrop for about 150 km between the study area and the Aranos Syncline.

The correlation of the Zenana Member with a stratigraphic unit in the Aranos Syncline poses a problem which, however, is felt to be of a terminological nature.

(a) The basal carbonate unit of the Zenana Member in the study area can be included in the Omkyk Member of the Zaris Formation in the Aranos Syncline (Germs, 1983). The overlying light-coloured quartzite of the upper unit of the Zenana Formation is similar to the Weissberg quartzite of the Dabis Formation. Therefore the sequence in the study area can be correlated with the Aranos Syncline and interpreted as an interfingering of the upper portion of the Kanies quartzite (Dabis Formation) with the lower part of the Omkyk carbonate as indicated on Fig. 4.12A and by Germs (1983, Fig. 5).

(b) Following Germs' (1983, Fig. 6) stratigraphic subdivision of the southern Aranos Syncline, the lower unit of the Zenana Member in the study area can be correlated with the Mara Member (Dabis Formation) and the quartzitic upper unit with the Kliphoek Member of the Dabis Formation (Fig. 4.12B).

Both contradicting correlations can be reconciled by an interfingering of the Dabis and Zaris Formations, as is also

Figure 4.12 (left): Correlation of the Zenana Member of the Zaris Formation. interfingering of the Dabis and Zaris Formations presents two possibilities for correlating the Zenana Member. Interpretation B is favoured in this memoir.

A - Carbonate rocks at the base of the Zenana Member are correlated with the Omkyk Member of the Zaris Formation and the overlying quartzite with the Kanies Member of the Dabis Formation, indicating an interfingering of both formations.

B - The two lithofacies of the Zenana Member are correlated with the Mara and Kliphoek Members of the Dabis Formation (Germs, 1983), respectively. In the study area this interpretation means that the contact between the Dabis and Zaris Formations is drawn at the basal carbonate of the Zenana Member and the overlying quartzite is included with the same member and therefore belongs to the Zaris Formation indicated in Table 4.1 by Germs (1983) where the Mara Member (Dabis Formation) is correlated with parts of the Omkyk Member (Zaris Formation). As the basal carbonate of the Zenana Member forms the first incursion of the Omkyk carbonate facies, it seems justified that the boundary is placed between the Dabis and Zaris Formations in the study area at the base of the Zenana Member, the quartzitic portion of which represents a recurrence of the Weissberg quartzite facies.

Facies (c) of the Grünental Member is lithologically very similar to the Omkyk Member of the Zaris Formation (Germs, 1983; SACS, 1980) and is therefore considered to be its equivalent (Fig. 4.12). It may also include portions of the overlying Hoogland Member and, in its lower part, the carbonate facies of the Zenana Member.

In the Aranos Syncline, the fine-clastic facies equivalent of the Omkyk and Hoogland Members is the Urikos Member (Germs; 1983). Consequently, facies (a) of the Grünental Member can be correlated with the latter. No equivalent for facies (b) of the Grünental Member is known in the Aranos Syncline further south.

5. PETROGRAPHY

5.1 CLASTIC SEDIMENTARY ROCKS

5.1.1 Method of Investigation

About 100 samples were collected from clastic rocks of most members of the Witvlei and Nama Groups in the study area. A few samples of Doornpoort and Kamtsas Formations were investigated for comparison since these rocks are a possible source for the Witvlei and the lower Nama rocks.

Petrographic compositions of the framework are depicted in triangular graphs (Fig. 5.1a-c). For a definition of R (rock fragments) the writer follows Pettijohn *et al.* (1973) who include chert. Table 5.1 shows the three main constituents (Q, F, R), converted to 100% (values without brackets). Other components listed in Table 5.1 (in brackets) are generally subordinate constituents, but they are either

Sample No.	Locality	Area	Quartz %	K-feldsp.(%) (excl. microcline)	microcl. + plagiocl. (%)	Rock fragm. (%)	Mica (%)	Ore (%)	Matrix (%)	Cement (carb.) (%)	Topaz (%)
	Doornpoort Formation										
433	Kehoro-S 939	2218 BA	60	24	13	3	[1]	[8]	[4]	-	-
434	Anstatt 250	2318 AA	71	15	9	5	[1]	0	-	-	-
	Kamtsas Formation										
430	Gobabis	2218 BD	75	15	<1	10	-	-	[<1]	-	[2-3]
431	Springboklaagte 20	2317 BB	67	22	5	6	-	-	[<1]	-	[<1]
432	Mentz 65	2318 BA	70	22	5	3	(<1]	-	-	[5]	-
	Tahiti Member										
120	Court 32	2318 BA	73	18	<1	8	[5]	-	[15]	[5]	-
123	Schwarzwald 344	2219 CB	70	20	2	8	-	-	-	-	-
124A	Schwarzwald 344	2219 CB	62	25	5	8	-	-	-	-	-
124B	Schwarzwald 344	2219 CB	65	25	5	5	-	[3]	-	-	-
429	Tahiti 61	2218 DC	67	28	3	2	-	-	-	-	-
	Constance Member										
141	Klippiespan 238	2218 CC	93	5	2	-	[10]	[10]	[25]	[10]	-
142	Court 32	2318 BA	80	20	<1	-	[5]	[20]	[4]	[2]	-
146	Kanabis 54	2218DC	84	15	1	-	[12]	-	[10]	-	-
150	Aais I 31	2318 BA	78	20	1	1	[3]	[4]	[8]	[8]	-
403	Court 32	2318 BA	95	5	<1	-	[12]	[18]	[15]	[5]	-
	Simmenau Member										
152	Kanabis 55	2218 DC	78	10	2	10	-	-	-	[8]	
156	Breitenberg 51	2218 DC	95	-	<1	<5	-	-	-	++	
160	Styria 52	2218 DC	88	5	<1	7	-	-	-	-	
161	Constance 230	2218 CC	76	12	<1	12	-	-	-	-	
162	Gobabis	2218 BD	81	5	1	13	-	-	-	-	
164	Josephine 226	2218 CC	74	18	1	7	-	-	-	-	
165	Doreen 227	2218 CC	79	12	<1	9	-	-	-	-	
166	Eilenriede 53 (lower member)	23128 BA	71	13	1	15	-	-	-	-	
167	Eilenreide 53 (middle member)	2318 BA	85	10	-	5	-	-	-	-	
168	Eilenriede 53 (upper member)	2318 BA	85	5	1	9	-	-	-	-	
169	Kuduberg 60	2218 DD	80	13	<1	6	-	-	-	[5]	-
170	Dakota 35	2318 BA	69	13	1	15	-	-	-	[2]	-
171	Poortje 287	2219 DA	75	15	<1	10	-	-	-	+	-
172	Poortje 287	2219 DA	79	12	<1	9	-	-	-	-	-
173	Valerie 291	2219 DA	83	5	1	11	-	-	-	+	-
174	Josephine 226	2218 CC	69	15	1	15	-	[2]	[5]	-	[2]
418	Eilenriede 53 (lower member)	2318 BA	82	10	-	8	-	-	-	-	-

able 5.1: Mineralogical composition of arenites. Continued overleaf

Sample No.	Locality	Area	Quartz (%)	K-felds. (%) (excluding microcline)	microcl. + plagiocl. (%)	Rock frag. (%)	Mica (%)	Ore (%)	Matrix (%)	Cement (carb.) (%)	Topaz (%)
	La Fraque Member										
197	Okambara 219	2218 CA	90	10	-	-	[2]	[18]	[5]	[35]	-
210	Eilenriede 53	2318 BA	90	10	-	-	[2]	[8]	-	[15]	-
	Okambara Member										
215	Kowas 233 (uppermost unit)	2218 CC	95	<1	<1	4	-	-	-	[40]	-
219	Bildah 220	2218 CA	85	15	<1	-	[1]	[8]	-	[15]	-
222	Okambara 219	2218 CA	85	15	<1	<1	-	[8]	[10]	-	-
228	Achenib 247	2318 AA	92	8	<1	1	[1]	[1]	-	-	-
413	Constance 230	2218 CC	86	13	<1	1	[<1]	[8]	[3]	-	-
414	Kuduberg 60 (arenaceous facies)	2218 DD	84	15	<1	1	[3]	[10]	[3]	-	-
	Weissberg Member										
232	Kehoro-S 939 (facies c, congl.)	2218 BA	90	10	-	-	[2]	[13]	-	-	[2]
235	Nudom 161 (facies c, congl.)	2218 BA	91	4	-	5	[<1]	[3]	-	-	-
236	Wonderboom 301	2218 BC	84	11	<1	5	-	-	-	-	[2]
237	Lee Enfield	2218BC	91	5	<1	4	-	-	-	-	[5]
238	Nudom 161	2218 BA	78	15	<1	7	-	-	-	-	[13]
240	Kowas 233 (lowermost member)	2218 CC	89	4	-	7	-	-	-	-	[<1]
250	Kuduberg 60	2218 DD	83	13	<1	4	-	-	-	-	[<1]
251	Styria 52	2218 DC	87	8	<1	5	-	-	-	-	-
253	Nudom 191	2218 BA	88	5	2	5	-	-	-	-	[3]
254	Eilenriede 53	2218 DC	87	8	1	4	-	-	-	-	-
255	Grünental 151	2218 AD	85	10	<1	5	-	-	-	-	-
259	Nautabis 268	2317 BB	98	<1	-	2	-	-	-	-	-
264	Soetblomspan 596 (lowermost member)	2219 DA	95	3	-	2	-	-	-	-	-
417	Okambara 219 (lowermost member)	2218 CA	93	4	-	3	-	-	-	-	-
425	Scheidthof 293	2218 CA	98	1	-	1	-	-	-	-	-
426	Okambara 219	2218 CA	97	3	-	<1	-	-	-	-	-
	Zenana Member										
252	Grünental 151	2218 AD	83	13	2	2	-	-	-	-	-
275	Eilenriede 53	2218 DC	96	4	-	-	-	-	[<1]	+	-
276	Soetblomspan 596	2219DA	97	2	-	1	-	-	-	[1]	-
277	Mahagi 302	2218 BA	90	5	-	5	-	-	-	-	[2]
278	Mahagi 302	2218 BA	86	8	1	5	-	-	[<1]	-	[4]
	Grünental Member										
265	Blumenau 58	2218 DD	96	4	1	-	[3]	[15]	[4]	[26]	-
266	Friedland 56	2218 DC	86	7	2	5	[2]	-	[15]	-	-
268	Karossewie 72	2218 DB	92	7	1	-	[10]	[3]	[3]	[35]	-
269	Tygerpoort 285	2219 CB	89	5	1	6	-	-	[1]	-	-
270	Kaukurus-E 79	2218 DB	81	10	1	8	[2]	[3]	[10]	-	-
271	Karossewie 72	2218 DB	85	10	1	4	[<1]	[3]	[10]	-	-
272	Karossewie 72	2218 DB	86	8	2	4	[<1]	[2]	[10]	-	-
273	Brakputs 79	2218 DB	90	7	2	1	[1]	[10]	[25]	-	-
274	Spatzenfeld 70	2218 DD	83	9	3	5	[1]	[5]	[12]	-	-
378	Okasewa 103	2218 AD	81	18	1	-	[10]	[3]	[30]	-	-

Table 5.1 (continued): Mineralogical composition of arenites. Minerals identified from thin sections, K-feldspar by staining. Numbers without brackets are expressed as percentages of the total QRF used for the triangular diagrams (Figs 5.1a to 5.1c), while numbers in brackets are expressed as percentages of the total number of grains counted

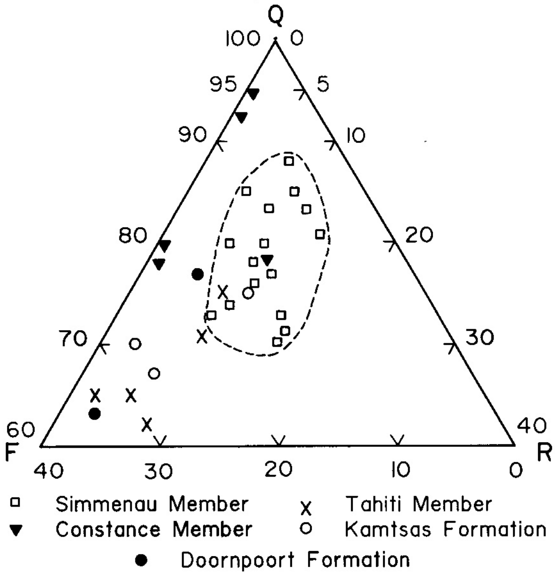


Figure 5.1a

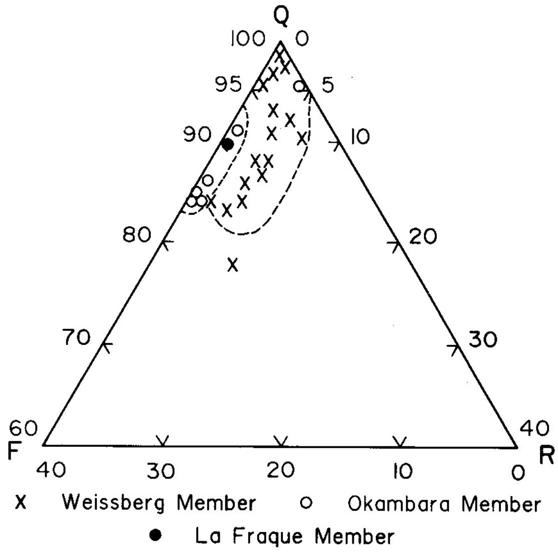


Figure 5.1b

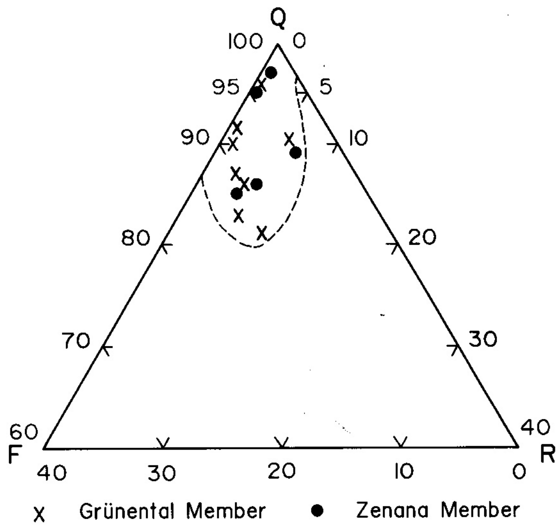


Figure 5.1c

Figure 5.1: Petrographic composition of framework of clastic sediments from pre-Witvlei source rocks, as well as Witvlei and Nama arenites, showing percentage of quartz (Q), feldspar (F), rock fragments and chert (R)

a - Arenites of Doornpoort and Kamtsas Formations (pre-Witvlei) and Tahiti, Constance and Simmenau Members (Court Formation, lower Witvlei Group)

b - La Fraque and Okambara Members (Buschmannsklippe Formation, upper Witvlei Group) and Weissberg Member (lowermost Nama Group)

c - Grünental and Zenana Members (Zaris Formation, lower Nama Group)

characteristic for a certain source area or remain crucial for classification.

The classification of sandstone follows Folk (1968) and that of textural maturity Folk (1951) (cf. Klein, 1978; Conolly, 1978; Germs, 1972a; Pettijohn *et al.*, 1973). A generally important indicator of maturity, viz. the degree of roundness of mineral grains, could in many cases not be observed since pressure solution or secondary overgrowth obliterated the original outlines and resulted in the development of either straight contacts with triple junctions, or sutured grain boundaries (Plate 5.1). In some cases a thin clay or hematite coating still defines the original outlines of quartz grains, indicating that most of these were originally rounded to well-rounded. Optically oriented secondary overgrowth of quartz in most cases fills interstices.

Amongst the feldspars, K -feldspar is more abundant than plagioclase and plagioclase exceeds microcline. Feldspar is not equally distributed in all grain-size classes; the generally anhedral orthoclase grains have the same size as the largest quartz grains. Optically oriented overgrowth of plagioclase is implied by the euhedral form of many of these.

For granulometric analysis, between 200 and 300 grains per thin section were measured, a number considered sufficient to provide all necessary information according to Friedmann (1958). The values obtained were grouped in intervals of 0.25 Φ. Cumulative curves were constructed using a graph paper designed by Friedman (1958) which allows for conversion of values of size distribution in thin sections to sieve size distribution. Percentiles were read from the graphs (5%, 16%, 25% = first quartile, 50% = median, 75% = third quartile, 84% and 95%), while mean and sorting values were calculated according to formulae by Folk and Ward (1957). The results are shown in Table 5.2. As certain trends of the size distribution stand out more clearly on log -probability paper than on arithmetic-scale paper, the construction of those curves was carried out from count values obtained graphically from the arithmetic curves.

5.1.2 Doornpoort Formation

Two samples of submature to mature, very fine-grained sandstone (Table 5.2) comprise more than 25% unstable constituents of which about 25 % are feldspar (Table 5.1 and Fig. 5.1a). Most of the feldspar is sericitised. The sandstone falls into the category of subarkose to arkose. Poor

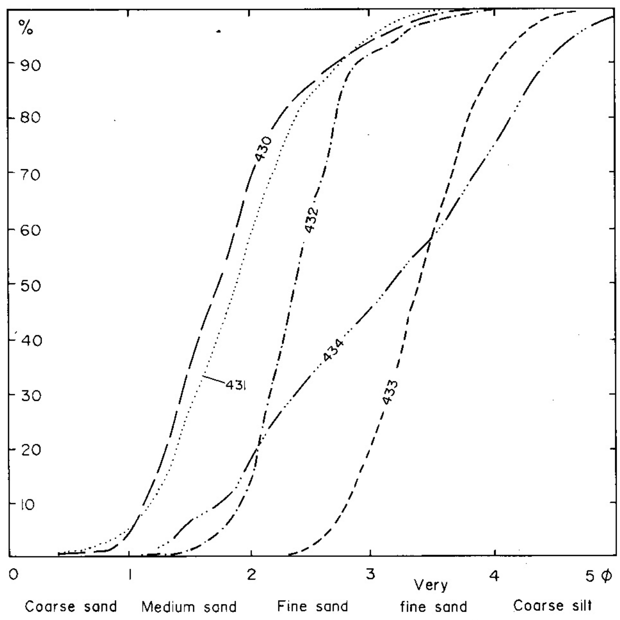


Figure 5.2: Cumulative curves of grain-size distribution in the Doornpoort Formation (samples 433, 434) and Kamtsas Formation (samples 430, 431, 432). See Table 5.1 for location of all samples

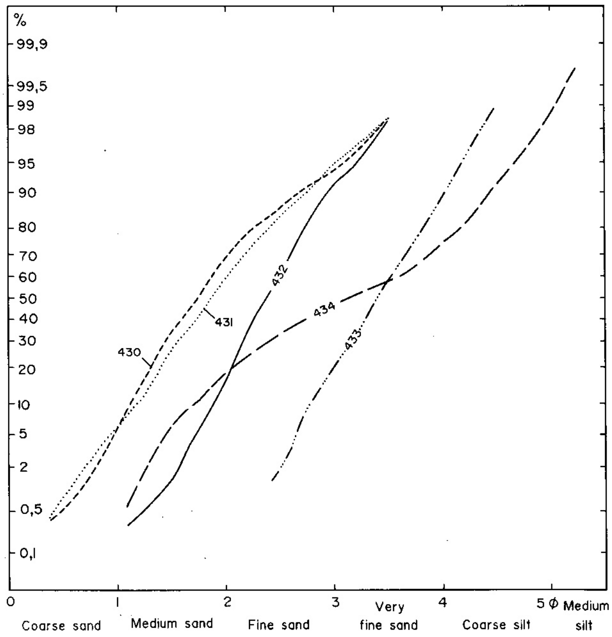


Figure 5.3: Log-probability histogram of grain-size distribution in the Doornpoort Formation (samples 433, 434) and Kamtsas Formation (samples 430, 431, 432). See Table 5.1 for location of all samples

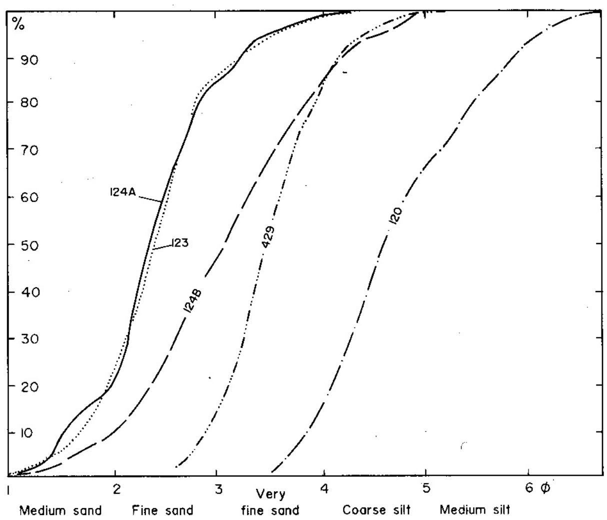


Figure 5.4: Cumulative curves of grain-size distribution in the Tahiti Member. See Table 5.1 for location of all samples

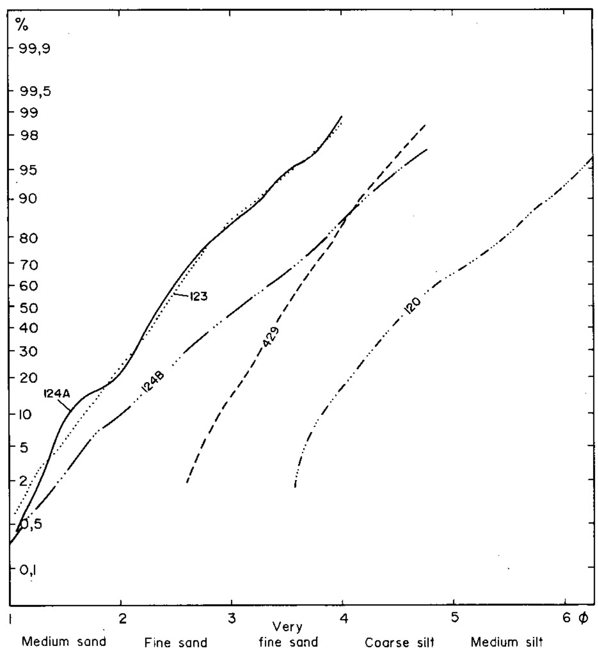


Figure 5.5: Log-probability histogram of grain-size distribution in the Tahiti Member

sorting of sample 434 (Fig. 5.2 and 5.3) is caused by a bimodal grain-size distribution, the fine-grained components averaging around 0.05 mm (= +4.25 Φ), the coarser around 0.25 mm (= +2.00 Φ). Both size classes form either alternating laminae which are between 0.1 and 0.5 mm thick, or they are mixed, indicating rapidly changing energy and transport conditions. The percentage of matrix is insignificant, consisting mostly of disintegrated lithic fragments (pseudomatrix), accompanied by subordinate diagenetic micas.

5.1.3 Kamtsas Formation

Three samples of medium- to fine-grained, moderately sorted arkose (Table 5.2) reveal the submature nature of the sediment. No matrix is present and secondary mica (chlorite) in interstices amounts to less than 1 %. The topaz content of 2-3% (Table 5.1) in the northernmost sample (430) deserves attention because of the concentration of the same heavy mineral in certain younger samples in the northern Witvlei Synclinorium. An occurrence of car-

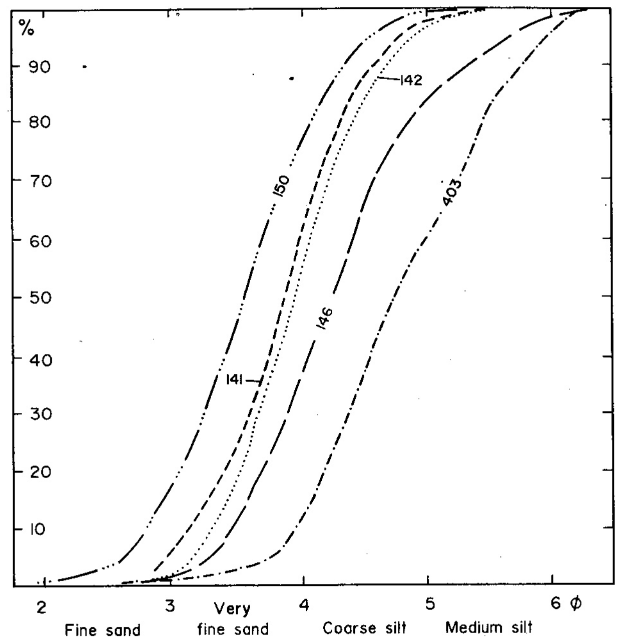


Figure 5.6: Cumulative curves of grain-size distribution in the Constance Member

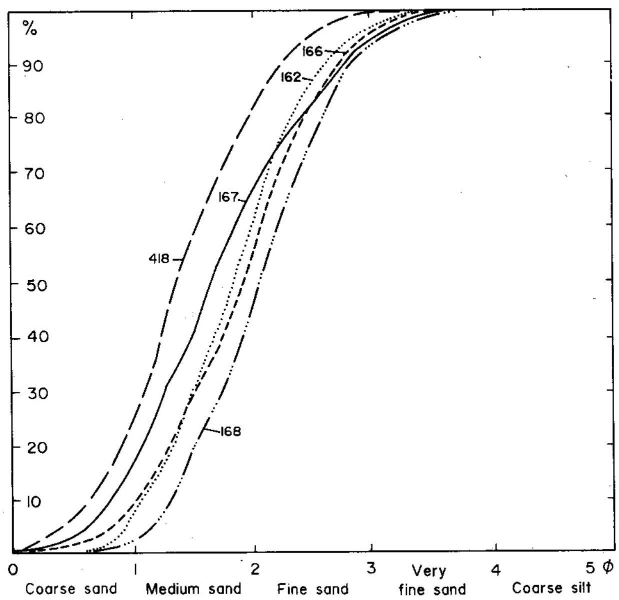


Figure 5.7: Cumulative curves of grain-size distribution in the Simmenau Member in the southern and northern Gobabis Synclinorium

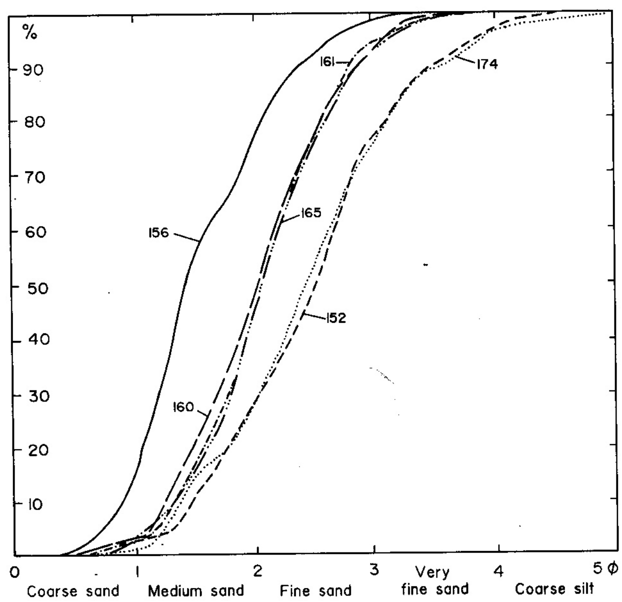


Figure 5.8: Cumulative curves of grain-size distribution in the Simmenau Member in the southwestern Gobabis and southern Witvlei Synclinoria

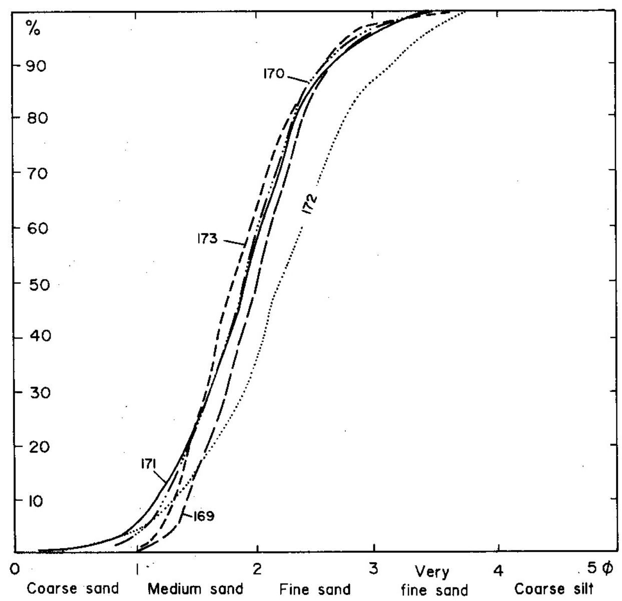


Figure 5.9: Cumulative curves of grain-size distribution in the Simmenau Member in the southern and eastern Gobabis Synclinorium

bonate-cemented, fine-grained arkose from the Hoesberg on Josephine 226 (2218 CA) contained about 33% micritic cement; the quartz grains are corroded.

5.1.4 Court Formation

5.1.4.1 Tahiti Member

Five moderately sorted samples of fine-grained sandstone to siltstone fall within the classification of an arkose (Tables 5.1 and 5.2; Figs 5.4 and 5.5). Disintegrated mica schist makes up a high percentage of pseudomatrix in thin section 120 while less-weathered schist particles constitute most of the rock fragments. Sample 120 is immature, while the others are texturally submature.

5.1.4.2 Constance Member

The grain sizes of the five samples from the arenaceous portion of the member cluster within a narrow range, from fine sandstone to coarse siltstone (Table 5.2 and Fig. 5.6). The rocks are characterised by up to 25% clayey matrix and pseudomatrix and 12% detrital mica (Table 5.1), i.e. they are arkosic wackes (in the sense of Pettijohn *et al.*, 1973, Fig. 5-3). Flakes of detrital mica may be up to 0.5 mm in diameter, many of them squeezed and broken between other

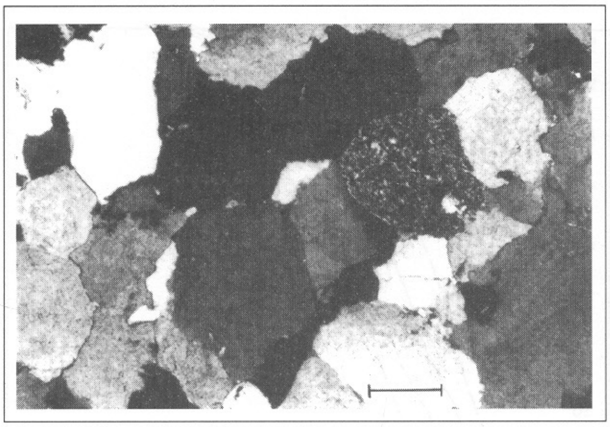


Plate 5.1: Moderately sorted, medium-grained, mature arenite of the Simmenau Member. Boundaries of quartz grains are straight or sutured; note triple junctions, secondary quartz overgrowth and rock fragment in upper-right quadrant. Eilenriede 53. Sample 167, X nic., bar is 0.3 mm long

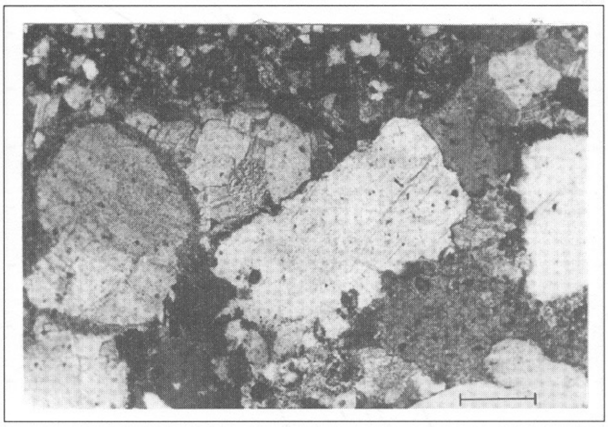


Plate 5.2: Conglomerate of the Simmenau Member. Note unsorted, subrounded to angular quartzite grains; rock fragment along upper margin; carbonate matrix either dusty micrite (lower right-hand side) or strongly recrystallised (left side above centre). Quartz grain on left-hand side is surrounded by a micritic seam. Schönborn 79F. Sample 158, X nic., bar is 0.3 mm long

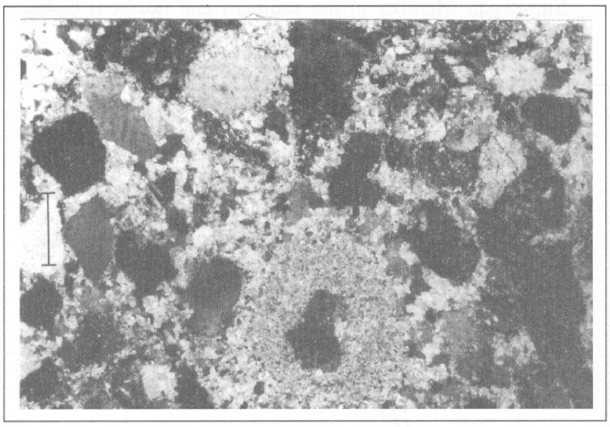


Plate 5.3: Calcareous sandstone of the marginal facies of the Bildah Member. It contains 50% siliciclastic material (quartz, plagioclase, rock fragments including fine-crystalline lava on right side). Carbonate matrix is mostly sparry. The coating of ooids has been obliterated by recrystallisation (lower centre). Sachsenwald 186 (940). Sample 184, X nic., bar is 0.3 mm long

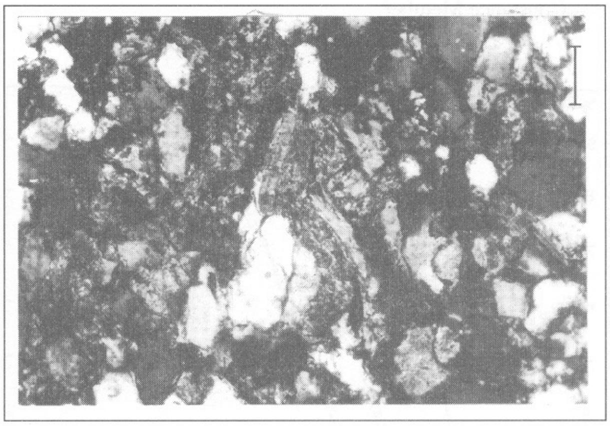


Plate 5.4: Very fine-grained, moderately sorted, impure arenite of the Grünental Member with 25% matrix (pseudomatrix) consisting of phyllosilicate. The rock is a subarkosic wacke. A large detrital mica flake, partly altered, can be seen bent around a quartz grain before being finally broken during compaction. Brakputz 79. Sample 273, X nic., bar is 0.1 mm long

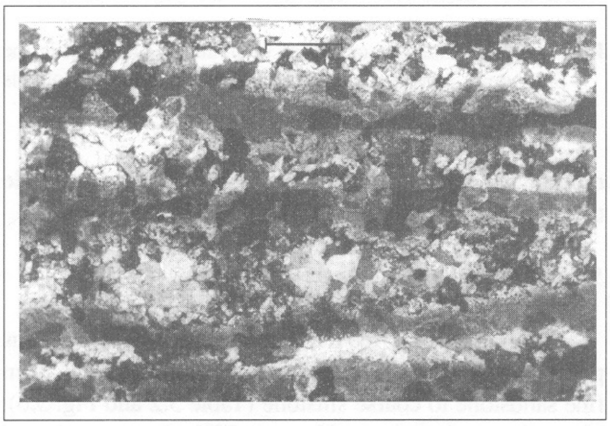


Plate 5.5: Laminated carbonate (rhythmite) of the "deep-water" facies of the Gobabis Member. X nic. Dark lamination passes through sparite crystals without being influenced by the outline of these crystals. This is ascribed to secondary recrystallisation. Bar is 0.3 mm long

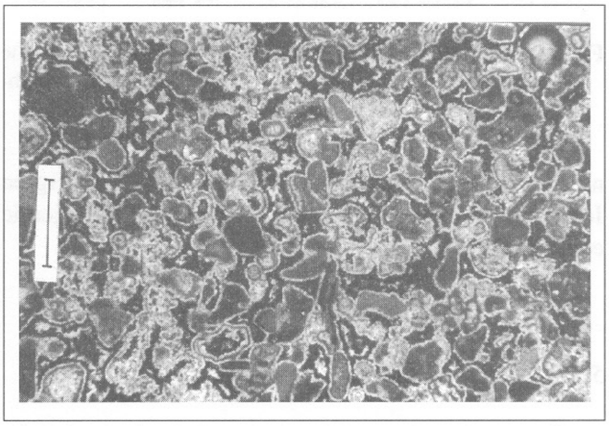


Plate 5.6: Grainstone of the Constance Member. Coated grains are irregularly formed and are surrounded by rims of light-coloured, fibrous carbonate. Several of the cores are leached and filled by large rhombs of the same type of dolomite as in the matrix. Northwestern portion of Tahiti 61. Sample 153, X nic., bar is 0.5 mm long

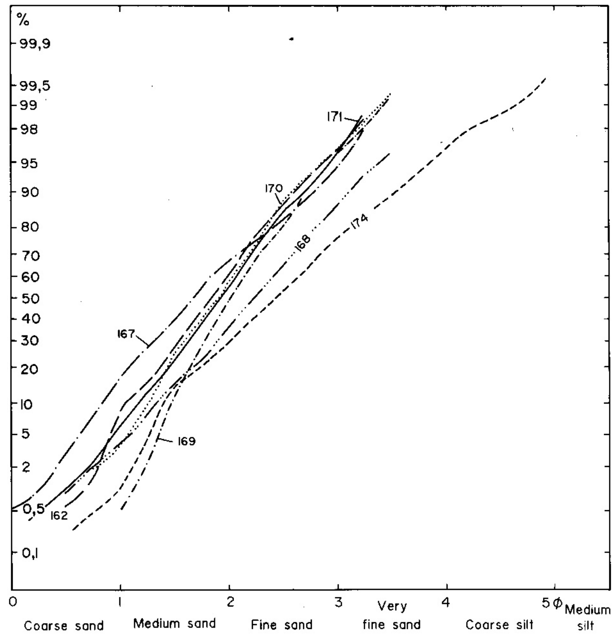


Figure 5.10: Log-probability histogram of grain-size distribution in the Simmenau Member

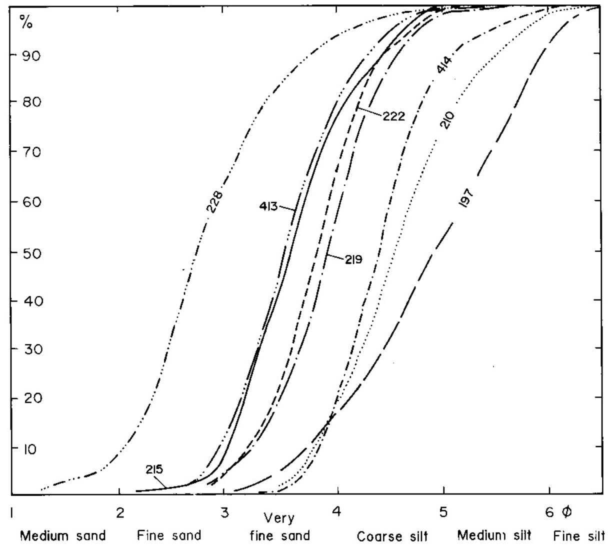


Figure 5.11: Cumulative curves of grain-size distribution of the La Fraque (samples 197 and 210) and Okambara Members (samples 215, 219, 222, 228, 413 and 414)

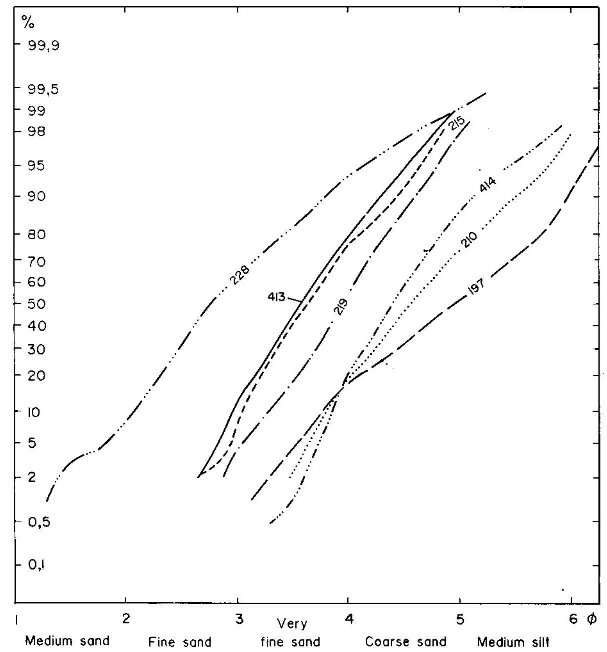


Figure 5.12: Log-probability histogram of grain-size distribution in the La Fraque (samples 197, 210) and Okambara Members (samples 215, 219, 228, 413, 414)

detrital grains; some are superficially chloritised. Depending on the percentage of clay, the rock is either matrix- or grain-supported. Carbonate, one of the main constituents of this member, constitutes up to 10% in the form of a cement. It has caused strong corrosion of the quartz grains.

5.1.4.3 Simmenau Member

Most of the fine- to medium-grained rocks show moderate, and subordinate portions, good sorting (Table 5.2; Figs 5.7-5.10); they are devoid of a matrix. These properties characterise them as submature, locally even mature arenites (Plate 5.1). The presence of nearly equal amounts of feldspar and rock fragments (Table 5.1) places the samples in the marginal region between subarkose and sublith arenite. Most of the lithic particles consist of fine-grained sediment, extremely fine-grained, biotite-bearing metasediment and chert. The carbonate cement in sample 169 (about 5%) is unevenly distributed within the rock and some of it consists of dolomite rhombs.

Two thin sections of the conglomerate on Schönborn 79F showed the following composition of the clasts: 75% medium-grey, banded micritic, locally pelletoidal, carbonate, 20% fine- to medium-grained arkose and quartzite, 5% chert. These components are set in a sparry, sandy carbonate matrix which makes up about 15% of the whole rock (Plate 5.2). Several of the smaller clasts are superficially coated by carbonate. Size and percentage of the clasts are highly variable.

5.1.5 Buschmannsklippe Formation

5.1.5.1 Bildah Member

In the vicinity of the Witvlei Ridge, the otherwise carbonaceous succession contains abundant arenaceous matter, and the dolomite passes into a carbonate-cemented sandstone. Its clastic components consist mainly of quartz which is accompanied by subordinate feldspar and chert. Grain sizes increase towards the basin margin. The quartz grains are angular and corroded. Where the carbonate content exceeds 40%, the clasts float in the cement and many are superficially to multiply coated. With less than 30% carbonate cement the rock becomes clast-supported, carbonate filling the interstices. In this case interstitial, secondary phyllosilicates occur, the flakes aligned either parallel or perpendicular to the boundaries of neighbouring grains; some bridge the gap between two adjoining grains. Where

Sample No.	ø5	ø16	ø25 (Q1)	ø50 (Md)	ø75 (Q3)	ø84	ø95	S	Sorting Class	Me	Wentworth class
Doornpoort Formation											
433	2.65	2.92	3.10	3.40	3.75	3.92	4.23	0.49	good to moderate	3.41	very fine sand
434	1.42	1.92	2.17	3.18	3.96	4.29	4.67	1.08	poor to moderate	3.13	very fine sand
Kamtsas Formation											
430	0.98	1.23	1.38	1.72	2.10	2.42	3.13	0.63	moderate	1.79	medium sand
431	0.96	1.32	1.46	1.88	2.28	2.52	3.02	0.61	moderate	1.91	medium to fine sand
432	1.70	1.99	2.10	2.39	2.67	2.77	3.27	0.44	good to moderate	2.38	fine sand
Tahiti Member											
120	3.71	3.99	4.20	4.62	5.30	5.63	6.20	0.79	moderate	4.75	coarse silt
123	1.48	1.84	2.03	2.42	2.75	2.94	3.59	0.59	moderate to good	2.40	fine sand
124A	1.43	1.79	2.07	2.37	2.76	3.03	3.54	0.65	moderate	2.40	fine sand
124B	1.65	2.22	2.50	3.10	3.75	4.00	4.57	0.89	moderate	3.11	very fine sand
429	2.75	3.05	3.20	3.50	3.80	4.03	4.45	0.51	moderate to good	3.53	very fine sand
Constance Member											
141	2.98	3.32	3.55	3.90	4.22	4.45	4.82	0.56	moderate to good	3.89	very fine sand
142	3.19	3.48	3.65	3.96	4.33	4.53	4.91	0.52	moderate to good	3.99	sand to coarse silt
146	3.31	3.63	3.83	4.26	4.73	5.05	5.65	0.71	moderate	4.31	coarse silt
150	2.65	2.97	3.18	3.58	3.98	4.21	4.55	0.60	moderate	3.59	very fine sand
403	3.80	4.11	4.30	4.75	5.35	5.55	5.97	0.69	moderate	4.80	coarse silt
Simmenau Member											
152	1.00	1.65	1.88	2.52	2.95	3.28	3.87	0.84	moderate	2.48	fine sand
156	0.73	1.00	1.15	1.45	1.98	2.18	2.68	0.60	moderate	1.54	medium sand
160	1.15	1.38	1.58	2.02	2.45	2.68	3.13	0.63	moderate	2.03	fine to medium sand
161	1.18	1.47	1.65	2.00	2.45	2.67	3.10	0.59	moderate to good	2.05	fine to medium sand
162	0.93	1.25	1.43	1.83	2.22	2.45	2.87	0.59	moderate to good	1.84	medium sand
165	1.11	1.52	1.73	2.08	2.55	2.73	3.16	0.61	moderate	2.11	fine sand
166	0.85	1.22	1.42	1.92	2.33	2.57	2.96	0.66	moderate	1.90	medium to fine sand
167	0.21	0.97	1.20	1.66	2.25	2.60	3.07	0.84	moderate	1.74	medium sand
168	1.17	1.45	1.64	2.08	2.55	2.70	3.17	0.61	moderate	2.08	fine to medium sand
169	1.35	1.53	1.69	2.02	2.35	2.52	2.95	0.49	good to moderate	2.02	fine to medium sand
170	1.11	1.38	1.53	1.90	2.26	2.43	2.89	0.53	moderate to good	1.90	medium to fine sand
171	0.95	1.33	1.53	1.90	2.30	2.45	2.95	0.58	moderate to good	1.89	medium sand
172	1.10	1.52	1.78	2.21	2.68	2.92	3.45	0.72	moderate	2.22	fine sand
173	1.23	2.41	1.52	1.80	2.18	2.40	2.78	0.49	good to moderate	1.87	medium sand
174	1.26	1.54	1.90	2.44	3.00	3.26	3.91	0.83	moderate	2.41	fine sand
418	0.40	0.81	1.01	1.40	1.85	2.10	2.50	0.64	moderate	1.44	medium sand
La Fraque Member											
197	3.34	3.98	4.36	4.94	5.63	5.82	6.12	0.88	moderate	4.91	coarse to medium silt
210	3.70	3.96	4.15	4.58	5.01	5.35	5.8	0.66	moderate	4.63	coarse silt
Okambara Member											
215	2.94	3.15	3.28	3.64	3.98	4.26	5.65	0.69	moderate	3.68	very fine sand
219	2.95	3.43	3.62	3.95	4.25	4.45	4.80	0.54	moderate to good	3.94	sand to coarse silt
222	2.95	3.38	3.56	3.88	4.16	4.33	4.75	0.51	moderate to good	3.86	very fine sand
228	1.84	2.22	2.42	2.74	3.28	3.57	4.15	0.69	moderate	2.84	fine sand
413	2.82	3.10	3.25	3.58	3.96	4.14	4.55	0.52	moderate to good	3.61	very fine sand
414	3.77	3.96	4.09	4.40	4.75	4.91	5.50	0.50	good to moderate	4.42	coarse silt

Table 5.2 (continued on opposite page): Grain-size parameters of clastic sediments in phi units. Percentiles graphically measured from respective cumulative curves. S = sorting, Me = mean value after Volk and Ward (1975).

$$S = \frac{\varphi_{84} - \varphi_{16}}{4} + \frac{\varphi_{95} - \varphi_{5}}{6.6}; \quad Me = \frac{\varphi_{16} + \varphi_{50} + \varphi_{84}}{3}$$

For definition of phi units, sorting class and Wentworth class see Pettijohn *et al.* (1973) or Firbridge and Bourgeois (1978). For localities see Table 5.1

Sample No.	ø5	ø16	ø25 (Q1)	ø50 (Md)	ø75 (Q3)	ø84	ø95	S	Sorting Class	Me	Wentworth class
Weissberg Member											
235	0.90	1.34	1.58	2.32	2.95	3.30	3.75	0.92	moderate to poor	2.32	fine sand
236	0.38	1.12	1.35	1.93	2.47	2.70	3.20	0.82	moderate	1.92	medium to fine sand
237	1.15	1.54	1.77	2.13	2.51	2.73	3.02	0.58	moderate to good	2.13	fine sand
238	0.25	0.80	1.01	1.53	2.06	2.20	2.65	0.71	moderate	1.51	medium sand
240	2.47	2.62	2.72	2.96	3.28	3.40	3.85	0.40	good	2.99	fine to very fine sand
250	1.10	1.43	1.64	2.20	2.88	3.11	3.53	0.79	moderate	2.25	fine sand
251	0.45	0.83	0.98	1.38	1.84	2.10	2.70	0.65	moderate	1.44	medium sand
253	1.02	1.38	1.60	2.05	2.51	2.66	2.98	0.64	moderate	2.03	fine to medium sand
254	0.58	0.98	1.20	1.70	2.23	2.52	3.00	0.76	moderate	1.73	medium sand
255	1.50	2.03	2.30	2.72	3.12	3.29	3.69	0.65	moderate	2.68	fine sand
259	2.43	2.67	2.78	3.07	3.35	3.58	4.05	0.47	good to moderate	3.11	very fine sand
264	2.48	2.67	2.77	3.00	3.26	3.43	3.75	0.38	good to very good	3.03	very fine to fine sand
417	2.56	2.77	2.91	3.18	3.49	3.68	3.96	0.44	good	3.21	very fine sand
425	1.45	1.96	2.10	2.47	2.68	2.96	3.33	0.54	moderate to good	2.46	fine sand
426	2.00	2.45	2.58	2.92	3.28	3.49	3.92	0.55	moderate to good	2.95	fine to very fine sand
Zenana Member											
252	2.75	3.13	3.23	3.33	3.65	3.79	4.01	0.36	good to very good	3.42	very fine sand
275	2.60	2.71	2.80	3.05	3.29	3.44	3.92	0.38	good to very good	3.07	very fine to fine sand
276	1.79	2.09	2.27	2.62	2.84	3.01	3.33	0.46	good to moderate	2.57	fine sand
277	0.55	1.12	1.29	1.68	2.15	2.38	2.90	0.68	moderate	1.73	medium sand
278	0.98	1.33	1.53	2.00	2.49	2.68	3.30	0.69	moderate	2.00	medium sand
Grünental Member											
265	4.40	4.80	5.12	5.61	5.95	6.12	6.37	0.63	moderate	5.51	medium silt
266	2.54	2.70	2.95	3.27	3.65	3.80	4.35	0.55	moderate to good	3.26	very fine sand
268	4.23	4.47	4.61	5.13	5.64	5.84	6.20	0.64	moderate	5.15	medium silt
269	2.08	2.31	2.41	2.67	2.90	3.12	3.60	0.43	good	2.70	fine sand
270	2.28	2.55	2.67	3.04	3.55	3.82	4.28	0.62	moderate	3.14	very fine sand
271	2.28	2.59	2.74	3.10	3.36	3.61	4.10	0.53	moderate to good	3.10	very fine to fine sand
272	2.65	2.88	3.00	3.33	3.70	3.93	4.35	0.52	moderate to good	3.38	very fine sand
273	2.58	2.90	3.07	3.34	3.92	4.14	4.52	0.60	moderate	3.49	very fine sand
274	2.73	2.98	3.07	3.33	3.73	3.90	4.22	0.46	good to moderate	3.40	very fine sand
378	3.00	3.52	3.85	4.45	5.17	5.50	6.15	0.98	moderate to poor	4.49	coarse silt

Table 5.2 (continued)

bordering the Witvlei Ridge, the member is represented by a conglomerate devoid of carbonate, with quartz pebbles up to 5 cm across, set in a coarse-grained to gritty matrix with grain sizes continuous from matrix to clasts.

Another terrigenously influenced portion of the Bildah Member on Sachsenwald 940 (2218 BA) consists of marl and sandy carbonate with approximately 50% carbonate and 50% clasts (Plate 5.3). Grains of quartz, feldspar, macrocrystalline quartz aggregates and chert measure up to 0.5 mm across, whereas fragments of fine, micaceous sandstone and basic lava are subangular to rounded and between 0.2 and 1.5 mm across. Lava fragments consist mainly of perthitic feldspar. Clastic components form carbonate-cemented layers alternating with carbonate-dominated laminae.

5.1.5.2 La Fraque and Okambara Members

Samples from these members reveal moderate sorting and fine-sandy to silty grain sizes (Table 5.2, Figs 5.11 and 5.12). The immature subarkoses contain up to 10% matrix (Table 5.1). Transition from pure carbonate to quartzite occurs, in many instances leading to a marly cement in which carbonate is intimately mixed with clay. Hematite, causing the reddish-brown colour of the rocks, is either concentrated in the pore space or finely disseminated in the matrix, or detrital hematite grains are scattered throughout the sediment and may be enriched in certain laminae. In thin section the rocks of the La Fraque Member are characterised by alternating wavy to lenticular siltstone and fine-grained sandstone layers between 0.1 and 1 cm thick, and interspersed pelitic layers. The siltstone and sandstone layers consist of up to 80% quartz, subordinate plagioclase and detrital mica in a sparry carbonate cement and are grain-supported, whereas the pelitic laminae consist of mudstone, finely disseminated hematite which may be concentrated in thin seams, detrital and diagenetic mica (up to 25%), and micritic carbonate. Silt-sized quartz grains occur scattered and are rare in the pelitic beds. Layering is enhanced by variable amounts of silty material (the higher the percentage, the lighter the colour) and hematite, the latter being enriched in the pelitic zones and effecting a red-brown colour banding.

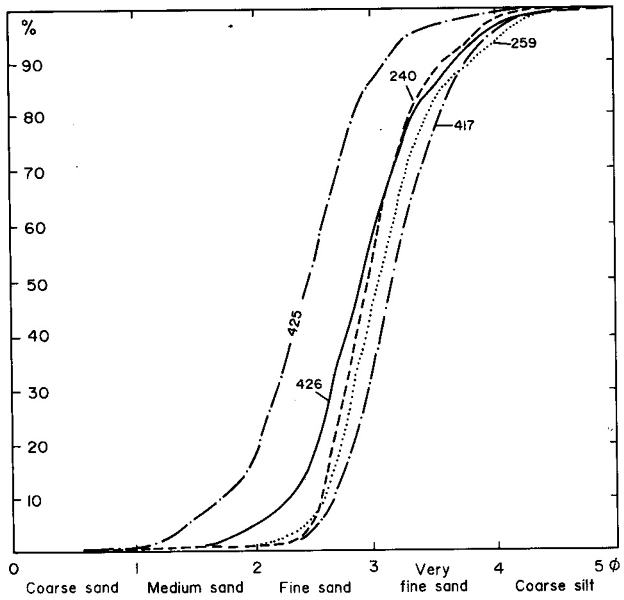


Figure 5.13: Cumulative curves of grain-size distribution in the Weissberg Member in the central and southern Witvlei Synclinorium (facies (a))

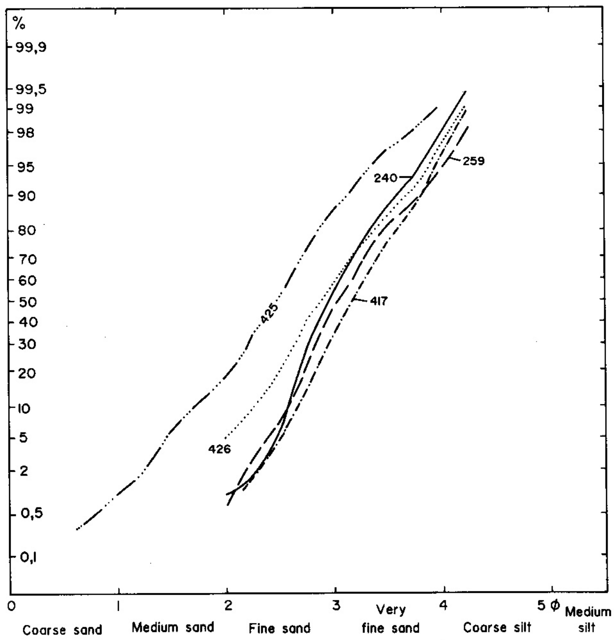


Figure 5.14: Log-probability histogram of grain-size distribution in the Weissberg Member in the central and southern Witvlei Synclinorium (facies (a))

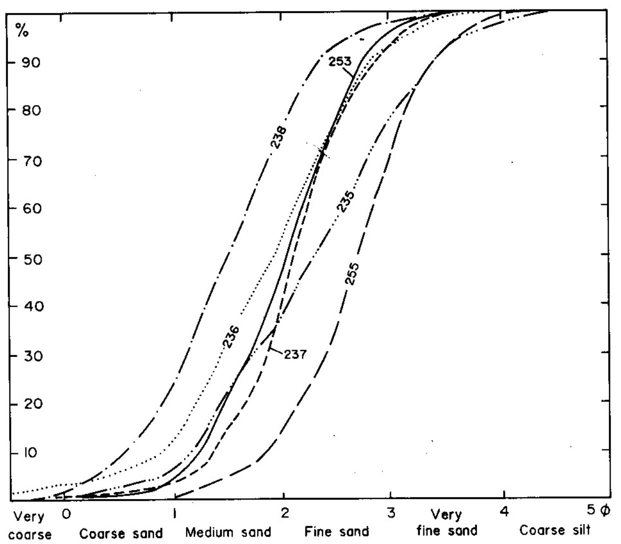


Figure 5.15: Cumulative curves of grain-size distribution in the Weissberg Member in the northern Witvlei Synclinorium (facies (b) and (c))

Most samples of the Okambara Member were taken from the arenaceous middle unit (Tables 5.1 and 5.2). Minor differences were noticed between the arenaceous middle unit of the carbonate-dominated facies in the Witvlei Synclinorium and the quartzitic facies in the southern Gobabis Synclinorium. The latter (sample 414) is devoid of a carbonate matrix, it also has better sorting and contains about 10% clastic ore grains which are either irregularly distributed or enriched in seams that accentuate crossbedding.

In contrast to the Bildah Member, the overlying La Fraque and Okambara Members do not become coarser grained toward the Witvlei Ridge. Instead, the La Fraque Member consists of mudstone and siltstone with up to 40% detrital mica partly altered to chlorite. Framework grains are concentrated in layers up to a few millimetres thick and are matrix-supported. At Witvlei, the La Fraque Member consists of very fine sandstone to coarse siltstone which is mostly matrix-supported, slightly calcareous and contains much hematite and a little mica.

5.1.6 Dabis Formation

5.1.6.1 Weissberg Member

Facies (a) in the southern and central Witvlei Synclinorium consists of mature or supermature quartz arenite (orthoquartzite): well-sorted, fine- to very fine-grained sandstone (Table 5.2) with generally less than 5% feldspar and rock fragments (Table 5.1). The few grains with recognisable original shapes are rounded to well-rounded. Close grouping and the steep slope of the cumulative curves indicate good sorting and uniform grain sizes, whereas their short tails show a negligible percentage of coarser and finer grain classes (Figs 5.13 and 5.14).

Facies (b) which prevails in the northern portion of the Witvlei Synclinorium and the Gobabis Synclinorium, differs from facies (a) in that it is moderately sorted and submature. According to the feldspar content of 5 % to 14%, the sandstones are subarkoses. The less mature and less homogeneous nature of the sediments, compared with facies (a), is revealed by flatter cumulative curves, while their tails indicate a considerable addition of coarser and finer

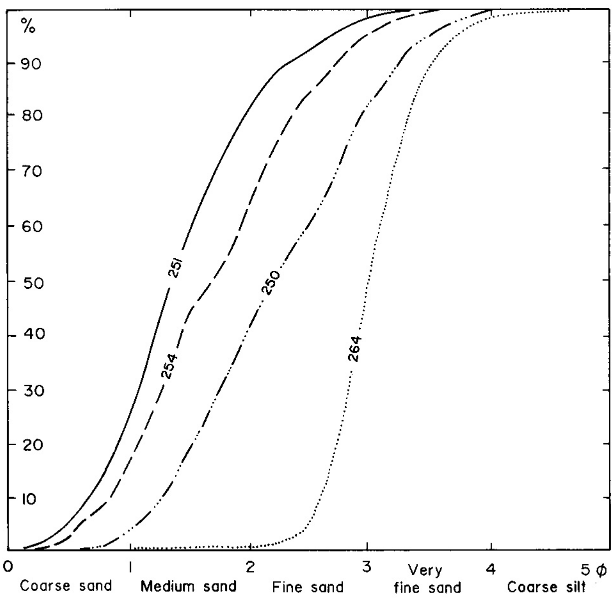


Figure 5.16: Cumulative curves of grain-size distribution in the Weissberg Member in the Gobabis Synclinorium (facies (b))

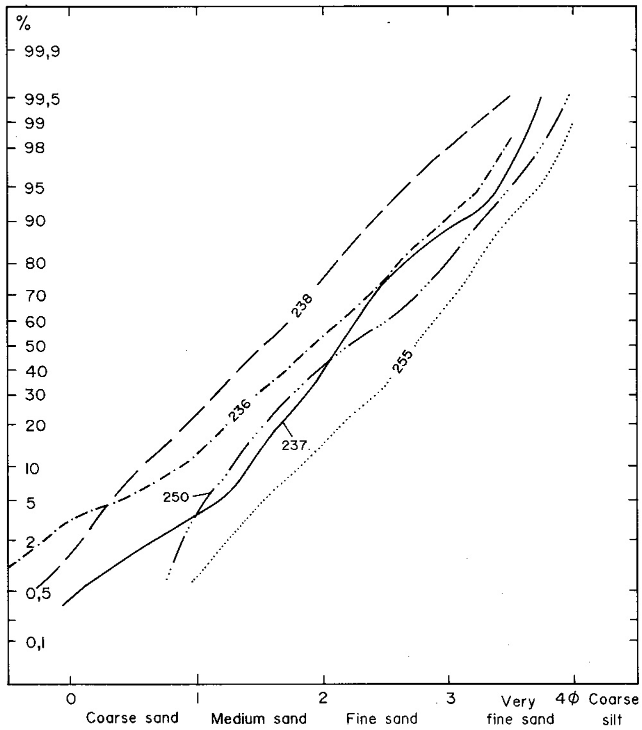


Figure 5.17: Log-probability histogram of grain-size distribution in the Weissberg Member in the northern Witvlei and the Gobabis Synclinoria (facies (b))

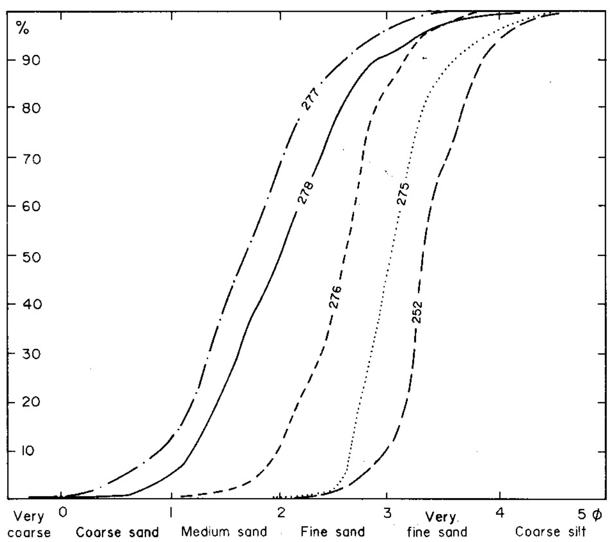


Figure 5.18: Cumulative curves of grain-size distribution in the Zenana Member. Samples 277 and 278 are from a fan deposit in the northern Witvlei Synclinorium (see Figure 4.9)

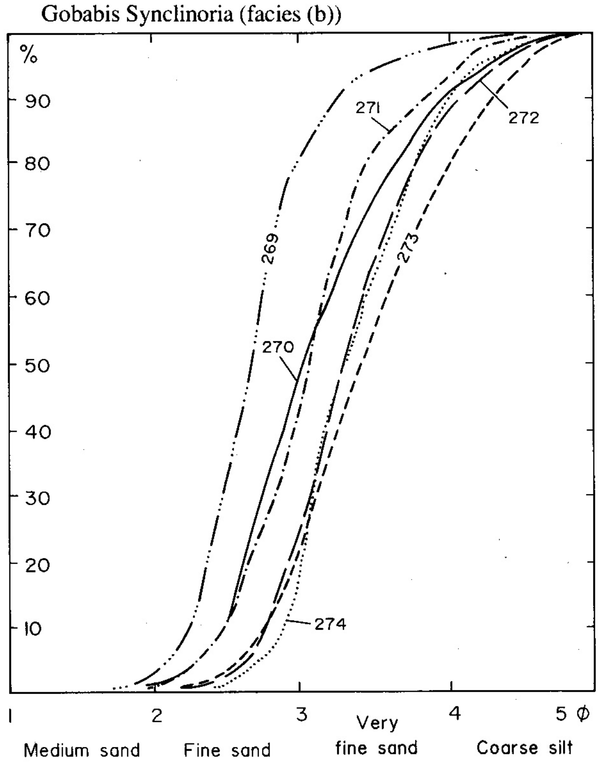


Figure 5.19: Cumulative curves of grain-size distribution in the Grünental Member in the Gobabis Synclinorium (facies (b))

size classes (Figs. 5.15 to 5.17). Several of the cumulative frequency curves and the corresponding log-probability graphs (Fig. 5.17) show bi- or plymodal grain-size curves (samples 235, 236, 237, 250, 254).

In the northern Witvlei Synclinorium (areas 2218 BA and BC) the arenite contains a strikingly high concentration of topaz (Table 5.1, samples 232, 236, 237, 238, 253). The size of most of the topaz grains falls within the coarser half of the cumulative curves.

The conglomeratic facies (c) of the Weissberg Member in thin section 232 is unsorted, with grain sizes continuous between clasts (maximum 1 cm across) and fine-sandy matrix. Cement, averaging 10 to 15%, consists mainly of hematite which gives the dark colour to the rock; silica cement is subordinate and prevails only in those portions in which hematite is absent. Clasts are composed mainly of monocrystalline quartz with undulatory extinction, aggregates of interlocking quartz crystals up to 4 rum across,

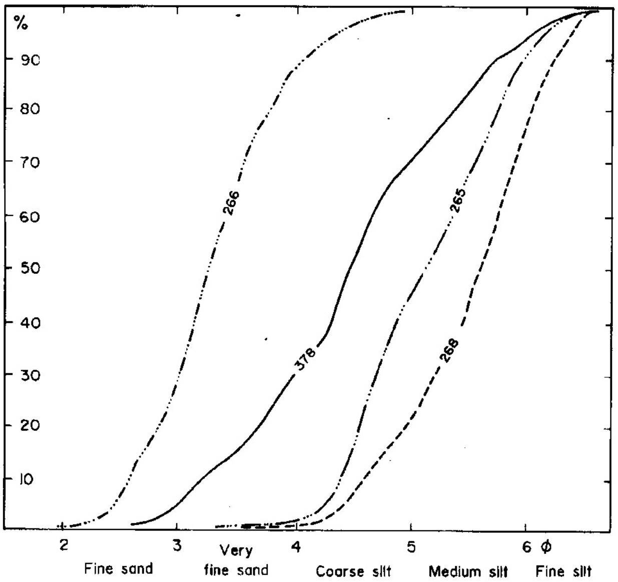


Figure 5.20: Cumulative curves of grain-size distribution in the Grünental Member (facies (a) and transition to facies (b))

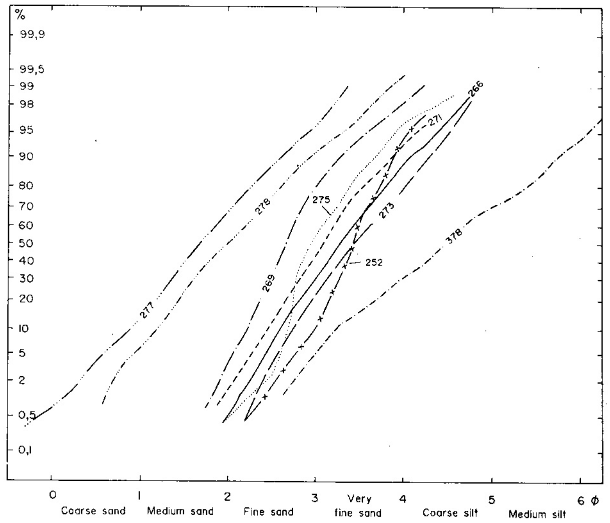


Figure 5.21: Log-probability histogram of grain-size distribution in the Zenana (samples 252, 275, 277, 278) and Grünental Members (samples 266, 269, 271, 273, 387)

and subordinate amounts of chert, unstable lithic fragments and ore. Detrital mica within the matrix amounts to 2%. A quartzitic intercalation within the conglomerate (sample 235, Fig. 5.15) is less well-sorted than the average quartzite of facies (b); a bimodal grain-size distribution could be observed. The rock is grain-supported and is cemented mainly by silica with subordinate hematite. The latter also forms detrital grains. Thin dark seams (dust lines) within quartz grains show that the larger grains were subrounded to rounded.

5.1.7 Zaris Formation

5.1.7.1 Zenana Member

The petrographic composition and texture of the arenaceous upper unit of this member turned out to be very similar to facies (a) of the Weissberg Member (compare samples 252,275 and 276 of Figs 5.18 and 5.21 with Figs 5.13 and 5.14 and corresponding samples of Tables 5.1 and 5.2). This can even be found in regions where the Weissberg Member itself is developed in facies (b). The cumulative curves of the Zenana samples indicate a slightly better sorting than graphs of facies (a) of the Weissberg Member. The grains (where their outlines can be recognised) are rounded to well rounded, and texturally the rock can be classified as supermature.

At the same stratigraphic level, a succession of massive quartzite in the northern corner of Mahagi 302 consists of a moderately sorted subarkose (samples 277 and 278, Fig. 5.18, Table 5.2). As in samples of the Weissberg Member from this region a concentration of topaz occurs (Table 5.1).

5.1.7.2 Grünental Member

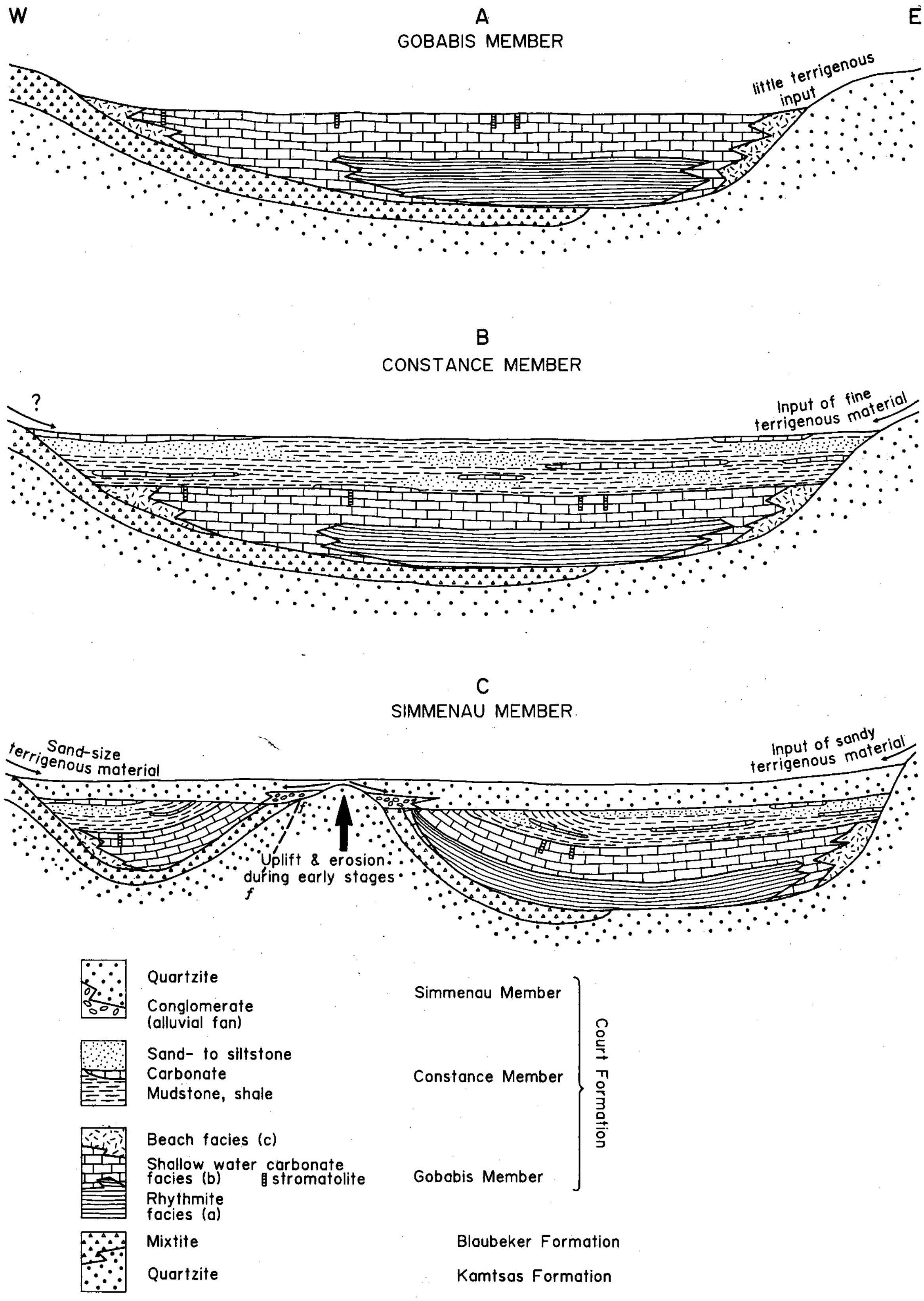
Most rocks from the Grünental Member have a relatively high proportion of matrix (generally about 15%, Table 5.1). They contain abundant phyllosilicates, a large portion of which are of secondary origin (chlorite, sericite). A high percentage of detrital micas (biotite) indicates that the matrix is weathered mica schist, and at least part of it should be considered a pseudomatrix (Plate 5.4). Layering of the arenaceous rocks is accentuated by laminae enriched in

Figure 5.22 (right): Schematic east-west section showing the development of the Court Formation. The Tahiti Member at the base has been omitted. Not to scale.

A - Gobabis Member: Wide and relatively shallow basin with deposition of rhythmite (laminated carbonate) in the centre under anoxic conditions below wave base (facies (a)); unlaminated, thick-layered dark carbonate in shallow water (reworked by wave action) (facies (b)); and a narrow rim of beach facies with reworked bedrocks, flat-pebble conglomerate and coated grains (facies (c)). Except for facies (c) which carries locally derived material, the sediments are devoid of terrigenous detritus.

B - Constance Member: Fine-grained terrigenous material (mud, silt) with subordinate carbonate layers which was deposited in a lagoonal or flood-plain environment.

C - Simmenau Member: An uplift, possibly combined with faulting, during early Simmenau times subdivided the Court basin into a smaller western and a larger eastern portion. The axis of the uplift coincides approximately with the Nina Anticline (Fig. 1.2). On the resulting elevated area erosion cut down into the Nosib Group in the southwestern part of the study area, while near Gobabis only the upper Court Formation was affected, indicating that the uplift decreased towards the northeast. The eroded material was deposited as an alluvial fan conglomerate. An uplift of the highlands bordering the basin to the west intensified erosion there, producing mainly sand-sized material, part of which was transported by rivers into the Simmenau basin during the upper stage of the member and covered most of the uplifted area


W
A
E
GOBABIS MEMBER
little terrigenous input
B
CONSTANCE MEMBER
?
Input of fine terrigenous material
C
SIMMENAU MEMBER
Sand-size terrigenous material
Input of sandy terrigenous material
f
Uplift & erosion during early stages
f
Quartzite
Conglomerate (alluvial fan)
Simmenau Member
Sand- to siltstone
Carbonate
Mudstone, shale
Constance Member
Court Formation
Beach facies (c)
Shallow water carbonate facies (b)
stromatolite
Rhythmite facies (a)
Gobabis Member
Mixtite
Blaubeker Formation
Quartzite
Kamtsas Formation

ore grains, accumulation and orientation of detrital mica flakes, and by thin, irregular laminae and lenses of secondary ore (hematite?).

The arenaceous rocks consist of very fine sand and silt and are subarkoses and arkosic wackes (Table 5.2); sorting is moderate to good and the cumulative curves of the sand-sized specimens resemble those of facies (a) of the Weissberg Member (Figs 5.19, 5.20 and 5.21). Where roundness can be observed it has reached an advanced stage. Samples lacking or having only very little matrix are siliceous-cemented (e.g. 269). Most of the samples were taken from the Black Nossob River valley, where facies (a) and (b) of the Grünental Member interfinger. The only sample from the Witvlei Synclinorium (facies (a), sample 378) is finer grained, less well-sorted and has a higher proportion of matrix, whereas sample 269, the easternmost sample taken from facies (b), contrasts markedly with sample 378 (see Tables 5.1 and 5.2). In thin section, sample 269 shows pressure solution and abundant overgrowth with secondary silica which has caused a merging of neigh-bouring grains, elongation normal to load and layering, straight grain contacts, suturing and even interpenetration of grains. These diagenetic changes obscure the original dimensions of the grains and may have caused the shift to a coarser grain size in the log-probability curve in Fig. 5.21.

5.1.8 Interpretation of data from clastic rocks

5.1.8.1 General

Interpreting sedimentary environments by means of petrographic composition and grain-size distribution alone is ambiguous because the effect of depositional processes depends on parameters which are not necessarily specific for one sedimentary regime only. The most important factors influencing the nature of a clastic sediment are (Reineck and Singh, 1973; Pettijohn *et al.*, 1973):

- type of source rocks;
- distance of transport;
- hydrodynamic factors active during transport;
- post-depositional changes.

Generally grains <1 mm across (about +3.25 Φ) are kept in suspension, grains up to 1.0 mm (0.0 Φ) are transported by saltation, whereas still larger ones move by rolling and sliding (Reineck and Singh, 1973). Changing conditions of transport are especially evident in the case of tidal sediments with reversing flow (Allen, 1984). By this mechanism fine silt and clay (suspension transport) are separated from the coarser grains transported as bed load.

As shown in Table 5.2, the mean grain size of matrix -rich samples is generally <0.12 to 0.09 mm (+3.1 Φ to +3.5 Φ) which conforms with the limits of suspension transport as mentioned above. From this one can infer that the presence of matrix, together with detrital mica, is controlled by suspension transport. No linear relationship, however, could be found between mean size of framework grains and the percentage of matrix and/or detrital mica of a sediment in the samples from the study area.

The clastic sediments of the Constance, La Fraque, Okambara and Grünental Members were mainly transported in suspension, whereas those of the other members as bed load. The cumulative curves show, however, that the modes of transport are not uniform and therefore the grain size classes are never completely separated. This is apparent in the case of both samples from the Doornpoort Formation. The probability curve of sample 434 identifies two grain-size classes, possibly as a result of various flow regimes and the modes of transport (e.g. short floods or meandering rivers) or of different source areas.

From compositional and granulometric data it can be assumed that all clastic sediments, including those of the Doornpoort and Kamtsas Formations, are multicyclic, considering, for example, the paucity of unstable framework minerals and the very limited suite of heavy minerals, of which only extremely stable ones like rutile, zircon, garnet and locally topaz, are present. Furthermore, multicyclicity is indicated by the advanced degree of maturity of the siliciclastic sediments. Comparing the Q/F/R diagrams (Figs. 5.1a-c) one notices that on average the content of feldspar and rock fragments of the progressively younger sediments decreases (from feldspar-rich in the case of the Doornpoort and Kamtsas samples to very quartz-rich in the case of the Witvlei and Nama sediments). This indicates increasing maturity and sorting with decreasing age of the sediments and is either due to longer transport (implying transgression), or to deposition in an environment in which former depositional processes were renewed, e.g. repeated reworking onshore or in the tidal zone.

5.1.8.2 Tahiti Member

The Q/F/R values of most samples from the Tahiti Member group together with the Kamtsas samples (Fig 5.1) which is strongly indicative of derivation of this lowest Witvlei unit from the underlying Kamtsas sediments.

5.1.8.3 Simmenau Member

The relative high percentage of rock fragments in the Simmenau Member indicates little alteration of the unstable components, possibly because of a short transport distance. Most cumulative curves of the Simmenau Member coincide with those of the Kamtsas Formation, suggesting that Kamtsas rocks were the source of the Simmenau quartzite. Considering that several thousand metres of Kamtsas quartzite must have been removed prior to the deposition of the Buschmannsklippe Formation along the western margin of the Witvlei Synclinorium, where the Bildah Member immediately overlies Doornpoort Formation, one can assume that most of the arenaceous sediments of the Simmenau Member were derived from this area. The Simmenau quartzite is probably also derived from the uppermost part of the Doornpoort Formation. The latter assumption would explain the in homogeneity of sample 174, the probability curve of which shows that it is composed of three grain-size classes with a prominent portion of silt-sized material (Fig. 5.9).

5.1.8.4 Bildah Member

A very limited supply of sediment from the Witvlei Ridge into the basin during lower Buschmannsklippe time is manifested by thin arenaceous intercalations along the basin margin. The source of the material was the Kamtsas

- Formation which in this region is a fine- to coarse-grained, pebbly quartzite. Sorting was significant over a short distance of a few hundred metres. Conglomeratic components are found only in a narrow zone surrounding the ridge, indicating the position of the shoreline; further away from the ridge the grain sizes decrease from coarse to fine.

An even more restricted provenance is inferred for the clastic intercalation in the Bildah Member on Sachsenwald 940. This contains grains of basic lava which were derived from volcanics. Volcanic rocks of the Marienhof Formation at present outcrop only a few hundred metres to the west. This implies that Marienhof strata were exposed during the deposition of the Buschmannsklippe Formation and could have been transgressed by it.

5.1.8.5 Constance, La Fraque and Okambara Members

Cumulative curves for the Constance and La Fraque Members (Figs 5.6 and 5.11) indicate that the grains were fine enough to be transported mainly as suspended load. The units overlie mainly chemical sediments and are interpreted as having accumulated during incipient clastic deposition when erosion of the source area and/or transport into the basin were limited. Such conditions are also indicated by the continuation of carbonate .accumulation during these times.

No regional trend in the size variation of siliciclastic grains in the Constance Member has been detected (Fig. 5.6), possibly only because of the very restricted occurrence of the member in the study area.

The clastic components of the Okambara Member are intermediate in position between those of the La Fraque and the Weissberg Members (Fig. 5.11). The La Fraque and Okambara Members can be considered to be part of one continuous sedimentary process. Samples 413, 215 and 228 (Fig. 5.11), from the quartzitic middle unit of the Okambara Member, show that the grain sizes increase slightly from the central to the southern portions of the Witvlei Synclinorium. This unit thickens in the same direction suggesting a provenance area to the southwest. Along the margin of the Witvlei Ridge the grain sizes of the La Fraque and Okambara deposits are finer than within the basin, e.g. the La Fraque Member consists of only silt- and mudstone. This means that the Witvlei Ridge did not contribute to the sediments of both members. Similar conditions can be deduced for the Gobabis Synclinorium. In the southwestern portion of the Gobabis Synclinorium both members have a finer grain size than the Kamtsas Formation (Table 5.2) and do not become finer offshore. On the contrary, sample 414 reflects an onshore fining. These trends make an eastern source of the clastic sediments of the La Fraque and Okambara Members unlikely and exclude the arenites of the Kamtsas Formation in that region as a supplier of sediments.

5.1.8.6 Weissberg Member

The maturity of the arenaceous constituents of the Weissberg Member increases from the northern Witvlei Synclinorium toward the Witvlei Ridge; sorting is better (Table 5.2) and grain size decreases (Fig. 5.15). The conglomerate in the Kehoro area (facies (c), Fig. 4.10) is an indication that the provenance area was situated nearby and to the north. The Witvlei Ridge formed a barrier, preventing coarser grains from being transported further south. Likewise, the grain size of sediments supplied from the east into the basin south of the Witvlei Ridge was restricted to finer fractions as can be inferred from the cumulative curves from the Gobabis Synclinorium which are similar to those from the northern Witvlei Synclinorium (cf. Figs 5.15 and 5.16). Only the fine-grained arenaceous fraction could be transported into the area of the central and southern Witvlei Synclinorium (Fig. 5.13).

5.1.8.7 Zenana Member

Orthoquartzite of the upper unit of the Zenana Member, which is a reappearance of the Weissberg facies, yielded uniform cumulative curves throughout most of the investigated area (Fig. 5.18), even in those regions where the Weissberg Member is less mature, indicating a longer transport distance. Only in the northeast (in area 2218 BA, see Fig. 4.11), is a change to onshore conditions perceptible, but even here the maturity has increased (samples 277 and 278 of Fig. 5.18), compared with the Kehoro conglomerate of the Weissberg Member (facies (c)) in the same area.

Topaz concentrations in the Weissberg and Zenana Members of the Kehoro area (2218 BA) might have formed in consequence to the enrichment of this heavy mineral in the Kamtsas Formation of the Gobabis area (Table 5.1). The local topaz content indicates that Kamtsas sediments, probably derived from the southeast, were redeposited during Dabis and Zaris times.

5.1.8.8 Grünental Member

Orthoquartzite of the Grünental Member occurs in the eastern portion of the study area (sample 269, Fig. 5.19). From there to the west grain sizes generally decrease and the amount of matrix increases to form a matrix-rich siltstone (sample 378, Fig. 5.20). In the region of the Black Nossob River, matrix-bearing, fine-grained sandstone to siltstone prevails. The coarser grain classes, transported as bed load, settled first, while the finer ones were carried as suspension further westward into the basin.

Sample No.	Locality Farm	Area	Lithofacies	CA++ (%)	Mg++ (%)	non-carb.
Gobabis Member						
125	Court 32	2318 BA	indistinctly laminated carb.	100	0	2
126	Court 32	2318 BA	indistinctly laminated carb.	100	0	-
127	Josephine 226	2218 CC	indistinctly laminated carb.	52	48	-
128	Tahiti 61	2218 DC	laminated carb.	100 (56)	0 (44)	8
129	Schwarzwald 344	2219 CB	sandy carb. with intraclasts	56	44	+
130	Schwarzwald 344	2219 CB	sandy carb. with intraclasts	52	48	+
131	Gobabis	2218 BD	sandy carb. with intraclasts	56	44	+
132	Hener 345	2219 CB	indistinctly laminated carb.	56	44	
133	Gobabis	2218 BD	unlaminated carbonate	48	52	
134	Gobabis	2218 BD	laminated carbonate	55	45	
135	Schwarzwald 344	2219 CB	sandy carb. with intraclasts	56	44	+
136	Klippiespan 238	2218 CC	unlaminated carbonate	100	0	
137	Klippiespan 238	2218 CC	flat-pebble conglomerate	100	0	
138	Klippiespan 238	2218 CC	unlaminated carbonate	100	0	10
140	Klippiespan 238	2218 CC	flat-pebble conglomerate	100 (56)	0 (44)	
400	Tahiti 61	2218 DC	laminated carbonate	100	0	12
460	Josephine 226	2218 CA	algal laminated carbonate	52	48	-
464	Frank 221	2218 CA	sandy to pebbly carbonate	52	48	25
Constance Member						
141	Klippiespan 238	2218 CC	fine sandstone, calcareous	100	0	65
142	Court 32	2318 BA	slightly calcareous siltstone	56	44	++
143	Frank 221	2218 CA	sandy carbonate	52	48	+
144	Constance 230	2218 CC	oolitic carbonate	52	48	0
145	Tahiti 61	2218 DC	calcareous mudstone	53	47	35
146	Kanabis 54	2218 DC	carbonate	55	45	-
147	Court 32	2318 BA	dark carbonate layer	55	45	15
148	Court 32	2318 BA	dark carbonate layer	58	42	11
150	Aais I 31	2318 BA	slightly calcareous sandstone	100	0	++
151	Tahiti 61	2218 DC	oolite (uppermost portion)	50	50	0
153	Tahiti 61	2218 DC	oolite (uppermost portion)	52	48	-
154	Tahiti 61	2218 DC	carbonate (uppermost portion)	52	48	-
458	Constance 230	2218 CC	carbonate (uppermost portion)	52	48	-
Simmenau Member						
159	Schönborn 79F	2218 DC	dark carbonate, stromatolite	52	48	0
Bildah Member						
175	Eilenreide 53	2218 DC	laminated white carbonate	54	48	15
177	Schönborn 79F	2218 DC	laminated light-coloured carb.	52	48	-
178	Gobabis	2218 BD	laminated light-coloured carb.	55	45	-
179	Okambara 219	2218 CA	laminated pink carbonate	52	48	-
180	Okombuka 218	2218 CA	laminated light-coloured carb.	52	48	
181	Held 84	2218 AD	calcareous sandstone	100 (52)	0 (48)	65
182	Held 84	2218 AD	calcareous sandstone	52	48	40
183	Sachsenwald 940	2218 BA	impure carbonate	52	48	+
184	Vergelegen 162	2218 BB	laminated reddish carbonate	52	48	0
185	Nudom 161	2218 BA	laminated pink carbonate	52	48	0
187	Orochevley 216	2218 CA	laminated light-coloured carb.	55	45	-
188	Okambara 219	2218 CA	laminated pink carbonate	52	48	-
191	Schönborn 79F	2218 DC	laminated light-coloured carb.	52	48	-
192	Keizaub 68	2218 DD	laminated light-coloured carb.	52	48	-
193	Styria 52	2218 DC	laminated pink carbonate	52	48	0
194	Okambara 219	2218 CA	pink, thin carb. (uppermost)	100	0	6
195	Orochevley 216	2218 CA	laminated, light-coloured carb.	52	48	-
196	Orochevley 216	2218 CA	laminated, light-coloured carb.	52	48	-

Table 5.3 (continued on following two pages) :

Sample No.	Locality Farm	Area	Lithofacies	CA++ (%)	Mg++ (%)	non-carb.
381	Okambara 219	2218 CA	laminated pink carbonate	52 (100)	48 (0)	-
383	Okasewa NW 120	2218 AD	recrystallised, light-grey carb.	52	48	-
396	Kehoro 183	2218 BA	laminated, stylolitic carbonate	56 (100)	44 (0)	1
397	Held 84	2218 AD	sandy, massive carbonate	52 (100)	48 (0)	25
437	Kehoro 183	2218 BA	pink, laminated carbonate	52	48	0
451	Kehoro 183	2218 BA	pink, laminated carbonate	55	45	0
452	Kehoro 183	2218 BA	laminated, stylolitic carbonate	52 (100)	48 (0)	0
453	Kehoro 183	2218 BA	laminated, stylolitic carbonate	52 (100)	48 (0)	0
462	Okombuka 218	2218 CA	sandy to pebbly carbonate	53	47	20
463	Okombuka 218	2218 CA	sandy carbonate	53	47	10
466	Frank 221	2218 CA	stromatolitic carbonate	52	48	0
La Fraque Member						
197	Okambara 219	2218 CA	silty carbonate	100	0	+
200	Okambara 219	2218 CA	fine calcareous sandstone	100	0	45
201	Witvlei	2218 AD	calcareous siltstone	55	45	++
202	Poortje 287	2219 DA	reddish, homogen. mudstone	54	46	30
357	Okambara 219	2218 CA	calcareous siltstone	100	0	60
359	Okambara 219	2218 CA	silty carb. to calc. siltstone	100	0	60
360	Okambara 219	2218 CA	silty carbonate	100	0	+
364	Okambara 219	2218 CA	indistinctly laminated carb.	100	0	-
370	Bildah 220	2218 CA	silty to fine-sandy carbonate	100	0	45
371	Bildah 220	2218 CA	silty to muddy carbonate	100	0	+
372	Bildah 220	2218 CA	calcareous siltstone	100 (57)	0 (43)	++
379	Okambara 219	2218 CA	silty carbonate	100	0	+
419	Doreen 227	2218 CC	sedim. breccia, sandy carb.	100	0	35
420	Doreen 227	2218 CC	pseudom. after aragonite	100	0	0
Okambara Member						
undifferentiated						
207	Kehoro S 939	2218 BA	pink,massive carbonate	54 (100)	46 (0)	15
208	Kehoro S 939	2218 BA	fine-sandy carbonate	100 (52)	0 (48)	+
229	Nautabis 268	2317 BB	algal laminated carbonate	52	48	-
230	Springboklaagte 20	2317 BB	slightly oolitic carbonate	100 (58)	0 (42)	20
361	Held 84	2218 AD	fine, calcareous sandstone	52	48	40
362	Held 84	2218 AD	sandy carbonate	55	45	+
382	Kehoro N 185	2218 BA	massive carbonate	52	48	12
386	Kehoro N 185	2218 BA	laminated carbonate	55	45	-
407	Schönborn 79 F	2218 DC	slightly laminated carbonate	56 (100)	44 (0)	0
lower unit						
206	Frank 221	2218 CA	silty, algal laminated carbon.	100	0	25
213	Otjimbondona 225	2218 CC	silty, algal laminated carbon.	100 (61)	0 (39)	20
214	Frank 221	2218 CA	calcareous sandstone	100 (56)	0 (44)	30
223	Okambara 219	2218 CA	irregularly laminated carbon.	100 (56)	0 (44)	10
392	Constance 230	2218 CC	edgewise conglomerate	56 (100)	44 (0)	20
395	Constance 230	2218 CC	edgewise conglomerate	55	45	5
427	Okambara 219	2218 CA	flat-pebble conglomerate	100 (58)	0 (42)	10
436	Okambara 219	2218 CA	edgewise conglomerate	100	0	20
middle unit						
217	Okombuka 218	2218 CA	sandy carbonate	55	45	35
219	Bildah 220	2218 CA	slightly calcareous sandstone	52	48	++

Table 5.3 (left, above and overleaf): Calcium and magnesium content of carbonate rocks. Data in brackets mean that a minor percentage of this carbonate is present apart from a major percentage of the other mineral. If percentage of non-carbonates was estimated (not determined by "Karbonat-Bombe"), symbols mean: - less than 30%; + between 30 and 60%; ++ more than 60%

Sample No.	Locality Farm	Area	Lithofacies	CA++ (%)	Mg++ (%)	non-carb.
upper unit						
203	Otjimbondona 225	2218 CC	sandy carbonate	52	48	+
204	Okambara 219	2218 CA	sandy stromatolite with chert	50	50	20
205	Okambara 219	2218 CA	stromatolite	52	48	-
209	Otjimbondona 225	2218 CC	reworked silty carbonate	52	48	30
211	Nautabis 268	2317 BB	massive carbonate	52	48	0
212	Nautabis 268	2317 BB	flat-pebble conglomerate	55	45	-
215	Kowas 233	2218 CC	flat-pebble conglomerate	52	48	45
216	Okambara 219	2218 CA	sandy flat-pebble conglom.	52	48	+
218	Kowas 233	2218 CC	sandy flat -pebble conglom.	52	48	+
224	Okambara 219	2218 CA	flat-pebble conglomerate	52	48	-
226	Springboklaagte 20	2317 BB	sandy flat-pebble conglom.	100 (56)	0 (44)	+
256	Springboklaagte 20	2317 BB	oolite	56	44	0
260	Nautabis 268	2317 BB	sandy carbonate	52	48	+
363	Okambara 219	2218 CA	sandy flat pebble-conglom.	55	45	+
365	Okambara 219	2218 CA	pelletoidal carbonate	50	50	20
366	Okambara 219	2218 CA	slightly sandy carbonate	52	48	+
367	Okambara 219	2218 CA	massive carbonate	52	48	-
369	Okambara 219	2218 CA	massive carbonate	52	48	-
373	Okambara 219	2218 CA	pelletoidal carb. with chert	56	44	30
374	Okambara 219	2218 CA	sandy carb. with silica spherul.	52	48	+
387	Okambara 219	2218 CA	stromatolite	52	48	-
388	Okambara 219	2218 CA	stromatolite	52	48	30
389	Okambara 219	2218 CA	algal laminated carbonate	52	48	-
390	Okambara 219	2218 CA	stromatolite	52	48	+
465	Otjimbondona 225	2218 CC	algal mound	55	45	-
Zenana Member						
231	Nautabis 268	2317 BB	massive dark carbonate	52	48	-
258	Nautabis 268	2317 BB	massive dark carbonate	56	44	-
288	Kuduberg 60	2218 CC	thin-layered dark carbonate	52 (100)	48 (0)	5
290	Okasewa 102	2218 AD	weakly lamin. dark carbonate	53 (100)	47 (0)	0
295	Styria 52	2218 DC	irregul. bedded dark carb.	55 (100)	45 (0)	2
Grünental Member						
265	Blumenau 58	2218 DD	calcareous siltstone	100	0	++
267	Styria 52	2218 DC	dark carbonate	100	0	0
268	Karossewie 72	2218 DB	calcareous siltstone	52	48	++
279	Grünental 151	2218 AD	carbonate with spherulites	100	0	12
280	Grünental 151	2218 AD	carbonate with spherulites	100	0	-
281	Grünental 151	2218 AD	carbonate with spherulites	100	0	-
282	Grünental 151	2218 AD	carbonate with spherulites	100	0	10
283	Blumenau 58	2218 DD	medium to dark carbonate	100	0	15
284	Grünental 151	2218 CA	carbonate with spherulites	100	0	-
293	Blumenau 58	2218 AD	impure dark carbonate	100	0	20
294	Suliman 215	2218 AD	massive, dark carbonate	100 (57)	0 (43)	0
358	Held 84	2218 AD	light coloured carbonate	100	0	-
375	Grünental 151	2218 AD	carbonate with spherulites	100	0	-
376	Grünental 151	2218 AD	carbonate with spherulites	100	0	-
377	Grünental 151	2218 AD	carbonate with spherulites	100	0	-
384	Suliman 215	2218 CA	massive, light to dark carb.	100	0	0
Omkyk Member						
285	Tigerpforte 59	2317 BC	homogeneous dark carbonate	100	0	0
286	Tigerpforte 59	2317 BC	indistinctly laminated carb.	100	0	-
287	Tigerpforte 59	2317 BC	dark carbonate	100	0	-
291	Graswêreld272	2317 BB	faintly layered, dark carbonate	58 (100)	42 (0)	0
292	Tigerpforte 59	2317BC	lam. dark carb. with intraclasts	100	0	-

Table 5.3 continued

5.2 CARBONATE ROCKS

5.2.1 Calcium and magnesium contents

Altogether 150 samples of carbonate and carbonate-bearing rocks were analysed for calcium and magnesium contents by means of X-ray diffraction. The Ca^{2+}/Mg^{2+} ratio within a carbonate crystal influences the X-ray reflex: Calcite with 100% Ca^{2+} has a d-value of 3.05 A, whereas dolomite with 50% Ca^{2+} and 50% Mg^{2+} has a d-value of 2.886 A. Accordingly, the angle of the main reflex (of the 104 crystal plane) increases from 29.4° to 30.9°. The relation between spacing of the crystal planes and the ratios of the two different cations is linear and allows for a rapid and simple determination of the calcite and dolomite percentages from $Ca^{100}Mg^{O}$ to $Ca^{50}Mg^{50}$) (Richter, 1974; further references in Friedman and Sanders, 1967, p. 276).

The results (Table 5.3) show that only the end-members of the continuous series calcite-dolomite (see Bissell and Chilingar, 1967) occur, i.e. either calcite or nearly pure dolomite. Several of the XRD diagrams prove the coexistence of calcite and dolomite. If one of these, as is usually the case, forms a minor constituent, this is indicated by brackets in Table 5.3. Coexistence of the two minerals is explained by:

- synsedimentary mixing of two different carbonate types, e.g. in flat-pebble and edgewise conglomerates, with clasts of a composition different from that of the matrix;
- diagenetic filling of open-space structures (see Chilingar *et al.*, 1967), e.g. most of the sheet cracks and tubular structures in the Bildah dolomite are filled with calcite;
- selective dolomitisation which may be influenced by primary sedimentary features, grain-size distribution and original composition (Friedman and Sanders, 1967, p. 297 ff.);
- partial dedolomitisation.

The percentage of non-carbonates in Table 5.3 was determined either with the “Karbonat-Bombe” (Müller and Gastner, 1971) - if exact values are given - or estimated from thin sections, polished plates and the intensity of carbonate peaks in XRD-diagrams. Use of the “Karbonat-Bombe” for distinguishing between calcite and dolomite, according to the method proposed by Müller and Gastner (1971), failed in the case of the dolomitic samples for which a high but inconstant calcite content (up to 50%) was wrongly indicated, most probably due to the immediate reaction of dolostone with concentrated HCI.

From Table 5.3 certain trends in the calcite/dolomite relation in the stratigraphic subunits of the Witvlei and Nama Groups become obvious:

Limestone prevails in the southern portion of the basin in which the *Gobabis Member* was deposited, while in its northern part both laminated and unlaminated facies consist of dolomite. No relation between depositional conditions and carbonate constituents is evident, with the possible exception of the marginal areas (samples 127 and 132). Generally the carbonate contains few impurities.

Dolomite makes up most of the carbonate of the *Constance Member*.

The only carbonate sample from the *Simmenau Member* is a stromatolitic dolomite.

Dolomite also prevails throughout the *Bildah Member*; it is poor in, or even devoid of, impurities. Samples 181 and 397 consist of calcite together with dolomite and might reflect dedolomitisation. Sample 194, a slightly clayey limestone, shows that in the uppermost, well-layered Bildah Member, calcite rather than dolomites prevails.

Calcite persists throughout the *La Fraque Member* into the lower unit of the *Okambara Member*. Most of these rocks have strong admixtures of fine siliciclastic to muddy components. Sample 202, a dolomitic mudstone from the Netso Syncline, could reflect specific conditions along the margin of the basin.

The dolomite content of the samples from the lower unit of the Okambara Member is mainly associated with the carbonate clasts of conglomerates and indicates deposition of reworked dolomite crusts in a calcite-precipitating environment. Large dolomite rhombs with several growth phases float in a calcitic clayey matrix and are of late diagenetic origin. The carbonate rocks of the middle and upper units are composed of dolomite, often with substantial clastic impurities. Samples 229 and 230 from the southern-most Witvlei Synclinorium, where the succession is disturbed by faulting, consist of dolomite and mainly calcite, indicating that the first sample belongs most probably to the upper and the second one to the lower unit of the Okambara Member. In a thin section of sample 230, dolomite rhombs replace cores and parts of the cortices of ooids and also occur in the cement, this indicating their diagenetic origin.

Samples of the Okambara Member from areas where the triple subdivision has not been developed (Kehoro region, Gobabis Synclinorium), consist generally of dolomite.

Carbonate of the *Zenana Member* is comprised mainly of dolomite while similar-looking carbonate rocks from the upper Zaris Formation (*Grünental and Omkyk Members*) are predominantly calcitic.

5.2.2 Microfacies

For a study of composition and texture of carbonate rocks, including the examination of sedimentary structures, one hundred and forty-five thin sections and polished slabs were investigated.

5.2.2.1 Gobabis Member

Five microfacies (Wilson ,1975) could be distinguished:

Microfacies (i) consists of parallel black calcite seams, alternating with pink to light-grey ones. They persist laterally, a few of them showing loop bedding (cf. Engster and Kelts, 1983, Plate 12.3.b). The laminae are 0.06 to 1.0 mm thick and comprise sparite. In thin section, the light laminae appear clear and translucent; the dark ones have a brownish tinge which, however, is faint enough to reveal the outlines of the calcite crystals (Plate 5.5). The matter causing this tinge is a diffuse, amorphous fluff, forming bands with generally sharply defined upper and lower boundaries. The organic content of two samples from Gobabis and Tahiti was estimated at 0.1 % and 0.2% by E.I. Robbins (U.S. Geological Survey, pers. comm., 1985). Other minor constituents mentioned by Robbins are: hematite, up to 5% pyrite, rutile, tourmaline, 1 to 5% clay, little silt. The colour of the dark bands is probably caused by organic matter, clay, or both.

Under polarised light two different kinds of dark laminae were found:

(a) the carbonate crystals are optically disordered and the dark seams transect the crystals without being bound by crystal faces (plate 5.5);

(b) the dark seams, locally together with small portions of the neighbouring light-coloured ones, show uniform extinction over up to 5 mm in length. Tiny crystals within the uniformly aligned bands deviate from this orientation.

Light laminae always consist of optically disordered carbonate crystals. These features might be explained by recrystallisation as a reaction to compressive stress normal to bedding, dependent on the percentage of dark matter.

Laterally type (a) may pass into type (b) and vice versa in the same lamina, or type (a) replaces type (b) for a short distance. Apart from the presence/absence of the dark matter, lamination can be further accentuated by change in crystal size, change of the direction of elongation of the carbonate crystals and by inclusion of hematite seams and lenses.

A thin section from Gobabis shows less regular and distinct lamination than that from Tahiti 61. Thicker layers within a laminite (0.5 to 1.0 cm thick) can be interpreted as storm generated (Tucker, 1983).

If lamination becomes indistinct, individual light-coloured laminae are laterally intermittent or even lenticular whereas dark intercalations swell, and in the end a fairly even distribution of the fluffy matter is reached, as seen in microfacies (ii).

Microfacies (ii) is unlaminated, micritic to sparry dark carbonate that forms thin to massive layers. Both microfacies (i) and (ii) contain up to 2% of silt-sized quartz grains and chert aggregates. It is not possible to distinguish detrital quartz grains from those of diagenetic origin. Angular outlines of grains may be due to abrasion as well as corrosion; aggregates of euhedral quartz indicate a diagenetic origin. As undulatory extinction was observed in quartz forming less than 1-mm-wide veinlets, this is no criterion for the detrital nature of quartz. Detrital mica flakes are rare; dolomite rhombs occur locally.

Microfacies (iii) is defined by sandy carbonate, varying percentages of detrital material and mainly corroded quartz grains the size of which varies from silt to coarse sand. There are also subordinate amounts of orthoclase, plagioclase, chert and quartz aggregates.

Microfacies (iv) is an ooid grainstone (Dunham, 1962) with normally packed, superficially and multiply coated ooids (the latter generally subordinate) up to 1.5 mm across, together with carbonate pellets, chert nodules and superficially coated intraclasts. Micritic cortices of the ooids are distinct from the lighter-coloured sparite of the cores and the drusy cement. Besides carbonate, the cores of ooids may also consist of angular quartz grains or even chert which may be derived from diagenetic replacements of carbonate by silica. Authigenic, fine-crystalline radial quartz aggregates locally replace calcareous cement. Composite and broken ooids indicate phases of reworking and resedimentation. On Tahiti 61 oolitic grainstone with ooids (0.3 mm across) and coated pellets (up to 3 mm across) occurring at the base of the member are made of micritic to sparry carbonate nuclei, surrounded by one to three micritic coatings. The coated grains are lined by fibrous sparite while late-diagenetic drusy sparite fills the interstices. Locally the coated grains are concentrated in layers.

Microfacies (v) is a lithoclastic grainstone, termed an intraformational pebble conglomerate (Plate 4.2), formed by micritic to microsparry laminated clasts in sparry groundmass. Clasts are between 2 mm and 10 cm long, tabular, either imbricated or oriented parallel to bedding. Some small clasts (less than 5 mm long) are superficially coated; larger clasts are slightly deformed, indicating that reworking affected a not-yet-lithified sediment. The clasts are yellowish grey in colour and stand out from the dark-grey matrix.

Microfacies (i) and (ii) combine to build thick successions with transitions between both. Such rocks occur as clasts in microfacies (v) (Plate 4.2). Microfacies (iii), (iv) and (v) occur together in the same individual layer, e.g. on the southern limb of the Netso Syncline where the Gobabis Member includes intraclast-bearing (derived from facies (v)) oolitic (facies (iv)) carbonate with locally up to 50% quartz grains (facies (iii)).

5.2.2.2 Constance Member

Two microfacies could be distinguished:

Microfacies (i) is a micritic, slightly clayey carbonate, forming wavy, discontinuous, irregular brownish-coloured zones from 0.2 to 0.5 mm thick. Intercalated are even thinner (0.05 to 0.20 mm) crinkled layers and lenses of light-coloured, sparry carbonate with silt-sized, corroded quartz, subordinate plagioclase and little muscovite. These detrital minerals are either scattered or form grain-supported lenses within the sparite, but also occur within the micrite. If sparite intercalations are absent, the rock is a brownish mottled, massive micrite. This microfacies forms well-defined dark layers in the lower portion of the member on Court 32 and was also found along the south-eastern Witvlei Synclinorium where it contains a higher clay content, is reddish to dark-grey in colour and consists mainly of sparite.

Microfacies (ii) is a grainstone of generally overpacked oolite. Superficially coated grains of subrounded to irregular shape predominate, contrasting with spherical, multiply coated ooids with up to 10 cortices. They measure up to 1.5 mm across and consist of micritic coats around microsparry to drusy carbonate nuclei. Several ooids have been corroded during transport and the fragments have been newly coated; the nuclei of others consist of several small ooids or pellets, enwrapped in common cortices. Radially structured ooids are subordinate; this may be due to recrystallisation. A seam of fibrous, radially oriented carbonate crystals lines the ooids; interstices have been clogged by a drusy carbonate mosaic. Where in contact with each other, neighbouring coated grains may be flattened (Plate 5.6), showing that they were not yet lithified during compaction of the rock. Nuclei of several ooids were leached out and the moldic cavity has been filled by drusy carbonate (including dolomite rhombs) and chert (plate 5.6). Dolomite rhombs with dust seams indicate repeated growth phases (Chillinar *et al.*, 1979). Corroded quartz grains and euhedral plagioclase crystals occur scattered within the cement and the coated grains. Most of them are authigenic as can be deduced from the fact that they do not only form the nuclei of ooids, but also penetrate into the cortices from within or outside, or they lie within the coating, partly replacing it. A euhedral plagioclase crystal has pierced the cortices of two neighbouring grains, bridging the narrow space between them. This microfacies was recognised in the southern portion of Constance 230 and in the uppermost part of the member in the northwestern portion of Tahiti 61. It was also observed in the southwestern corner of Frank 221 in sandy carbonate. There the rock consists of about 40% detrital grains, up to 1 mm across (mostly quartz, quartz aggregates and chert, subordinate feldspar), and 60% micritic to sparry carbonate, made up mainly of superficially coated carbonate pellets and lumps (intraclasts) up to 0.5 mm across.

5.2.2.3 Simmenau Member

Dolomite is the only carbonate rock within this member displaying indistinct zones, layers and patches of medium-grey micrite to microsparite, alternating with light-grey granular sparite. A few small quartz grains might be authigenic. The individual zones are up to two millimetres thick, badly defined and have no sharp boundaries. Irregular domal structures indicate that the texture may be of algal origin.

Carbonate cement has been found in several thin sections of the quartzite (Table 5.1); this explains structures which appear at many localities on the weathered surface of the rock. They consist of spherical, rather indistinctly bounded zones, measuring up to 2 cm across and coloured darker than the surrounding harder quartzite. Further weathering leads to the formation of cavities. In several cases the inner portion of the sphere is preserved as quartzite and set free by disintegration of a thin quartzite zone (Plate 4.3). In other instances cross sections on weathered surfaces show a central spherical cavity which is surrounded by a parallel ridge, joining the surrounding unweathered quartzite along a shallow depression. These structures stem from the decomposition of sandstone by removal of its cement which is believed to have been carbonate. Initially finely disseminated in the rock, carbonate was diagenetic ally enriched in a concretionary manner, replacing much of the siliceous cement. Later it was more readily removed by weathering than the siliceous cement in the surrounding quartzite.

5.2.2.4 Bildah Member

(i): The *standard microfacies* of this member is a fine crystalline, micritic to fine-sparry, light-coloured dolomite (Table 5.3). Faint banding, caused by different crystal sizes and shades of colour and slight variations of the generally insignificant clay content, form indistinct laminae. The average thickness of 0.1 to 0.3 mm of the laminae may increase when several of them merge. Silt-sized corroded quartz grains are subordinate; they are either irregularly scattered or concentrated in bedding-parallel horizons which are only one grain thick. Most of these quartz grains are detrital, possibly wind-blown (Wilson, 1975), as indicated by their layered disposition, together with rare detrital mica flakes. In most cases the detrital seams are associated with darker laminae or occur along sheet cracks, supporting their interpretation as algal-related sedimentary structures: the lamination originated from algal mats which produced carbonate when covered by water, while during low water their growth was impeded and mud, occasionally silt, was trapped at the surface of the mats. Chert and authigenic quartz aggregates form small patches and bedding -parallel layers which are up to one millimetre thick and consist of fibrous aggregates, oriented perpendicular or parallel to bedding. The latter case is interpreted as the infilling of sheet cracks.

Microfacies (ii): A thin section from the uppermost, thinly bedded portion of the Bildah Member shows features similar to those in the underlying portion, but the rock contains more silt-sized and pelitic clastics, both scattered or enriched in thin layers. The latter, however, are several grains thick and, compared with rocks from the lower portion, indicate restricted or suppressed algal growth. Haematite is concentrated in seams, causing a macroscopically visible banding. Although the XRD analysis of sample 194 identified calcite only (Table 5.3), a few dolomite rhombs were found under the microscope.

Microfacies (iii): On the southern margin of the Witvlei Ridge a grainstone microfacies of sandy carbonate, containing coated grains, was formed under coastal conditions. The carbonate of the ooids, their matrix and that of intermittent layers devoid of ooids, is micrite to fine crystalline sparite. Spherical, multiply coated ooids reach up to 0.7 mm; irregular to elongated, superficially coated ones 2.0 mm; their nuclei consist of either carbonate or siliciclastic particles (quartz, rock fragments). Coated grains abound if the detrital component of the sandy carbonate is below 40%, become less abundant with rising clastic percentage, and are absent in rocks in which siliciclastic detritus exceeds 60%. Fibrous carbonate lines the ooids and many of the quartz grains. Similar sediments have been encountered on Sachsenwald 940, where the carbonate of both cement and cortices is drusily recrystallised, thereby obliterating fabric and outlines of the ooids (Plate 5.3). Dolomite rhombs are plentiful in thin sections from both localities, as is seen in Table 5.3.

Pebbly to fine-sandy carbonate occurring in the western portion of Okombuka 218 (2218 CA) includes carbonate clasts as well as many superficially coated quartz grains and quartzite pebbles; multiply coated grains are absent. Small irregular cavities are lined or filled by banded fibrous dolomite, forming spelaeothems (Plate 5.7).

Microfacies (iv): Oolite, occurring on Orochevley 216, is distinguished from all the other rocks containing coated grains in that:

- it consists mainly of spherical ooids and subordinately of coated fragments;
- multiply coated grains are predominant;
- the ooids are up to 3 mm in size (Plates 4.6, 5.8).

Nuclei consist of corroded quartz grains, quartzite fragments and crushed ooids, lined by up to 15 cortices, but all transitions between superficially and multiply coated grains were found, e.g. larger, less mobile rock fragments were less often coated; a 5.6-mm-long quartzite particle is enveloped by one layer only. The oblong to irregular shape of the ooids is due to:

- the form of the siliciclastic core;
- the irregular shape of fragmented earlier ooids (plate 5.8);
- different thickness of cortices, caused by uneven abrasion during transport, leading to eccentric ooids (plate 5.8);
- coalescence of two or more small ooids (Plate 5.8).

Many ooid fragments have not been coated again, indicating a destructive final stage of transport. Coarse sparite cores have grown either by leaching of an original nucleus and by subsequent carbonate infilling (Friedman, 1964) of the moldic pores, or by recrystallisation of a primary carbonate nucleus (Plate 5.8). Euhedral quartz crystals and aggregates of euhedral crystals, which occur scattered in both ooids and cement, are authigenic. Dolomite rhombs are abundant.

Thin sections show that the walls of vertical dewatering or degaussing pipes occurring in the Bildah carbonate are about 7 mm thick and indistinctly bounded on both sides. Some pipes are surrounded by a 0.1- to 0.3-mm-thick zone of dolomite rhombs and/or seams of chert and quartz. This zone is intergrown with the wall rocks. The filling of the pipes consists of very large carbonate crystals (up to 10 mm) and subordinate euhedral quartz crystals (up to 5 mm) (Hegenberger, 1987, Fig. 3). Three phases of diagenesis can be determined: (i) formation of pipes, (ii) lining of the wall with dolomite and/or chert, (iii) filling of the central conduit with calcite and/or megaquartz.

5.2.2.5 La Fraque Member

Whereas in all massive carbonate successions of this member at least small amounts of silt-sized siliciclastic grains were found, pure carbonate forms thin laminae within calcareous siltstone. The pure carbonate is micritic to microsparry, whereas the silty carbonate is sparry. This difference can be observed where the silty material is only a minor component. Dolomite rhombs are rare in thin section, corresponding to the results obtained from X-ray diffraction (Table 5.3).

A thin section of the sedimentary breccia from Doreen 227 reveals that the floatstone (classification see Wilson 1975, Fig. 1-6) consists mainly of imbricated flat pebbles, embedded in a matrix of calcareous, very fine- to medium-grained sandstone. The latter is composed of 25% sparite and 75% corroded quartz grains, less than 1 % plagioclase and microcline and few detrital mica flakes. The pebbles consist either of homogeneous sparite or normally packed oosparite with multiply coated grains, 0.3 to 0.8 mm across, made entirely of carbonate. Recrystallisation tends to obliterate the ooids.

Carbonate boulders which form part of the sedimentary breccia contain large columnar crystals, up to 7 cm high, rising from bedding planes. They consist of elongated sparite crystals in a microsparry surrounding and form palisade- and fan-shaped aggregates, oriented perpendicular or oblique to layering. They penetrate higher layers and locally can comprise 80% of the rock, leaving sedimentary lenses in between. Disseminated clay causes the reddish colour of the rock; it is pink in the palisade structures and deep red-brown in the surrounding sediment (Plate 5.11).

5.2.2.6 Okambara Member

Carbonate rocks of this member consist mainly of alternating quartz-rich and quartz-poor laminae, with the individual layers generally less than 1 mm thick. The size of siliciclastic grains is usually in the range of silt and fine sand; medium to coarse grains occur locally; detrital mica flakes are rare. Dolomite is subordinate in the lower unit but forms the predominant carbonate mineral of the middle and upper units (Table 5.3). Variations of the percentage of euhedral dolomite rhombs are evident in thin sections of the different units; the crystals are zoned, contain up to 5 rims, possibly indicating repeated stages of dolomitisation (Chilingar *et al.*, 1979); the long axes of the crystals are up to 0.3 mm.

The tripartition of the member in the central and southern Witvlei Synclinorium, comprising lower and upper carbonate-dominated units, separated by an arenaceous unit with different sedimentary structures, is also obvious on micro-scale. The *lower unit* is characterised by thin lamination and several lithologies can be distinguished in thin section:

- laminae 0.1 to 1.0 mm thick and rich in quartz grains, cemented by carbonate and separated from each other by 0.4 to 0.25 mm of micritic to sparry carbonate with or without scattered quartz grains;
- clay pellets (about 0.06 mm across or flattened parallel to bedding) concentrated in red-brown coloured layers, while layers devoid of or poor in them are a lighter colour;
- bedding-parallel zones, up to 0.75 mm thick, consisting of fibrous carbonate with crystal axes oriented perpendicular to layering.

These three types may form discrete horizons or occur combined, e.g. clay pellets were found enriched in fibrous carbonate seams together with quartz grains; quartz-rich layers, only one grain-size thick, alternating with pure carbonate laminae, are interpreted as having been caused by algal mats, while thicker quartz layers were most probably accumulated by traction currents and the clayey laminae originated during slack-water periods; layers one centimetre and thicker are considered to be storm generated.

Another carbonate-dominated rock type in the lower unit of the Okambara Member is an edgewise conglomerate forming successions 20 to 50 cm thick, alternating with laminites of the type described above. The edgewise conglomerate consists of up to 10-cm-long and 0.2- to 2.0-mm-thick red-brown flakes which appear as laths in thin sections or two-dimensional outcrop (Plate 4.11) and are made of sparite with abundant clay pellets. The flakes do not intersect each other but stand at different angles to the bedding planes, forming fan-shaped to roughly imbricated groups. The laths are set in a lighter-coloured, sparry and in places slightly sandy carbonate matrix. Locally the rock is strongly dolomitised. Most of the flakes are straight; a few are slightly bent. One edge is well defined whereas the opposite one is indistinctly bounded.

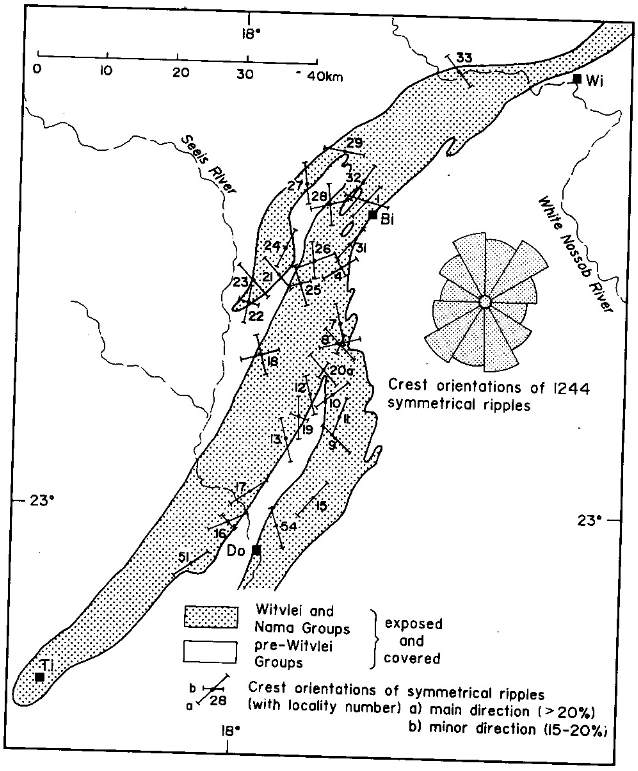


Figure 5.23: Orientation of symmetrical ripple marks in the Weissberg Member of the central and southern Witvlei Synclinorium. For localities refer to corresponding numbers in Table 6.1. Bi-Bildah, Do-Doornpoort, Ti-Tigerpforte, Wi-Witvlei

The edgewise conglomerate consists of thin clayey crusts which were peeled off a dried surface during desiccation and transported for a short distance by low-energy currents and collected in discrete layers. The sharply bounded side indicates the former surface.

In the predominently sandy *middle unit* the carbonate component is mainly contained in flat-pebble conglomerates with a calcareous sandstone matrix.

The texture of the usually crinkled algal mats in the *upper unit* is similar to but less regular than that of the straight laminated mats of the lower unit. The rock is formed by irregular wavy to kinky layers, which consist of micritic to sparry carbonate, containing little clay, scattered clay pellets, carbonate pellets, coated carbonate grains, ooids and intraclasts. Intercalated are zones of silt-sized quartz grains and layers of very coarse, drusy to fibrous carbonate and chert (Plate 5.10).

Chert is much more abundant in the algal mats of the upper unit than in those of the lower one (Plate 4.8); haematite forms seams and thin layers. Stromatolitic structures consist of dome-shaped layers which are either rich in quartz grains and up to 0.25 mm thick, or made of coarsely recrystallised carbonate, chert, or a combination of the three components (Plate 5.10). Stromatolitic layering may be accentuated by open-space structures (Wolf, 1965, 1978) with an internal filling of fibrous sparite (0.2 to 0.5 mm thick) along the surface of the walls, followed by acicular to drusy chert bands, 0.1 to 3 mm thick, (possibly stromatactics, Plate 4.8), and in places by a dark-coloured seam of clay or hematite (Plate 5.10). The voids are considered to be sheet cracks caused by dewatering and desiccation of the sediment and/or by the pressure of gas that originated during decay of algal mats. The thickness of layers of the first magnitude varies from 0.2 to 4.0 mm, with less evident layers of a second magnitude in between, caused by differences in the crystal size and the percentage of mud. Vertical to inclined desiccation cracks intersect the sediment. They were filled with the same material as is found in the sheet cracks, viz. micrite to sparite, rich in silt-sized quartz grains, indicating syn-sedimentary to early diagenetic origin; fibrous to granular sparite and chert are of a later diagenetic stage. Rocks rich in sheet cracks and desiccation fissures developed reticular textures.

Flat-pebble conglomerates, locally more than one metre thick and either clast- or matrix-supported, are a common feature of the middle and upper units; their textures resemble those of the lower portion of the member (Plate 4.9). Generally rounded to subrounded, micritic, subordinate sparry clasts are set in a matrix of sandy sparite to calcareous sandstone. The platy pebbles are either oriented parallel to bedding or imbricated, or they are unoriented. Large quartz grains of the matrix which pierce clasts indicate that the pebbles were unlithified during reworking of the sediment. The mineralogical differences between matrix and clasts are accentuated by the lighter colour of the latter and point to a high-energy environment in which carbonate layers, deposited during lower energy conditions, were ripped off and reaccumulated.

An ooid-dominated microfacies appears associated with the stromatolite-bearing zone near the top of the unit (Plate 5.10). Superficially to multiply coated grains, 0.15 to 1.5 mm across, consist of pure sparite. Their outlines and internal structures are in most cases ill-defined due to recrystallisation; spheres lacking recognisable coats might be pellets, or their cortices were diagenetically obliterated. But primary differences between spherical grains can be observed under more favourable conditions since coated grains are up to 1.5 mm across, while the maximum size of pellets does not exceed 0.5 mm. Elongated micritic to sparry pisolites, up to 6.0 x 2.5 mm, are composed of oolite clasts, made of small spherules (ooids and pellets), which are enveloped by common outer coats. In several cases, ooids both within the pisolites or isolated, are broken, indicating numerous stages of reworking, transport and resedimentation. Some adjoining ooids and pisolites have slightly deformed each other. Cement and matrix of the coated grains consist either of pure granular sparite or recrystallised sandy carbonate with silty to coarse siliciclastic grain-size fractions.

Also associated with the stromatolitic zone of the upper unit are siliceous nodules, up to 1 cm across, scattered or concentrated in layers. In thin section they reveal a spherulitic structure, caused by the radial arrangement of fibrous and megaquartz aggregates (Plate 5.11). The fibres are length-slow, i.e. the c-axis (slow ray) is parallel to the fibres. This is a strong indication that the rocks are silicified evaporites (Folk and Pittman, 1971; Siedlecka, 1972). The spherulites occur in micrite to sparite, but are concentrated in very fine-grained terrigenous sediment, indicating that the evaporite-bearing solution preferred more permeable layers.

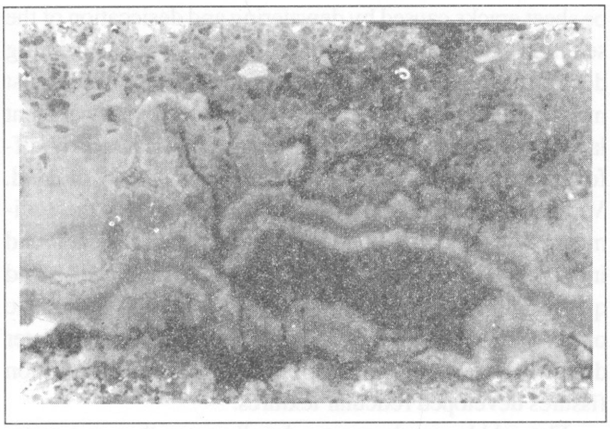


Plate 5.7: Sandy dolomite with superficially coated grains and small pebbles from the Bildah Member. These characterise near-shore deposition. Banded, fibrous carbonate crusts are possibly of vadose origin, having formed in palaeokarst cavities (spelaeothems). Okombuka 218. Sample 462A, thick section, vertical section of photograph covers 1.1 cm

Plate 5.8: Ooid grainstone from the Bildah Member. The rock consists of multiply coated ooids showing several phases of reworking and deformation during transport. Ooids are overgrown by fibrous carbonate. The drusy matrix contains secondary, euhedral quartz. Orochevley 216. Sample 190, X nic., bar is 1 mm long

Plate 5.9: Aggregates of elongated calcite crystals in the La Fraque Member forming fan- and palisade-like structures in a darker matrix. Crystals broaden upwards. Pseudomorphs after aragonite or gypsum. Southern portion of Doreen 227. Sample 420, // nic., bar is 1 mm long

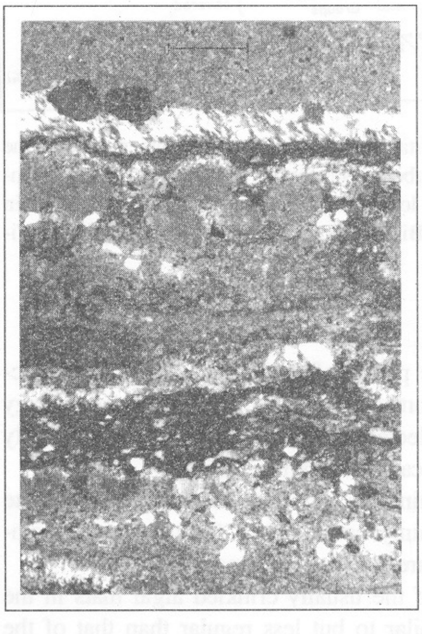


Plate 5.10: Stromatolitic structure from the upper unit of the Okambara Member. Several zones are visible (from top downwards): 1) Sheet crack, filled by fibrous chert and limonite seam, 2) coated grains and scattered quartz grains, 3) indistinctly laminated dolomite (micrite to sparite), 4) zone rich in quartz grains, 5) dark limonitic band, 6) micrite with scattered quartz grains. Okambara 219. Sample 388, // nic., bar is 0.2 mm long

The only occurrence of carbonate facies of the Okambara Member in the northern Witvlei Synclinorium consists of dolomitic, strongly recrystallised sparite; its red-brown colour is due to hematite or limonite, precipitated in interstices between carbonate crystals. Lamination is caused by differences in crystal sizes, colour banding (i.e. differing amounts of interstitial hematite) and the presence of thin, irregular seams of silt-sized quartz grains. One thin section of overpacked ooid grainstone is composed of usually deformed coated grains, 0.5 to 1.0 mm across; most cortices were obliterated by recrystallisation. Drusy, light-coloured carbonate lines the ooids; hematite (or limonite) fills the interstices of both cement and ooids (Plate 5.12).

Carbonate zones in the Okambara Member of the Gobabis Synclinorium, which in this area is predominately clastic, have a distinct microfacies. On the farms Schönborn 79F and Breitenberg 51, dusty brownish-grey micrite forms indistinctly stratified layers in the upper portion of the member. Layering is caused by a change of colour due to variations of clay content, by chert seams and locally by thin zones which contain scattered silt-sized quartz grains. Coated grains, stromatolites and flat-pebble conglomerates are absent.

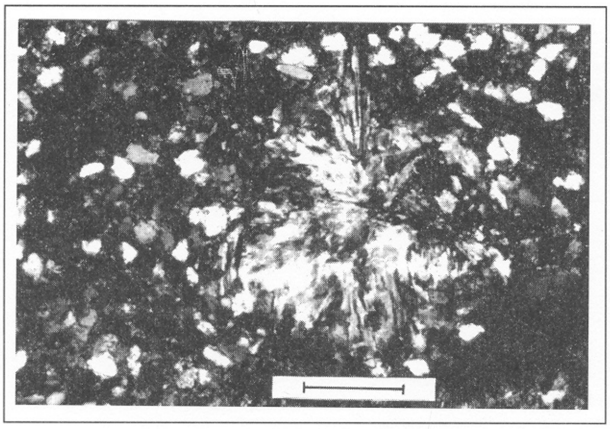


Plate 5.11: Chert spherulite in fine-sandy dolomite of the upper unit of the Okambara Member. Okambara 219. Sample 374, X nic., bar 0.5 mm long

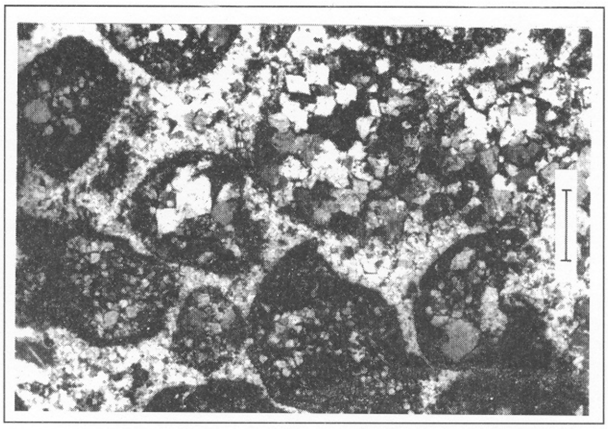


Plate 5.12: Ooid grainstone with overpacked oolite from the Okambara Member in the northern Witvlei Synclinorium. Many ooids are deformed by compaction. Ooids and matrix consist of drusy recrystallised dolomite with the largest crystals in the centres of the ooids. Dolomite crystals within the ooids are surrounded by red-brown material, probably limonite, giving a deep colour to the finer crystalline outer portions and a lighter colour of the centres; the matrix is drusy almost, colourless dolomite. Kehoro Noord 185. Sample 412, X nic., bar is 0.3 mm long

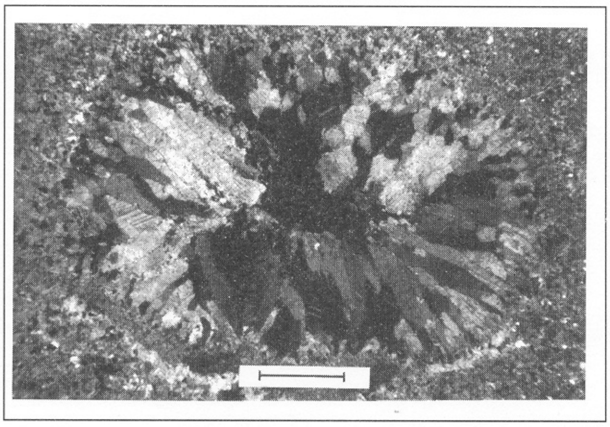


Plate 5.13: Spherulite in micritic to sparry limestone of the Grünental Member. Large bladed, centripetally oriented calcite crystals radiate from a point (or an axis, if the structure is elongated). The spherulites are interpreted as calcite pseudomorphs after gypsum. Grünental 151. Sample 279, X nic., bar is 1 mm long

5.2.2.7 Zaris Formation

The carbonate rocks from all three members of this formation display similar textures and belong to the same microfacies, viz. brownish-grey sparite, subordinate microsparite and micrite, with strongly recrystallised, indistinct zones and layers. The rock is either massive or banded, the latter caused by alternating fine (= darker) and coarser grained (= lighter) zones and patches, and by indistinct secondary hematite seams. In a few thin sections scattered silt-sized siliciclastic grains were observed along bedding planes. Quartz and rare plagioclase are strongly corroded. Calcareous siltstone of thin sections 265 and 268 displays a patchy extinction of the groundmass carbonate (26% and 35% micritic calcite and dolomite cement respectively, Tables 5.1 and 5.3). The areas of uniform extinction are indistinctly bounded, measure between 0.4 and 1.0 mm across and altogether form a mosaic pattern which is also visible between parallel nicols because of the pleochroism. The siliciclastic grains are scattered within the carbonate, This fabric is probably caused by the diagenetic to anchimetamorphic orientation of originally micritic carbonate crystals within certain portions of the cement.

At several localities in the Grünental Member, elongated, up to 5-mm-long calcite crystals form fan-shaped to radial aggregates, measuring up to 7 mm across (Plate 5.13). If the crystals do not meet at a central point but along a line, the aggregate is cylindricallly elongated. The most instructive occurrence of these structures has been found on Grünental 151, where originally they were mistaken for fossils. Another thin section from this locality consists mainly of aggregated quartz crystals: impure (clayey?) brownish carbonate spherulites (up to 0.4 mm across) are lined by fibrous chert which is oriented parallel to bedding. The length of the fibres decreases from the bedding -parallel equator of the spheres towards the poles where they are completely absent. This phenomenon can be observed even in nuclei elongated perpendicular to layering. It is explained by maximum growth parallel to minimum strain. Relics of dolomite rhombs within chert indicate that the latter formed subsequent to the former. Diagenetic chlorite is oriented parallel to the quartz fibres. Growth of the quartz fibres in an optically length-slow orientation supports the interpretation of the spherulitic structures as silicified evaporites (Folk and Pittmann, 1971; Siedlecka, 1972).

6. SEDIMENTARY STRUCTURES, PALAEOCURRENTS AND BASIN ANALYSIS

6.1 COURT FORMATION

6.1.1 Gobabis Member

The presence of three sedimentary facies which are characteristic for certain environments, and knowledge of their domains, allow reconstruction of the sedimentary basin of the Gobabis Member.

Facies (a) is a laminite (rhythmite) which is present in the central portion of the basin (Fig. 4.4). North of Gobabis it has been removed by erosion whereas further south it disappears under younger cover. This facies is up to 45 m thick and consists of alternating dark- and light-coloured laminae a fraction of a millimetre thick, forming distinct couplets.

Generally, deposition of rhythmite takes place either in a pelagic marine environment (Scholle, 1977; Scholle *et al.*, 1983; Tucker, 1983; Wilson, 1969), or in a lacustrine environment (Collinson, 1978; Dean and Fouch, 1983; Jackson, 1985; Picard and High, 1981). In both cases euxinic conditions in a sediment-starved basin create an anoxic or oxygen-depleted environment in which circulation of bottom water is restricted. The basin must be deep enough for the surface of the sediment to lie below the wave base and the zone of bioturbation. Laminite couplets originate from the rhythmic precipitation of carbonate, organic matter and clay during seasonal or annual changes. Changes of water temperature influence the amount of plankton produced and also of lime precipitated, whereas the clay content may reflect the energy of rivers entering the basin. This interpretation of laminites as seasonal sediments (varves) and the reconstruction of environmental conditions is mainly based on modern examples (Jackson, 1985; Picard and High, 1981).

Well-laminated rocks of the Gobabis Member contain an average of 27 couplets of dark-and light -coloured varves per centimetre thickness, i.e. one couplet is about 0.4 mm thick and one lamina about 0.2 mm thick. If it is assumed that varves reflect annually repeated changes in sedimentary conditions, a 45-m-thick laminite succession will have been deposited over a minimum period of about 110000 years. A longer period can in fact be assumed because laminae may be destroyed during storms or their formation may have been impeded during unfavourable conditions. This is indicated in the Gobabis laminites by dark layers more than 1 cm thick, by loop bedding and by sections where lamination is indistinct or even absent.

Absence or near-absence of detrital components and the type of lamination indicate that the facies was deposited under euxinic conditions. As bioturbation can be disregarded in Precambrian sediments, the minimum water depth is defined by the maximum depth of the wave base during fine weather, i.e. between 5 and 10 m in the case of lakes and oceans; storm waves, however, may reach down to 20 m (Tucker, 1983; Picard and High, 1981; Gary *et al.*, 1974). As portions of the laminae have been destroyed, it follows that water depth during their deposition must have exceeded 10 m. Local association of rocks characteristic of shallow water and littoral areas suggests that the water was not necessarily much deeper than this minimum depth. The absence of slump structures in the rhythmite indicates a low gradient of the sea bottom.

Facies (b) is a massive to thinly layered limestone. It overlies facies (a) and grades into it at the eastern margin of the southern Witvlei Synclinorium. This facies is considered a shallow -water sediment (less than 10 m deep). No terrigenous component is present.

South of Gobabis, the uppermost portion of facies (b), occurring immediately below the upper boundary of the member, is algal-laminated. Downward-curved laminae are interpreted as suspended algal mats which alternate laterally with structureless rock (Plate 4.1). These columnar structures are 30 to 70 cm high and about 10 cm wide. Cryptalgal lamination and algal build-ups occur north of the homestead on Josephine 226.

On Tahiti 61 and on Klippiespan (portion of Marguerite 238) a clast-supported sedimentary breccia is intercalated in the lower most portion of facies (b). On Klippiespan it forms a 0.3- to 0.5-m-thick layer about 3 m above the lower boundary of the member (Plate 4.2). These deposits are interpreted as debris flow.

Facies (c), calcareous sandstone to sandy dolomitic limestone with coated grains and flat pebbles, occurs at the southern margin of the Netso Syncline and is considered to be a shoreline facies. Its siliciclastic component is derived from the underlying Kamtsas quartzite. Basinwards, facies (c) passes within a few hundred metres into facies (b). Sandy carbonate overlying Kamtsas quartzite in the southwestern corner of Frank 221 is correlated with the shoreline facies of the Gobabis Member.

Interpretation: The Gobabis Member was deposited in a basin starved of detrital sediment. It was bordered by a hinterland which contributed very little clastic material, this being silt-sized and finer (Fig. 5.22, stage A). The initial transgression drowned a peneplain free of lag deposits. Peripheral terrigenous components have been derived by reworking of Kamtsas quartzite along the beach. Rhythmites occurring along the north-northeast-striking axis of the basin indicate seasonally controlled deposition under euxinic conditions at a depth just below wave base. As the basin filled, the water depth became shallow enough to inhibit development of lamination. The position of the eastern margin of the basin is well defined by several outcrops of shoreline sediments. On the western side of the basin the possible former position of the shore is defined by shallow-water carbonate with algal buildups and shoreline deposits (the correlation of the latter is equivocal).

6.1.2 Constance Member

The Constance depositional basin most probably coincided with that of the Gobabis Member; this can be assumed even in the western and northern parts of the Gobabis Synclinorium where erosion during early Simmenau times removed portions of the underlying strata (Figs 4.4 and 4.5). Shale and siltstone with layers of mudstone and impure dolomite are the main rock types (Table 5.3). All rock types contain a silty to fine-grained sandy silicilastic component (Table 5.2) which locally may predominate, thus forming silty to sandy intercalations. The influx of terrigenous material was distinctly higher than during deposition of the Gobabis Member.

On the farm Court 32, the lower portion of the member consists of variegated fine-grained sandy shale with mudstone layers and three dark-grey carbonate interbeds (samples 147 and 148 of Table 5.3). Further upward the amount of terrigenous components increases, and fine-grained sandstone and siltstone (Table 5.2) form several metres of parallel-bedded layers 0.2 to 0.5 m thick. The uppermost portion of the member consists of shale that contains several layers of sandstone and siltstone, 5 to 30 cm thick; the latter show low-angle cross-lamination (currents to the northwest) with wavy and lenticular bedding, scattered flute casts on the underside of the sandy layers and locally abundant pyrite cubes. Clay galls up to 1 cm across and coarse to pebbly quartz grains are locally scattered in the mudstone indicating deposition in very shallow water. On Tahiti 61, desiccation cracks occur in a 3-m-thick succession of shale, siltstone and mudstone.

Oolite in the uppermost portion of the Constance Member on the northern portion of Tahiti 61 (plate 5.6) alternates with quartzite and is overlain by typical Simmenau quartzite. Along the eastern margin of the southern Witvlei Synclinorium, the top of the member is formed by a several-metres-thick carbonate zone.

Interpretation: Composition of the sediments and the paucity of sedimentary structures indicate deposition in a generally low-energy environment in shallow water where carbonate precipitation was restricted and where occasional emergence occurred. Taking into consideration that the underlying Gobabis Member shows a distinct shallowing-upward tendency, continuation of this development is con-firmed by the sedimentary features of the Constance Member which can be interpreted as having accumulated during a lagoonal or flood plain stage of a shallowing basin (Fig. 5.22B).

6.1.3 Simmenau Member

The mainly psammitic deposits with subordinate psephitic components extend further to the east than the underlying members of the Court Formation (Fig. 4.6). The light-coloured subarkose is fine- to medium-grained and moderately sorted (Tables 5.1 and 5.2; Plate 5.1). Characteristic of the generally thick to massively bedded quartzite is a paucity of sedimentary structures. The thickest section through the member is exposed in the gorge of the White Nossob River and reveals several zones of different bedding thickness with a fining-upward tendency (Table 5.2, samples 418,167,168; Plate 4.4).

Clay pellets oriented parallel to layering or imbricated are locally abundant. In the gorge of the White Nossob River they occur in the lower most portion of the member. In most cases clay galls are interpreted as having formed when fragments from thin desiccated clay layers were transported for a short distance and buried by sand; however, subaqueous erosion of coherent mud beds can produce the same result (Potter *et al.*, 1980). Spherical cavities, as shown on Plate 4.3, are believed to have been produced by diagenetic processes.

Parting lineations in the Gobabis Synclinorium have a southwest-northeast orientation, whereas current ripples in area 2219 DA with an average wavelength of 15 cm, a height = amplitude of 1 cm and a ripple index (Tanner, 1967) of 15 indicate a transport direction to the northeast (Fig. 4.6; for exact localities see Table 6.1). Mineralogical and grain-size analyses showed that the main source of the siliciclastic sediments of the Simmenau Member was situated close to the depositional area and most probably consisted of Kamtsas quartzite and subordinate amounts of Doornpoort rocks. Following upwarping (Miller, 1983, p. 505) and erosion in the Southern Margin Zone of the Damara Orogen prior to the deposition of the Buschmannsklippe Formation, it can be assumed that much of the clastic material of the Simmenau Member was derived from the southwest, i.e. the Dordabis-Rehoboth area. Derivation of sediments from this region was also inferred by Germs (1972, 1983) for the slightly younger Dabis Formation that was deposited further to the south.

The polymictic conglomerate occurring at or near the base of the Simmenau Member along the western margin of the Gobabis Synclinorium contains pebbles of i.a. lower Simmenau quartzite which indicates an intra-Simmenau uplift and erosion in this region. Along the eastern margin of the Witvlei Synclinorium, a conglomerate occurs on the farms Frank 221 and Josephine 226. Absence of pebbles derived from the laminated facies (a) of the Gobabis carbonate excludes an eastern or southeastern source for this conglomerate. The unconformity marked by this conglomerate is greatest in the southwestern part of the Gobabis Synclinorium, where the conglomerate overlies Kamtsas quartzite directly; here the Gobabis and Constance Members of the Witvlei Group, the Blaubeker Formation and probably the top of the Kamtsas Formation of the Nosib Group have been removed by erosion. Carbonate clasts in the conglomerate are similar to the carbonate in the Constance Member and the unlaminated Gobabis Member. The erosional gap is smallest at Gobabis, where the conglomerate rests on the unlaminated facies of the upper Gobabis Member. The absence of the Constance Member in this region might be primary or due to erosion. The distribution and composition of the conglomerate and the amount of erosion indicate a provenance region to the southwest.

Interpretation: Due to the scarcity of data, a sedimentological interpretation of the Simmenau Member remains tentative. The succession accumulated rapidly in shallow water. The shallowing -upward trend in the lower portion of the Court Formation is followed by flood plain deposits of the Constance Member. In an ideal cycle (Picard and High, 1981, p. 247), coarse clastics of fluviatile origin would be expected higher up in the succession. In the Sirnmenau Member the coarsest grain sizes occur in the lower part of the succession and the sequence fines upward. The extensive, massively bedded Simmenau quartz-

1	Okambara 219, 0.5 km SE of homestead
2	Okambara 219, near homestead
4	Armhem 222, eastern portion
6	Josephine 226
7	Josephine 226
8	Otjimbondona 225, eastern boundary, near main road
9	Boundary Marguerite 238/Renette 232, in gorge
10	Helen 231, northern boundary
11	Helen 231, middle of eastern boundary
12	Helen 231, NW corner
13	Renette 232, western boundary, near homestead
15	Boundary Julia 239/Alice 237
16	Kowas 233, SE corner of northern portion
17	Kowas 233, northern portion, near road to Renette 232
18	Aoda 296, eastern boundary
19	Boundary Lacockshoop 297/Helen 231, at road
20a	Otjimbondona 225, at deserted homestead
21	Scheidthof 293, gorge near homestead
22	Scheidthof 293, southwestern gorge
23	Scheidthof 293, northwestern hill
24	Scheidthof 293, near northern beacon
25	Arnhem 222, at guano cave
26	La Fraque (southern portion of Okambara 219)
27	Boundary Okombuka 218/Okambara 219
28	Boundary Orochevley 216/Okambara 219
29	Weissberg (corner Owiniekiro 213/Sulinam 215)
31	Bildah 220, SW boundary
32	Corner Sulinam 215/Bildah 220/Okambara 219
33	Held 84, near White Nossob River
34	Valerie 291, hill near western boundary
35	Soetblomspan 596, at large pan
36	Tygerpoort 285, centre of syncline
40	Karossewie 72, SW portion
41	Kaukurus O 79, at homestead
42	Kirchberg 79, northern portion, at Black Nossob River
43	Spatzenfeld 70, NW portion
44	Blumenau 58, near homestead
46	Kuduberg 60
47	Kanabis 55, NE portion, near beacon
48	Kanabis 55, NE portion, near beacon
49	Boundary Eilenriede 53/Kanabis 54, along river
50	Nautabis 268, NE corner and central portion of farm
51	Nautabis 268, NE corner
54	Keerweeder (Gungams), ptn of Achenib 247, near homestead
56	Okambara 219, near homestead
57	Bildah 220, SW portion
59	Gobabis, south of town, east of Black Nossob River
60	Leeuwpoort 598, east of homestead
61	Leeuwpoort 598, at homestead
63	Groenrivier, ptn of Klein Keitsaub 59, at Black Nossob River
64	Kuduberg 60, east of homestead
65	Eilenriede 53, near White Nossob River, S side of main road
67	Scheidthof 293, gorge near homestead
68	Orochevley 216, western side
69	Otjimbondona 225, at deserted homestead
70	Boundary Otjimbondona 225/Doreen 227
71	Boundary Marguerite 238/Renette 232, in gorge
72	Helen 231, W of homestead
73	Kowas 233, northern portion, 3.5 km NE of homestead
74	Corner Smalhoek 236/Kowas 233/Nautabis 268
75	Nautabis 268, north of main road
76	Near northern corner Julia 239/Alice 237
77	Springboklaagte 209, 0.5 km W of cattle post S of main road

Table 6.1: Localities where sedimentary structures were measured (crossbedding, ripple marks, parting lineations). Numbers refer to Figures 4.6, 4.9, 4.10, 4.11 and 5.23

ite which is devoid of pelitic intercalations, suggests a braided stream deposit. The main direction of -transport was from the southwest. The conglomerate is interpreted as an alluvial fan, transported from an elevated area in the southwest (Figs 4.6 and 5.22 C). Sandy components of the matrix were derived from reworked quartzites and from as-yet-unconsolidated Simmenau sands, while dark carbonate matrix originated mainly from comminuted and partly dissolved Gobabis rocks. Stromatolites occurring in this horizon are considered to have grown in a lacustrine environment, the lakes having been dammed up by fan accretion. Similar occurrences were described by Buck (1980) from the Ventersdorp Supergroup of South Africa and by Molenaar and De Feyter (1985) from the Cretaceous of Italy.

6.1.4. Résumé of the Court Formation

In considering the development of the Court basin and its sediments, starting with rhythmites which are characteristic for "deep-water" conditions in both lakes and oceans, and shallowing upward to a terminal fluvial stage, two questions arise:

(1) was the environment of the lower Court Formation lacustrine or marine, and

(2) do the three members of the Court Formation constitute a continuous cycle or was deposition interrupted by discontinuities?

(1) "Deep-water" laminites (facies (a) of the Gobabis Member) are believed to have formed along the axis of the trough, an area extending in a north-northeast-south-southwest direction. To the west and east, this facies passes into the shallow water carbonates of facies (b). In the east a marginal facies (facies (c)), is preserved which is exposed only at one locality (Fig 4.4) in the west, but is assumed to be more widely distributed below younger rocks in the southern Witvlei Synclinorium. There is no indication that the Court Formation, as preserved today, might represent only the remnants of a once more extensive occurrence much of which was removed prior to deposition of the Buschmannsklippe Formation. Along the northwestern and western margin of the Aranos Syncline, the Court Formation is absent. The southern boundary of this formation might have been formed by the Osis Ridge. This topographic high was most prominent at the start of Nama deposition, but its influence on sedimentation decreased with time (Germs, 1972a, 1983). The existence of the Osis Ridge during the pre-Nama period can thus be assumed. The Gobabis Member extended to somewhere north of Gobabis (Fig 4.4) and was deposited in a north-northwesterly trending basin some 250 km long, 120 to 150 km wide and about 30 000 km^2 in area. This is similar to the average size of ancient lakes mentioned by Bradley (1963; quoted by Picard and High, 1981). This basin meets two criteria used to define a lake, i.e. it is not part of an ocean and it has a limited dimension (Horie, 1978). Another characteristic feature, viz. that it has only a limited duration compared with marine basins (Horie, 1978; Table 4.1), becomes apparent when counting of the varves of facies (a) resulted in a minimum duration of 110 000 years for that unit. Sedimentary features like flat-pebble conglomerates, oolites and stromatolites are mainly described from shallow marine environments but are also common in lacustrine environments (Collinson, 1978; Buck, 1980; Picard and High, 1981; Dean and Fouch, 1983). Lenticular bedding, characteristic of intra- and subtidal zones, has also been reported in lacustrine sediments (Reineck and Wunderlich, 1968). All this supports or at least does not contradict an interpretation of the lower Court Formation as being of lacustrine origin. However, as the margins of the Court basin are ill defined, it might have been connected in some way (e.g. by an inlet) with the Damara trough further west.

(2) The fine-grained flood plain sediments of the Constance Member, believed to form the terminal lacustrine deposits, are overlain by much coarser arenites of the Simmenau Member that were probably deposited in a braided fluviatile environment. However, transition or interfingering of both sediment types was found only locally. The ideal cycle in which the amount and grain size of clastic sediments increases upwards, is disturbed by the Simmenau Member which represents a subcycle with a locally reversed, fining-upward tendency. The same tectonic event that caused uplift and erosion of the source area in the west and southwest of the Witvlei Synclinorium during Simmenau times, probably affected portions of the Court basin, resulting in a southwesterly plunging uplift (anticline or horst) of the area between the Witvlei and Gobabis Synclinoria, heralding formation of the Nina Anticline (Fig 1.2). Erosion of this high, down to the level of the Nosib rocks in the southwest, furnished material for alluvial fan conglomerates which in their turn were buried by the quartzite of the upper Simmenau Member (Fig. 5.22C). With decreasing relief and erosional energy the sediment gradually became finer grained higher up.

The Court Formation is a sequence in the sense of the sequence stratigraphy (Van Wagoner *et al.*, 1988) bounded by a type 1 sequence boundary at the base and a type 1 or 2 boundary at the top. The basal Tahiti Member can be interpreted as a valley fill at a low stand basin floor. Correlation of the Witvlei Subgroup with units of the southern Damara Belt and the Naukluft Nappe Complex was proposed by Hoffmann (1989). The Gobabis and Constance Members correlate well with the Kudis Subgroup, considering that the latter was deposited in deeper parts of the Damara Basin. As shown above, the arenaceous sediments of the Simmenau Member were derived from uplifted terranes southwest of the study area, from where most of the eroded material was probably transported into the Damara trough, forming the turbiditic quartzites of the Auas and possibly also Kleine Kuppe Formations.

6.2 BUSCHMANNSKLIPPE FORMATION

6.2.1 Bildah Member

The faint, omnipresent lamination of the light-grey, thickly to massively bedded dolomite is cryptalglal in the sense of Aitken (1967). The laminae are either parallel to the palaeohorizon or form irregular humps and domes up to about one metre high and several square metres in area (Plate 4.5) which are interpreted as algal build-ups. Locally, LLH-C stromatolites (nomenclature after Logan *et al.*, 1964) have developed. More details are visible on slabs quarried at Kehoro 183 (Plate 6.1) which display flat and undulatory to wavy laminations with pronounced micro-unconformities (Collinson and Thompson, 1982, Fig. 8.6). Desiccation cracks are locally abundant; sheet cracks are presumably caused by gas generated in decaying algal mats during emergence or by shrinkage of sediment due to desiccation (Wilson, 1975). Desiccation leads to disruption of the laminae, resulting in unconformable contacts of the layers and generation of incipient flat-pebble and edgewise conglomerates (intraclast grainstone). The absence of well-developed intraclast conglomerates points to a low-energy environment. In the western Witvlei Synclinorium, however, crossbedding and small channels indicate internal reworking by currents.

Locally, discrete well-laminated columnar and pseudo-columnar stromatolites (LLH-S) are developed. The structures are more than one metre high and measure up to five centimetres across; their tops, however, stood not more than 5 cm above the surface of the surrounding sediment (Plate 6.2). Low elevation above the sediment indicates relatively calm water conditions (Wright, 1981). While the occurrence of stromatolites is not necessarily indicative of water depth (Hoffman, 1974), the other sedimentary structures mentioned above are characteristic of intermittent exposure in the peritidal zone (Shinn, 1968, 1983a and b; Ginsburg, 1975; Wilson, 1975). Slight reworking by waves and currents and the nature of columnar stromatolites generally indicate sedimentation in a protected, low-energy environment.

Birdseye structures (fenestrae) (Shinn, 1968, 1983a and b) are scarce; this might be explained by obliteration of the vugs during compaction of still-unlithified sediments, a process which does not affect other structures like cryptalgal lamination, mudcracks and tabular cracks as intensely (Shinn, 1983a and b; Shinn and Robbin, 1983).

Approaching the shoreline at the southern margin of the Witvlei Ridge and on the farm Sachsenwald 940, carbonate precipitation was accompanied by deposition of terrigenous sand with coated grains and clasts. Even nearer to the shore, calcareous sandstone and eventually gritty to pebbly sandstone with low-angle crossbedding were deposited.

On Held 84 (2218 AD), sandy carbonate layers are intercalated in the pure dolomite of the Bildah Member. Dome-like structures originate from the pure dolomite layers and disturb (penetrate) the bedding of the overlying strata. These structures resemble chicken wire and diapiric textures of anhydrite in sabkhas (Shinn, 1983b, Figs 38 and 39). In the Bildah Member, however, thin seams of sandy carbonate and/or silica traverse the structures parallel to bedding and indicate updoming of the layers. Therefore, the domal structures are explained as stromatolites which grew during deposition of the pure carbonate layers and were slightly or completely suppressed during siliciclastic sedimentation. They measure between 2 and 20 cm high, but can be as much as one metre high. In the latter case the otherwise vertically aligned structures are bent sideways (Plate 6.3).

Oolite and pisolite, together with clasts of Bildah and Eskadron rocks, fill a channel in the lower, algal-laminated Bildah Member on Orochevley 216 (2218 CA) (Plates 4.6 and 5.8). The ooids were transported by strong currents, most probably induced by storms, which cut into the Bildah and locally also into the Eskadron rocks. The multicoated ooids attest to an environment more agitated than that in which they were finally deposited. They are believed to have formed on the seaward side of a shoal situated west of the Witvlei Ridge. Existence of a shoal is also indicated by the presence, in the same area, of karst features dating from a period of synsedimentary emergence of the sea floor, and concomitant erosion.

Tubes or pipe-like structures, arranged parallel to each other, stand vertical to the palaeohorizon and cut approximately perpendicular through the lamination or at a high angle to it where the laminae are not even, e.g. in algal build-ups. They are plentiful in many areas of the central Witvlei Synclinorium (2218 CA) but were found scattered in the rest of the study area, e.g. in the quarry on Kehoro 183 (2218 BA) (Plate 6.4). In the uppermost, transitional portion of the Bildah Member the tubes are absent. The pipes measure between 1 and 2.5 cm across and are spaced 1 to 5 cm apart. On Kehoro 183 their vertical length exceeds 30 cm. They are usually filled with coarse crystalline calcite, or, less commonly, with chert and megaquartz. On Kehoro 183, structureless micrite and tabular carbonate fragments fill the interior of several tubes (Plate 6.4). Their cross sections, studding many bedding planes, form round to oval depressions or crater-like structures. The structures were formed as pipe-like cavities and filled either from above by lime mud and/or from below by fragments of the reworked and expelled laminite. Some of the tubes remained open for a while and were filled diagenetically by sparite and subordinate chert.

Similar pipes are known from other late Precambrian peritidal carbonate rocks from Namibia (Hegenberger, 1987) and California (Cloud *et al.*, 1974). They are explained as gas- and probably water-escape structures. Gas was formed by the decay of organic matter of the algal mats mainly at low tide. Pressure and the ability to penetrate the overlying sediment could have been increased by compaction of the lime mud, by dewatering during low tide or by groundwater rising with high tide. For more details see Hegenberger (1987).

Some samples from the Kehoro quarry exhibit post-sedimentary structures:

a) Extensive fracturing of the laminae created cavities and channels cutting irregularly through the algal laminite. These cavities are filled by a breccia consisting of angular equidimensional or tubular fragments of the laminite. The margins of these cavities can be sharp or highly disturbed. This feature must have been caused by a sudden

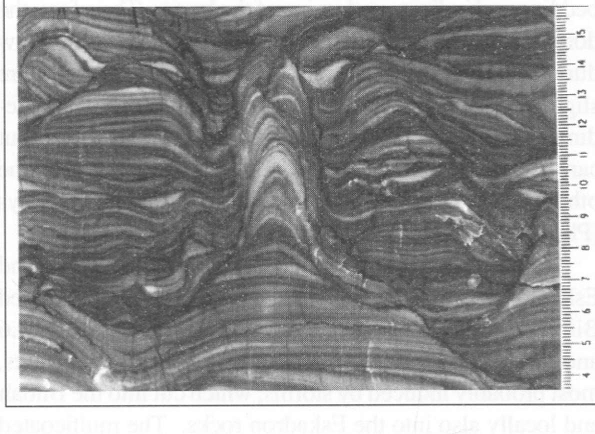


Plate 6.1: Algal laminite of the Bildah Member with discrete columnar stromatolite. Stylolites follow micro-unconformities and lamination. Kehoro 183. Scale in centimetres

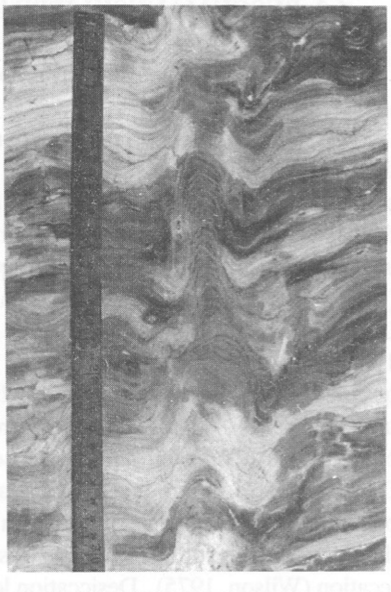


Plate 6.2: Algal laminite of the Bildah Member. Pseudocolumnar stromatolite, desiccation cracks and distorted fragments of algal mats. Scale in centimetres

Plate 6.3: Large columnar stromatolite (?) (light-coloured carbonate) in sandy (dark) and pure (light-grey) dolomite. Structure is inclined to bedding. Bildah Member, southern margin of Witvlei Ridge. Southwestern portion of Held 84. Scale in centimetres

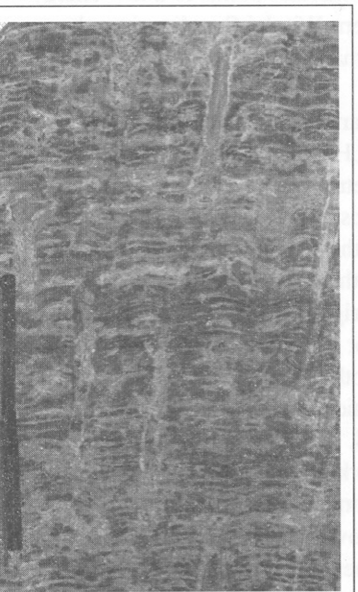


Plate 6.4 (left): Algal laminite in the Bildah Member with longitudinal section through tubular pipes. Tubes are aligned parallel to each other and are cut by the polished surface at a high angle, and are therefore not exposed over their entire length. The upward and downward continuation of many pipes is indicated by discolouration and/or downwarping of the laminae. Kehoro 183. Scale in centimetres

Plate 6.5: Internal reworking of algal laminite of the Bildah Member caused by slumping of the coherent but still unlithified sediment. Kehoro 183. Scale in centimetres

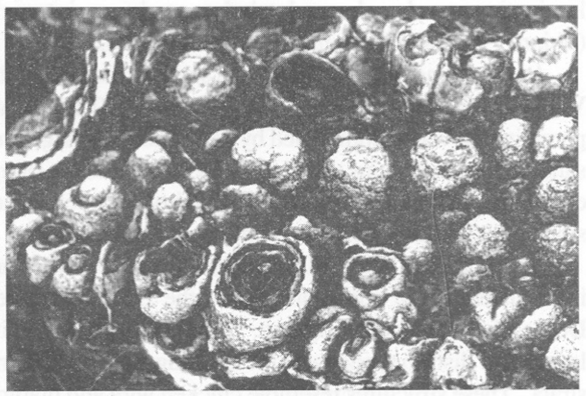


Plate 6.6: Stromatolites forming a characteristic layer in the uppermost portion of the Okambara Member. Okambara 219. Section about 30 cm long (see also front cover). *Photograph*: K.H. Hoffmann

event, e.g. movement of water or gas within the coherent but as-yet-unlithified sediment. P. Cloud (pers. comm., 1986) mentioned as possible causes solution and collapse or concussion due to seismicity, resulting in vigorous expulsion of ground water.

b) Plate 6.5 shows the effects of internal slumping with slumped and disturbed laminated mud overlying a truncation surface. These structures should be common in sediments cut by tidal channels whose stability changes frequently under peritidal conditions.

The upper 5 to 10 m of the Bildah Member are well and thinly stratified; algal lamination is scarce or absent, dolomite content decreases, whereas the supply of fine-grained terrigenous clastics increased slightly; this might have inhibited algal growth and carbonate precipitation. As this zone forms a transition to the La Fraque Member, a deeper water environment than during the preceding Bildah times may have existed.

Interpretation: The depositional basin of the Bildah Member did not coincide with that of the previously deposited Court Formation (Figs 4.2, 4.6 and 4.7). Rather the onset of Bildah deposition marked the initiation of a new sedimentary cycle in a greatly enlarged basin. The earliest carbonate precipitation may locally have percolated into the uppermost layers of Simmenau quartzite, simulating a transition between both members.

Accumulation of the Bildah succession took place after transgression of the sea onto a peneplaned surface. The relation to the underlying Court Formation is therefore paraconformable and a basal conglomerate is lacking. The same relationship exists between the Bildah Member and the Kamtsas Formation. In the Witvlei Synclinorium the Bildah Member transgresses across faults from Kamtsas upon Doornpoort and Eskadron rocks and this shows that tectonic displacement had affected the area between Kamtsas and Bildah time which had also locally resulted in weak folding of the Kamtsas rocks (Plate 3.1).

Three facies belts can be distinguished, viz:

A narrow coastal belt, influenced by clastic sediments from the borderlands and by wave action.

An extensive low-energy tidal flat area (standard facies belt 8 of Wilson, 1975) that was subjected to frequent subaerial exposures and in which most of the Bildah succession, comprising algal-laminated carbonate rocks, was accumulated.

An offshore shoal at the outer edge of the platform (tidal shelf) where oolite formed and strong wave action and even periodic emergence occurred. This coincides with standard facies belt 6 of Wilson (1975).

Uniform shallow-water facies throughout most of Bildah times, resulting in a maximum thickness of 60 m, requires that the rate of carbonate precipitation kept up with the rate of basin subsidence. From this it can be concluded that the transgression started in areas of maximum thickness of the member, i.e. in the west and north of the study area, from where it advanced southeastwards, leaving, however, the Witvlei Ridge an island.

It was only towards the end of Bildah times that deposition no longer kept pace with basin subsidence. Reduced carbonate precipitation and increasing depth of water, not balanced by siliciclastic input, caused the uppermost Bildah sediments to form a transition to the overlying La Fraque Member.

6.2.2 La Fraque Member

Shale, siltstone and fine-grained sandstone, the principal constituents of this member, usually display faint planar crossbedding, as well as wavy and lenticular bedding. These features are characteristic of deposition in tidal areas where they are generated by weak currents and gentle water movement (Reineck and Wunderlich, 1968; Reineck, 1975). Greater water depth than during deposition of the Bildah Member is inferred from the general absence of evidence for subaerial exposure.

Lamination in rocks ranging from silty carbonate to calcareous fine-grained sandstone is caused by changes in transport energy and by algal accretion. Clay galls within the upper portion of the member indicate local subaerial or subaqueous erosion.

On the farm Doreen 227 a thick zone of conglomerate and sedimentary breccia consists of floatstone with carbonate clasts, several of which are oolitic, set in a sandy carbonate matrix. From the oolite it can be inferred that the rocks, which now form the conglomerate, were deposited on a shoal, situated toward the west. From there the material was transported to its present site by either gravity flows or storm action. Boulders of pinkish calcitic micrite within the sedimentary breccia contain fans of elongated megacrystalline calcite. Similar occurrence of crystal fans are described by Grotzinger (1989) from shallow open marine environments of Precambrian platforms and are interpreted as a biotic precipitations of calcite or aragonite from oversaturated sea water.

Interpretation: The fine-grained to muddy sediments accumulated in a generally low-energy regime and the virtual lack of emergence suggests deposition in a shallow subtidal environment, probably in a shelf lagoon. This environment corresponds to standard facies belt 7 (Wilson, 1975). Absence of stromatolites is explained by reduced penetration of light in water enriched in terrigenous mud, and by sedimentation rates of the terrigenous constituents high enough to suppress algal growth (Cloud, 1942; Rezark, 1957; Walker *et al.*, 1983).

The lagoon was possibly separated from the open shelf by a barrier to the west and northwest which even in Bildah times existed as a high-energy zone. Rock fragments and boulders from this zone were transported into the lagoon during storms.

The dolomite content decreases from the upper Bildah Member to the La Fraque Member. This can either be caused by increasing water depth, or it may be explained chemically by a proportional relationship between hematite (which is ubiquitous in the La Fraque Member) and the degree of dedolomitisation (Chilingar *et al.*, 1979).

6.2.3 Okambara Member

6.2.3.1 Carbonate-dominated facies

Terminology: Unlithified carbonate and clay layers (crusts) break up when desiccated during subaerial exposure and the fragments may then be transported (Fagerstrom, 1967), slightly rounded, redeposited within a shorter or longer distance, and finally cemented by a matrix similar or dissimilar to the material of the clasts. The resulting sediments are often indiscriminately referred to as "edgewise conglomerate", "flat-pebble conglomerate", "desiccation breccia", "clay-pebble conglomerate" or "clay-pebble breccia" (e.g. Fairbridge, 1978b; Fagerstrom, 1978), terms synonymous with "intraformational (pebble) conglomerate" (Wilson, 1975).

Edgewise arrangement of clasts is caused by imbrication. In this memoir, "edgewise conglomerate" is applied to very thin fragments of calcareous clay crusts which are disposed edgewise to their substratum, many standing vertically on the bedding plane (Plate 4.11), while "flat-pebble conglomerate", as understood, contains thicker, less uniform, tabular to subrounded clasts arranged either parallel to or at an acute angle, or even randomly to stratification. "Intraformational conglomerate" is the generic term for both edgewise and flat-pebble conglomerates, and pebbles of both types are called "intraclasts" (Shinn, 1983b). "Clay-pebble conglomerate" describes accumulations of clay (mud) galls only.

Lower Unit

Several sedimentary structures characterise this unit.

(a) Cryptalgal laminations. These are caused by alternation of silty to fine-sandy carbonate layers which are between less than one millimetre and more than one centimetre thick.

(b) Edgewise conglomerate. This contains clasts 0.2 to 2.0 mm thick and up to 10 cm long which accumulated in layers and lenses up to 50 cm thick on a surface that in many cases was eroded and scoured (Plate 4.11). Commonly, transitions from well-laminated, undisturbed layers through layers in which clasts are in the process of being broken and ripped off, to entirely detached clasts are present.

(c) Flat-pebble conglomerate. This is less abundant than edgewise conglomerate and forms zones up to one metre thick. The angular to rounded pebbles of pure to slightly sandy carbonate are imbricated and set in a sandy carbonate matrix containing up to 50% siliciclastic material; this indicates that the clasts were derived from rocks formed in an environment quieter than that prevailing during deposition of the conglomerates and a storm-related origin in the shallow subtidal region is suggested by Sepkoski (1982). In places, one can observe how the edgewise conglomerate passes into flat-pebble conglomerate when both thin clay crusts and thicker carbonate layers were affected by rework-ing. Generally, however, environmental differences, like water depth and the energy of the transporting currents, dictate that only one or the other will be deposited, the currents forming edgewise conglomerate apparently being too weak to transport the clasts of the flat-pebble conglomerate, and the currents giving rise to the latter being strong enough to destroy friable mud flakes. Desiccation cracks abound in the country rocks of both types of conglomerate, suggesting that they were deposited in a periodically emerging environment and that both flat-pebble and edgewise conglomerates resulted from desiccation of layers during subaerial exposure.

Various depositional processes have been proposed for edgewise and flat-pebble conglomerates. Shinn (1983b) suggested that clasts of flat-pebble conglomerates were transported by strong currents (tidal streams or storms) and accumulated either on supratidal flats or in tidal channels. The delicate particles of the edgewise conglomerate may have been transported either by wind (Fagerstrom, 1978) or by weak water currents (Fairbridge, 1978; Shinn, 1983b). Hoffman (1975) interpreted edgewise conglomerates from the Proterozoic Rocknest Formation as having been deposited in the shallow subtidal to lower intertidal zone, whereas Betrand-Sarfati and Moussine-Pouchkine (1988) suggested storm-related sedimentation of what they called "flake conglomerate". According to Grotzinger (1986), edgewise conglomerate accumulated during storms in the subtidal zone below fairweather wave base, but not necessarily in a high-energy environment. Numerous siliciclastic intercalations with hummocky cross-stratification in the lower unit of the Okambara Member indicate that a large percentage of the sediments accumulated during storms. Therefore storms may playa dominant role in the formation of the edgewise and flat-pebble conglomerates.

As the zones of the peritidal milieu are intimately interrelated (e.g. the intertidal zone can be traversed by subtidal channels), structures which are characteristic of certain zones may occur next to each other.

(d) Hummocky cross-stratification occurs in a variety of the sandy layers which are intercalated in the carbonate, as mentioned in (c).

(e) Desiccation cracks. These are abundant in this unit and attest to frequent exposure.

(f) Scour marks resembling rill or wrinkle marks. These occur at the crest of mud-draped ripples in siltstone layers near the base of the unit. Both rill and wrinkle marks form in very shallow water during gradual emergence of a sedimentary surface (Reineck and Singh, 1973). Allen (1985) described wrinkle marks from an intertidal area and considered them to be soft-sediment deformation structures, evolving soon after emergence. Ripples also occur higher up in the succession in quartzite layers.

No coated grains and birds eyes, or indications of the former presence of evaporite minerals, were found; large algal build-ups, up to 10 m across, developed at a few localities.

The above sedimentary structures point to a storm-dominated environment in the shallow subtidal to lower intertidal zone.

Middle Unit

Quartzite is the dominant lithotype. It forms laminated beds up to one metre thick and contains layers with both planar and trough cross-stratification which are of medium scale, 6 to 20 cm thick and dip at a low angle (terminology after McKee and Weir, 1953). Vectors of the palaeocurrents follow a bi- to quadrimodal dispersal pattern, with modal classes about 180° and 90° apart (Fig. 4.9; for exact localities see Table 6.1). Hummocky cross-stratification and symmetrical ripple marks occur in places. Clay-gall conglomerates and subordinate layers of sandy carbonate containing carbonate clasts (flat-pebble conglomerate) are indicative of currents in an emergent surrounding.

The middle unit may have formed in a subtidal channelled belt containing mainly clastic sediments in a broader carbonate tidal flat environment. Channel migration and storms reworked and incorporated intertidal carbonate sediments (Galloway and Hobday, 1983; Reineck and Singh, 1973; Sellwood, 1978).

Upper Unit

Grainstone and rudstone predominates in this mainly calcareous unit. The siliciclastic content (in the form of intercalated sandy carbonate and quartzite layers) is higher than that of the lower unit but less than in the middle unit. Abundant internal reworking and discordances become apparent with the occurrence of flat -pebble conglomerates and layers of crossbedded quartzite and sandy carbonate (Plate 4.7). Hummocky cross-stratification occurs. No edgewise conglomerate was formed.

Many cryptalgal laminae are crinkled or crenulated and intersected by desiccation cracks (Plate 4.8); they include seams of chert (= stromatactics?) (Plate 4.8) and thin sandy layers parallel to lamination. They may form small domal stromatolites (Plate 4.8). Large dome-shaped algal build-ups occur in the lower portion of the unit (Plate 4.9). Neighbouring domes are separated from each other by abrupt changes in the inclination of the marginal algal layers, or are connected by undulating algal mats. In several cases unlaminated, indistinctly to horizontally layered carbonate is intercalated between the build-ups.

In the uppermost 10 to 15 m of the upper unit, a great variety of sedimentary structures occurs. Stromatolites developed in a 2- to 10-m-thick horizon about 5 to 8 m below the Weissberg Member that extends for 40 to 50 km along strike. They grew from crinkled algal mats to LLH-C domal forms which normally are between a few and 15 cm high and remain uncharacteristic (Plate 6.6). At some localities more specialised forms were found: club-shaped LLH-S structures, about 30 cm high, and discrete vertically stacked hemispheres, the overlapping laminae of which reach almost down to the base of the preceding ones (SH-C) (Logan *et al.*, 1964). In similar structures, discrete vertically stacked hemispheroids are overlain by closely laterally linked hemispheroids; their formula is SH-C LLH-C. If the upper laminae do not reach the base of the lower ones, their formula is SH-V/LLH-C. By means of overlapping laminae which are thickest on the flanks and thinnest on the top the stromatolites appear flattened. The largest specimens are nearly 50 cm wide and 30 cm high. Branching stromatolites are rare; apparent ramifications near the base are actually close laterally linked developing hemispheroids (Plate 6.7). Branching LLH-type columnar stromatolites also occur, probably with the formula LLH-S/LLH-C. Branching columnar stromatolites surrounded by flat-lying algal mats resemble "walled bioherms" described by Bertrand-Sarfati and Moussine-Pouchkine (1988) and are shown on Plate 6.8.

The stromatolite-bearing zone includes coated grains (ooids and pisolites) (Plate 5.10). Chert nodules up to one centimetre across within the stromatolite zone or immediately overlying it are either scattered or are arranged in distinct layers (Plate 5.11). The chert nodules are inter-preted as replacing evaporite minerals and some actually resemble nodular anhydrite (cf. Shinn, 1983b, Fig. 44 A).

In the uppermost few metres of the Okambara Member, above the stromatolite-bearing zone, the proportion of flat-pebble conglomerate, sandy intercalations (both sandy carbonate and quartzite) with intensive cross-stratification and channel-fill deposits increases, while the carbonate content decreases. This zone is characterised by bidirectional and herringbone cross-stratification (Plate 6.9), mud-gall con-glomerates, desiccation cracks and oscillation ripples in the sandy layers. These features are also present in the lower-most portion of the Weissberg quartzite.

The upper unit of the Okambara Member is of intertidal origin, having formed under higher energy conditions than the lower unit. This is indicated by more abundant sandy, cross bedded intercalations and flat-pebble conglomerates. Regression seems to have continued during deposition of the uppermost portion, resulting in upper intertidal and even supratidal environments.

On Kehoro N 185 cryptalgal laminites, stromatolites and the occurrence of oolite (Plate 5.12) suggest sedimentation in a tidal flat environment.

6.2.3.2 Clastic facies

Towards the Witvlei Ridge, the thickness of the Okambara Member decreases and the carbonate-dominated facies passes into a clastic facies. The boundary between both facies can roughly be drawn along the 60 m isopach (Fig. 4.9). Successions less than 60 m thick consist mainly of sandstone and quartzite, whereas in areas where the member exceeds 60 m, the additional thickness consists of carbonate. That means that the amount of clastic material is relatively constant in the Okambara basin; in the case of the carbonate facies, however, the mainly chemically accumulated constituents (varying from carbonate cement in the sandstones to intercalations and sequences of pure carbonate) account for the increase in thickness. This rule is modified by the increase in thickness of the sandy middle unit in the southern Witvlei Synclinorium.

As was the case with the La Fraque Member, the Witvlei Ridge contributed no clastic material to the Okambara deposits. Scattered symmetrical ripples indicate transport of the siliciclastic material by water. The nearest coarse-grained offshore intercalations in the fine-grained sandstone are clay-gall conglomerates. Still further basinward, a 3-m-thick sedimentary breccia consisting of sandy carbonate clasts in sandy matrix is intercalated and can be followed for several hundred metres perpendicular to the coastline. Sandstone boulders in the conglomerate indicate high energy of the reworking agents, probably during storms.

The clastic facies originated in a shallow marine beach zone. Currents and waves had lost most of their energy after crossing the vast tidal flats within tidal channels, and the fine detrital sediment was accumulated in a restricted, shallow inshore zone, thereby bringing carbonate precipitation to a halt (Walker *et al.*, 1983). This process was enhanced by storms which carried sand and clasts landward, bypassing the carbonate flat, and reworking some of the sediments. Immediately south of the Witvlei Ridge, this coastal clastic belt is about 5 km wide, while in the southeastern portion of the study area (Figs 4.2 and 4.9) it has a width of between 30 and almost 80 km.

Interpretation of the Okambara Member: The member accumulated in a channelled tidal flat environment that was frequently swept by storms. Indicators for various hydrodynamic regimes place the lower unit in a lower energy, subtidal to intertidal zone and the upper unit in a higher energy environment with increasing emergence toward the top. The middle unit is a clastic channel-fill succession. The overall energy level of the tidal flat could either have remained unchanged during Okambara times in which case only the tidal zones exposed in the study area shifted with progradation (shallowing), or the energy of the whole peritidal zone may have increased by subsidence of the hypothetical barrier west of the Witvlei Synclinorium, which apparently existed during deposition of the preceding members. The increasing clastic content of the Okambara Member heralded an approaching major environmental change after which the siliciclastic facies of the overlying Weissberg Member was deposited.

The clastic facies of the Okambara Member was laid down between the mainland situated in the east and northeast, and the carbonate tidal flats in the west and north respectively. The Witvlei Ridge formed a westward-projecting peninsula which divided the southern basin from the northern basin (Fig. 4.9). The supply of sediment from the Witvlei Ridge was negligible, as can, in all likelihood, be concluded for the mainland to the southeast. This implies that most of the clastic material in the basin south of the Witvlei Ridge was derived from the south and was transported by longshore currents, and/or from the shelf in the west from where it was transported by tidal currents and storms onto the shore, bypassing the carbonate facies through tidal channels. This depositional mechanism of the clastic facies is similar to that of the Wadden Sea where most of the detritus is brought in from the North Sea through tidal inlets by tidal currents (Bartholdy and Pheiffer Madsen, 1985; Van Straaten, 1978).

The quadrimodal dispersal pattern of cross-stratification (Fig. 4.9) is characteristic of tidal flats in general, and bimodal patterns with directions 90° apart point either in the direction of the slope or perpendicular to it, depending on the flow regime (Klein, 1967). A plot of all crossbedding directions (Fig. 4.9) presents two main and opposing vectors, which are aligned about a perpendicular to the trough axis, and a subordinate vector, directed into the basin. None of the dispersal patterns of palaeocurrent vectors has a relation to the source area of the sediments which is characteristic of peritidal environments (Klein, 1967).

6.2.4 Résumé of the Buschmannsklippe Formation

The Buschmannsklippe basin is not a continuation of the underlying Court basin. A major unconformity separates the two successions. The evolution of the Buschmannsklippe basin was initiated by a far-reaching transgression from the west and northwest that drowned a peneplain extending over most of the study area. While the eastern margin of the emergent basin can be inferred, its northern and western boundaries are not defined.

Most of the sediments were accumulated in a shallow subtidal to peritidal environment. The unchanged isopach patterns during deposition of the shallow water sediments of the three members (Figs 4.7 and 4.8), which reached a combined maximum thickness of about 350 m, indicate that sedimentation rate was in general accordance with basin subsidence. Slight deviations from this equilibrium between water depth and basin filling caused alternation of several hydrodynamic settings, resulting in repeated facies changes. The environment of the study area passed from intertidal to subtidal during deposition of the uppermost Bildah Member. During the lower Okambara Member, regression started which also enlarged the Witvlei Ridge towards the east and north, linking it with the Kalahari Craton, and thus caused bipartition of the basin and its facies belts.

The restricted tidal environment that prevailed during Bildah and La Fraque times required a barrier or shoal to the west of the study area. Its subsidence during upper Buschmannsklippe time might have been responsible for the increase of the energy level of the Okambara Member environment.

The percentage of clastic material increased upwards and the depositional environment changed from a carbonate platform at the base to mixed carbonate/siliciclastic but sand-dominated platform towards the top. In the Okambara Member, two facies belts can be recognised, a sandy coastal belt, and an offshore platform. The increase in detritus received by the basin probably heralded deposition of the siliciclastic Weissberg Member. However, the fine-grained siliciclastic material was not derived from the Kalahari Craton to the east but was transported by storms and/or currents from the south and west. That does not mean that the Kalahari Craton was not the original source of the sediments, but that their transport direction was influenced by currents, storms and tides. The distribution pattern of the clastic fraction, together with indications for a southern provenance area, led to the assumption that the Buschmannsklippe basin and the depositional province of the fluvial clastic sediments of the basal Kanies Member further south (Germs, 1983) could be linked. While in most parts of the study area, as in the Aranos basin further south, the lower contact of the Nama Group is a distinct unconformity, a transition between the Okambara Member of the Witvlei Group and the Weissberg Member of the Nama Group seems to exist in a portion of area 2218 CA.

Buschmannsklippe sediments formerly occurring to the west and northwest of the Witvlei Synclinorium have either been removed by erosion or were overridden by nappes of the Damara Orogen. The facies distribution in the study area, however, indicates that the basin extended towards the west and might even have opened in this direction, i.e. the Buschmannsklippe basin would represent a marine platform of the Damara ocean.

In the study area the unconformity between the Court and Buschmannsklippe Formations can be implied by the overstepping of the latter over the former. The sequence boundary between both could be of type 2 (Van Wagoner *et al.*, 1988). The marine flooding surface in the lower most part of the Weissberg Member overlying the Buschmannsklippe Formation (see below) forms the upper sequence boundary. A locally conformable contact with the Weissberg Member can be interpreted as a correlative conformity of the basal Nama unconformity.

6.3 DABIS FORMATION

6.3.1 Weissberg Member

A far-reaching transgression at the beginning of Weissberg times extended the depositional basin beyond the limits of the study area. The Witvlei Ridge, although submerged, still influenced deposition and divided two different arenaceous facies, viz. the orthoquartzite of facies (a) south of the Witvlei Ridge, and the subarkose of facies (b) north of the Witvlei Ridge. Generally, the purely siliciclastic sediments are thickest in the northwest and become thinner towards the south and east (Fig. 4.10).

The conglomerate at the base of the member in the Kehoro area **(facies (c)**) extends for 1 0 km in a north-south direction; its northern continuation is not known, however. This shape, together with the granulometric results, indicates that facies (c) is a delta fan.

Facies (b) is a submature quartzite. Its grain size decreases and sorting increases southwestwards from the Kehoro area. Coarse grains and small pebbles are concentrated on bedding planes or in the upper portion of layers. This results from winnowing by the action of strong currents which transported the finer size classes southwestwards across the barrier formed by the Witvlei Ridge and deposited them in the central and southern portion of the Witvlei Synclinorium as the orthoquartzitic lithofacies (a).

A thin to thick, parallel stratification is the main sedimentary structure observed within facies (b) of the Weissberg quartzite in the northern Witvlei Synclinorium, suggesting distribution and deposition by unidirectional currents rather than by tidal currents. Grain-size distribution indicates transport from northeast to southwest.

In the Gobabis Synclinorium, facies (b) comprises planar and trough crossbedded strata. Bimodal disperal classes (vectors 1800 apart) reveal northeastern and southwestern current directions (Fig. 4.1 0; for exact localities see Table 6.1), indicative of tidal flats and nearshore environments (Klein, 1967). Mud-gall conglomerates are widespread.

Where the Weissberg Member transgresses onto the Kamtsas Formation on the Witvlei Ridge, it contains scattered pebbles up to 7 cm across. These were derived from the conglomerate in the uppermost portion of the Kamtsas Formation. Cumulative curves of grain-size distribution in the Weissberg quartzite from the Gobabis Synclinorium are very similar to those in the northern Witvlei Synclinorium (Figs 5.15 and 5.16). However, only in the range of the fine grain sizes do curves from these two areas coincide with relevant curves from the central and southern Witvlei Synclinorium (cf. Fig 5.13). This, again, shows that coarser grain classes were concentrated in the northern Witvlei and the Gobabis Synclinoria, relative to the central and southern Witvlei Synclinorium. From this one can infer that even in the east, i.e. in the Gobabis Synclinorium, fine sediment was winnowed and deposited further west. It is not clear whether the southerly provenance area that supplied the clastic material of the Okambara Member was still present during Weissberg times.

Sedimentary structures in the lower most three to eight metres of **facies (a)** in the central and southern Witvlei Synclinorium are similar to those found in the uppermost arenite of the upper unit of the Okambara Member, i.e. mud-gall conglomerates, desiccation cracks (Plate 6.10), wave ripples, ripple lamination, low-angle planar and subordinate trough cross-stratification and herringbone cross-stratification (Plate 6.9). Irregular to angular vugs, up to 2 cm across, may have been left by weathered evaporite minerals. These features imply deposition in an intertidal environment with currents in opposite directions and frequent subaerial exposure. Tidal currents are also attested by the generally bimodal dispersal pattern of cross-stratification (Fig. 4.10). Trimodality, with directions 1800 and 900 apart, is interpreted as a rudimentary quadrimodal pattern.

The main succession of facies (a) of the Weissberg Member which overlies the lower most portion, is thicker-bedded and devoid of most of the sedimentary structures found in the basal portion. It is however, characterised by an abundance of ripple marks, most of which are symmetrical. Straight-crested to sinuous wave ripples are plentiful on many bedding planes. Interference ripples reveal changing wind or current directions (Plate 6.11). Periodic shallowing of water is indicated by the following:

- ladder ripples (Plate 6.13), caused by two ripple directions, in many cases with different wavelengths;
- flat-topped ripples (Plate 6.11) with scoured crests;
- double-crested, parallel-trail ripples, reflecting reduction of wavelength during shallowing of water depth;
- wrinkle marks showing runoff into the troughs of transverse ripples;
- desiccation cracks due to subaerial exposure (Plate 6.11).

Locally ripple trains were destroyed by scours. Spindle-shaped to parabolic flute marks (Allen, 1984) on the lower surfaces of layers (Plate 6.14) are the result of erosion by currents; their noses point upstream. Remnants of wind erosion resemble flute marks, but appear on the upper surface of beds (Collinson and Thompson, 1982). The same features, also on the top of a layer, may be generated by currents in cohesionless sediments (Eastler, 1978).

All these sedimentary features are due to changing hydraulic conditions (Allen, 1984). Most of the ripples, irrespective of whether symmetrical or asymmetrical, have a wavelength of 2.0 to 4.0 cm and a height of 0.5 to 0.75 cm, resulting in a ripple index (Tanner, 1967) of about 5. Crests of longitudinal (symmetrical) ripples extend parallel to the waves, i.e. perpendicular to the water movement. Orientation of ripple crests was measured at many localities in the domain of facies (a) (Fig. 5.23). They trend mostly between north-northwest and northeast, approximately parallel to the inferred coastline, but other

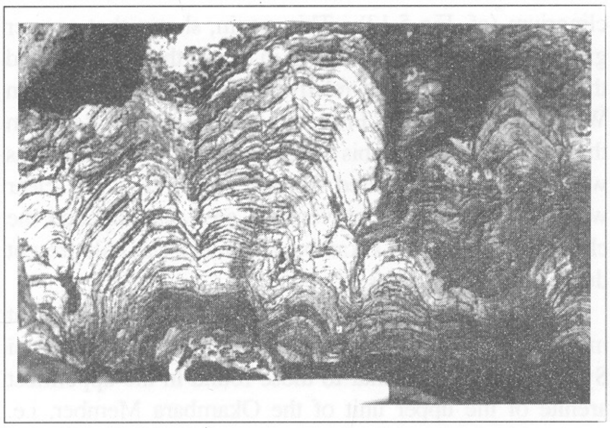


Plate 6.7: LLH-C stromatolites with indistinct branching near the base. Desiccation cracks and micro-unconformities are visible in the laminae of the stromatolites. Upper unit of the Okambara Member. Eastern portion of Arnhem 222. About 10 cm of the pen is visible

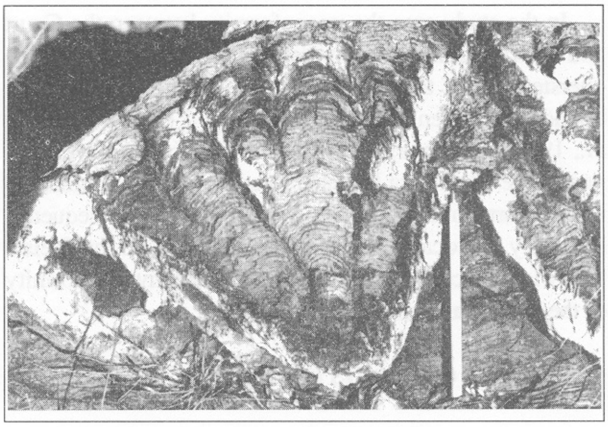


Plate 6.8: "Walled bioherm" (Bertrand-Sarfati and Moussine-Pouchkine, 1988) of branching stromatolites, surrounded by flat-laying algal mats. Upper unit of the Okambara Member. Okambara 219

Plate 6.9: Herringbone cross-stratification in the lowermost Weissberg Member. Otjombondona 225. About 8 cm of the pen is visible

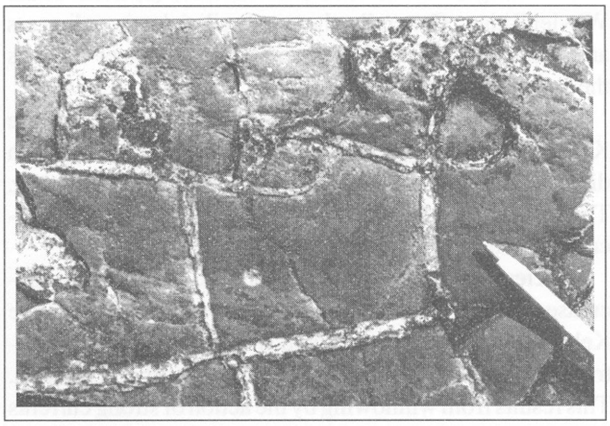


Plate 6.10: Desiccation polygons on bedding planes of the lowermost Weissberg quartzite. Nautabis 268. About 6 cm of the pen is visible

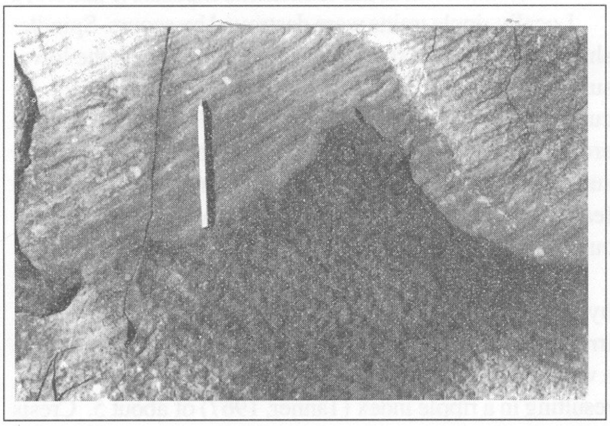


Plate 6.11: Interference ripples with superimposed desiccation cracks in upper layer and flat-topped ripples with truncated crests in lower layer. Both structures are caused by gradually shallowing water. Weissberg Member, facies (a). Scheidthof 293. Pen is 14 cm long

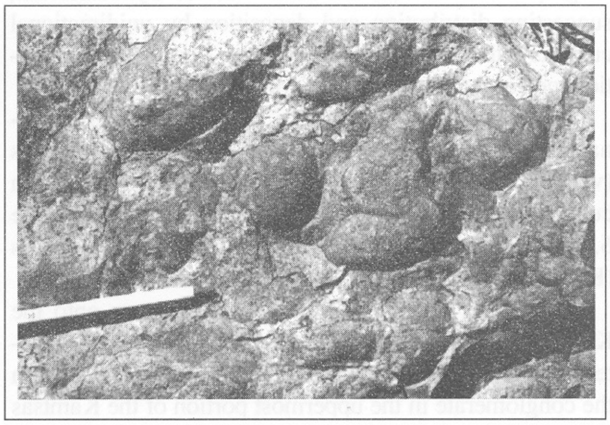


Plate 6.12: Load casts at the base of a sandstone layer. Zenana Member. Kuduberg 60. Visible part of pen is 9 cm long

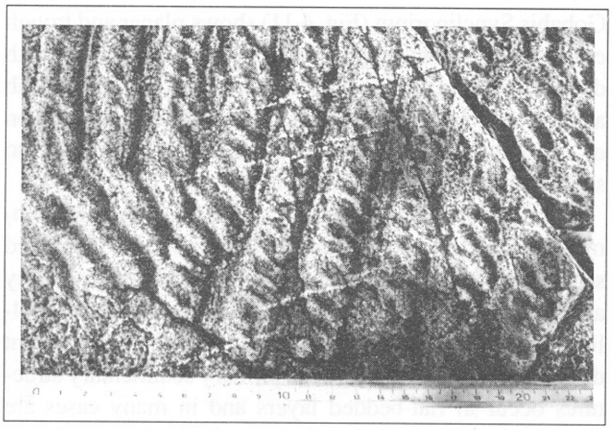


Plate 6.13: Interference or ladder-backed ripples superimposed on transverse ripples. Weissberg Member, facies (a). Okambara 219. Scale in centimetres

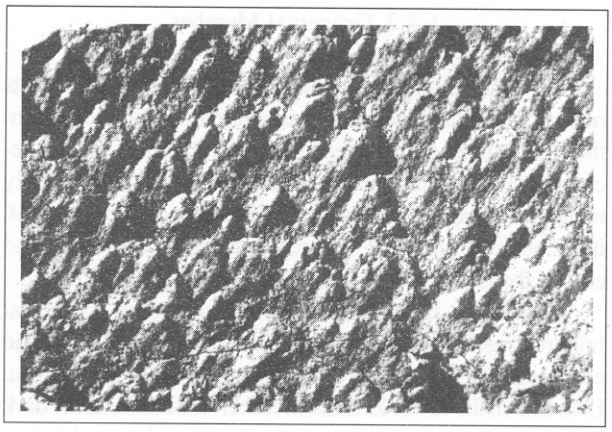


Plate 6.14: Parabolic flute marks on underside of layer; current from upper right to lower left. Weissberg Member, facies (a). Aida 296. Figure is about 25 cm wide

directions are also numerous.

However, the orientation of wave ripples in the intratidal environment is not meaningful for elucidating the provenance area of the Weissberg sediments.

Interpretation: The sedimentary structures and their vectors attest to deposition of facies (a) in the intertidal and shallow subtidal zones in a protected environment in which the fine clastics, brought in from a northerly and easterly direction, were concentrated. A fluvial to deltaic origin is obvious for facies (c) in the extreme north of the study area. Facies (b) has an intermediate position, displaying, in the northern Witvlei Synclinorium, the effects of strong currents either of fluviatile or of longshore marine nature. Separation of grain sizes in facies (a) and (b) is the result of a barring effected by the Witvlei Ridge which impeded southward transport of coarse detritus. The facies differences between the southern Witvlei Synclinorium (facies (a)) and the Gobabis Synclinorium (facies (b)) may be due either to a hypothetical barrier between the synclinoria which acted in a similar way to the Witvlei Ridge, or to the distance of about 50 km between the synclinoria which was sufficient to allow effective sorting by currents.

During Dabis times, as during deposition of the Buschmannsklippe succession, sedimentation kept pace with the rate of subsidence. The depocentre was also in the region west to northwest of the Witvlei Synclinorium. Thus, one can conclude that the basin of the Dabis Formation was structurally and probably also chronologically continuous with the Buschmannsklippe basin; it may have extended far to the west (Fig. 4.10) and was possibly connected with the Damara trough. Most of the clastic material was derived from the Kalahari Craton. The provenance area of the only locally developed facies (c), however, lies to the north, which indicates that a shoal separated the Nama basin and Damara trough, possibly as a result of incipient uplift within the Damara trough.

The sequence boundary between the Witvlei and Nama Groups lies within the lower most sandstone of the Weissberg Member, where the sedimentary structures in facies (a) indicate a reversal from shallowing of the upper Witvlei Group to deepening of the basin, resulting in the flooding of the Nama Group.

6.4 ZARIS FORMATION

6.4.1 Zenana Member

The base of the member is defined by an interruption of the arenaceous sedimentation and by deposition of dark - coloured carbonate (= **lower unit**) in the Witvlei and western Gobabis Synclinoria, and of dark shale in the Netso Syncline. Absence of all features characterising an intertidal environment with periodical subaerial emergence indicates that the carbonate most probably formed under shallow subtidal conditions. The lower unit is interpreted as a parasequence.

In the **upper unit**, siliciclastic sedimentation reoccurred, producing arenites similar to those of facies (a) of the Weissberg Member, but more mature. In the southern Gobabis Synclinorium and in the Netso Syncline, the arenites contain abundant clay-gall conglomerate, as well as desiccation cracks, asymmetrical ripple marks (currents from east), parting lineations (localities 46 and 48 of Fig. 4.11) and load casts (Plate 6.12). Most of these structures indicate a shallow-water environment with intermittent sub-aerial exposure. In accordance with the underlying and overlying members, the siliciclastic material is considered to have been derived from an easterly to northerly source area which was further away than it was during Weissberg times. This requires longer transport leading to better sorting and a uniform facies. The only indication of a nearby provenance was found in the northern Witvlei Synclinorium. Here the quartzite of the Zenana Member thickens to a coarser-grained, laterally confined body on the northwestern margin of the Eindpaal Syncline. This unit is interpreted as a proximal delta fan. Fine-grained quartzite and siltstone in the same horizon at the southeastern margin of that syncline are regarded as the distal portion of this fan (Fig. 4.11). It is certainly not coincidental that the fan deposits in the lower most Weissberg Member and in the Zenana Member occur in the same region, thus indicating a similar transport direction. They probably both belong to the same depositional system, implying a source in a region which was situated near or beyond the northern margin of the present Nama basin.

6.4.2 Grünental Member

The basin extended beyond the study area, and the limits of the Grünental depositional area remain unknown. From the distribution of the arenaceous components which decrease in size and proportion from east to west, and inversely to the percentage of pelitic and carbonate sediments present, it is concluded that the source of the clastics was situated east of the Gobabis Synclinorium on the Kalahari Craton.

Facies (a): Lenticular carbonate zones, up to several hundred metres across, are irregularly intercalated in pelitic clastics in the Witvlei and the southern Gobabis Synclinoria. Like that in the lower Zenana Member, the carbonate is dark in colour but in contrast to the latter it is nearly pure limestone. The thin to massively layered carbonate zones contain fine-grained clastic intercalations. Thin lamination, accentuated by chert seams and reminiscent of cryptalgal lamination, occurs sporadically. Massive dark limestone with local algal lamination, patches of rudstone and abundant Cloudina (Germs, 1972b) occurs near the homestead on the farm Suliman 215 (2218 CA) and is interpreted as reef talus. It is possible that bioherms were an essential constituent of the carbonate lenses in the Grünental Member. The extensive Driedoornvlakte bioherm further south (2316 DC) (Germs, 1983) occurs in the same horizon. Dark, indistinctly bounded patches within the grey limestone which are up to one centimetre across, are interpreted as thrombolites by J. Grotzinger (1990, pers. comm.).

Radiating calcitic structures in thin limestone layers interbedded in shale on Grünental 151 were interpreted as archaeocyatha and named *Buschmannia roeringi* by Kaever and Richter (1976). The organic origin was refuted by Debrenne and Lafuste (1979) who stressed the typically spherical shape of the structures and maintained that the calcite is pseudomorphic after gypsum. Glaessner (1980) supported this interpretation. In thin section the outlines of the spherulites and the free-ending, unrestrained bladed calcite crystals are clearly visible (Plate 5.13). Length-slow chert in spherulites from the locality of the pseudo-fossils may also be pseudomorphs after evaporite minerals.

Deposition in an offshore to subtidal lagoonal environment is envisaged for most of the dark limestone.

Facies (b): Fine sandstone of facies (b), exposed in the Netso Syncline and interfingering with facies (a) in the Gobabis Synclinorium (Fig. 4.11) shows planar and trough cross-stratification, ranging from large-scale troughs with dimensions of several metres to micro cross-lamination (rib and furrow). In the centre of the Netso Syncline orientation of cross-stratification reveals transport in a westerly to southwesterly direction (Fig. 4.11).

Parting lineations (primary current lineation) are common in the fine-grained and thinly layered to laminated sandstone of facies (b) in the Gobabis Synclinorium. They are formed under high flow velocity in shallow water (Collinson and Thompson, 1982) striking parallel to the current direction (Allen, 1984). These primary sedimentary structures occur in flat-bedded layers and in many cases are combined with a secondary feature, viz. parting-step lineation, caused by breaking of the laminae parallel to parting lineation. The majority of the parting lineations in the Black Nossob valley strike approximately east-west (Fig. 4.11; for localities see Table 6.1). If evaluated in combination with crossbedding a main transport direction from the east and northeast can be inferred (Fig. 4.11). Parting lineation has been observed in several environments (Allen, 1984), but is most abundant in fluvial sediments (Potter and Pettijohn, 1963). Indistinct flute marks, ripple marks and clay-gall conglomerates also occur in places.

The sedimentary structures, the unidirectional palaeocurrents and the absence of features of the peritidal zone indicate that facies (b) was deposited in a fluvial environment. Westwards, it interfingers with shallow marine sediments of facies (a).

Facies (c): Dark carbonate accompanied by subordinate shale and siltstone reaches a considerable thickness. The carbonate is either thickly layered to massive or, because of intercalation of lighter-coloured strata, has a thin-bedded to laminated stratification (Plate 4.12). Locally, the laminae show desiccation cracks and contain intraclasts. The clasts are either flat pebbles, having undergone short transport only, or are rounded, up to 1 cm thick and 5 cm long, oriented parallel or oblique to layering, and consist of light-coloured carbonate. The sedimentary features reflect short-lived, intermittent reworking in an environment that was starved of siliciclatic input. Deposition in an intertidal lagoon is envisaged. Scarcity of clastic debris can be explained by the distance from the source area in the east. Facies (c) is the equivalent of the Omkyk Member in the Aranos Basin.

7. METAMORPHISM

The mineral associations indicate that the Nama rocks in the study area were subjected to very low grade meta-morphism. These associations could also be produced by diagenesis (Winkler, 1976). In pelitic rocks, fibrous chlorite is the main secondary mineral, while diagenetic or metamorphic feldspar and quartz are very subordinate constituents in coarser-grained sediments. Illite content in five pelitic samples from the Constance, La Fraque and Grünental Members was too low for extraction.

Two types of metamorphism can be distinguished:

(a) The rocks of the Witvlei and lower Nama Groups in the study area were buried by the upper Nama strata and probably by still younger erosion products of the Damara Orogen (postulated by Horstmann, 1987). After denudation to almost the present level, Karoo rocks covered the area. The stratigraphic well Masetlheng Pan-I in western Botswana intersected a thick sequence of sandstones below the pre-Karoo unconformity at 1160 m. Of these the uppermost 1064 m were correlated with the Fish River and Schwarzrand Subgroups. The maturation gradient shows that nearly 2000 m of sediments were removed by erosion during post-Lower Permian times (Stoakes and McMaster, 1990), resulting in an originally nearly 4000-m-thick pile of overburden above the basal Nama sediments.

(b) The final regional metamorphism of the Damara Orogen, about 530 million years ago, affected the adjacent terrains of the foreland. It did not exceed very low grade (anchi-) metamorphism along the northwestern margin of the Aranos basin further south; but its intensity decreases from northwest to southeast (Ahrendt *et al.*, 1977; Weber *et al.*, 1983; Horstmann, 1987). The transition from very low grade metamorphism to diagenesis coincides with the southeastern border of the folded Nama Group (Ahrendt *et al.*, 1977). If applied to the study area it means that the Witvlei Synclinorium would still fall into the belt of very low grade metamorphism whereas in the Gobabis Synclinorium mainly diagenesis can be expected.

Using the calcite-dolomite geothermometer, Kasch (1983) calculated peak metamorphic temperatures of 380°C and 343°C for two carbonate samples from Grünental 151 and Owinieikiro 213 respectively. Both samples were collected near the southern boundary thrust of the Southern Margin Zone of the Damara Orogen (Fig. 3.1) where the highest temperatures affecting Nama rocks in the study area can be expected.

8. DEPOSITIONAL AND TECTONIC HISTORY OF THE STUDY AREA

The sedimentary rocks of the study area reflect the orogenic and depositional history of the southern margin of the Damara trough and its foreland. While the Tsumis and Nosib strata and possibly also the Court Formation of the Witvlei Group were deposited during the rifting stage of a passive margin, the Buschmannsklippe Formation of the Witvlei Group was deposited during the drifting stage, i.e. in a flexure basin of the passive margin in the sense of Klein (1991). The basal Nama unconformity is believed to indicate the reversal from an extensional to a convergent basin.

The Witvlei Group was deposited along the northern and western edge of the Kalahari Craton on the southern shelf of the Damara trough. Similar strata have been found only in the Naukluft Nappe Complex and in the Rosh Pinah-Witpütz area (Hoffmann, 1989). During the middle Damara period the basin of the passive margin was situated further north- and westward than the basin of the active margin stage during the upper Damara, when the depositional area extended south- and eastward due to loading of the foreland. Deposition of the Witvlei Group took place mainly in areas where today basement highs separate the internides of the Damara Orogen from the externides, or which were over-ridden by nappes.

The oldest tectonic structures of Damaran age are post-Nosib-pre-Witvlei faults that accompanied up warping of the southeastern margin of the Damara trough (Table 8.1). In this region, which included the area of the later Witvlei Synclinorium, Eskadron rocks were displaced against Nosib sediments. The throw on some faults amounted to several thousands of metres. Intense erosion removed the Nosib quartzite on the upthrown fault blocks. This tectonic zone can be followed from east of Rehoboth to the region northeast of the study area and could indicate the latest stage of rifting. Most of the eroded material was transported northwestwards into the Damara basin, while only a minor portion found its way southeastwards into the Court Formation basin to form the quartzite of the Simmenau Member.

Some open, megascopic folds in the Kamtsas Formation can be recognised on aerial photographs in area 2218 CA. They are discordantly overlain by the Buschmannsklippe Formation and therefore pre-date the latter.

Although the margins of the Court basin have not been clearly delineated, it probably formed an isolated, elongated depression extending in a northeast-southwest direction near the northwestern margin of the Kalahari Craton. Minor uplift occurred during early Simmenau times, possibly as the last pulse of rifting.

The Buschmannsklippe succession overlies the Court succession paraconformably. They are separated by an interval of unknown duration which was a period of tectonic quiescence. With the onset of deposition of the Buschmannsklippe Formation, a new sedimentary cycle started. A period of slow (eustatic) subsidence began which proceeded on a peneplaned floor from the west and northwest. A barrier is postulated to have separated the Buschmannsklippe basin from the Damara trough. It however had no influence on the sedimentary environment during deposition of the upper Okambara Member.

Sedimentation of the Witvlei Group was terminated by a regression. A new sedimentary cycle started with the far-reaching transgression of the Nama Group. Its depositional area extended beyond that of the preceeding group. Therefore the Nama Group nearly everywhere overlies the older rocks unconformably, with a possible exception in the central and southern Witvlei Synclinorium. The isotopic composition of carbonates from the upper Witvlei Group and the lower Nama Group are similar (Kaufmann *et al.*, 1991). These authors believe that the unconformity which separates both groups represents a significant hiatus. While a hiatus at the base of the Nama Group is evident for most of the Nama basin, marking a sequence boundary, an apparent transitional contact between the Buschmannsklippe Formation and the Dabis Formation in parts of the study area can be interpreted as a correlative conformity in a region where the marine flooding surface of the Nama was preceded by shallowing only, without subaerial erosion.

The basal unconformity marking the transgression of the Nama Group can be found nearly everywhere in the Nama basin from the Gobabis area to the northern Cape and is believed to mark the reversal of plate motion from spreading to convergence, i.e. the lower Nama Group was deposited during initial closing of the Damara ocean under the load of the approaching thrust wedge (Stockmal *et al.*, 1986). The latter, however, remained below sea level and did not shed sediments into the foreland basin.

During deposition of the Dabis and Zaris Formations, clastic material was mainly derived from the Kalahari Craton in the east. With the possible exception of a small fan deposited during Dabis and Zaris times in the northern part of the study area no source area for the clastic sediments lying within the Damara Orogen is known. This indicates that the closure of the Damara ocean had not advanced far during lower most Nama times. Deposition of alternating clastic and carbonate units during the upper Witvlei and lower Nama periods is mainly ascribed to minor tectonic pulses affecting the hinterland rather than the basin of deposition which, in the study area, remained for most of that time within the peritidal and shallow subtidal zones.

The Nama sediments probably did not extend eastward beyond the Kalahari Line, situated in central Botswana (Reeves, 1979; Meixner, 1984a; see Fig. 1.1). In Masetlheng Pan-l well in the Nosop-Ncojane Basin in western Botswana (Fig. 1.1) the base of the Karoo Sequence at 1162.7 m below surface is mainly underlain by red beds of terrestrial origin (Stoakes and McMasters, 1990). A probably marine incursion between 2223.5 and 2284.5 m below surface consists of siltstone and shale and yielded a Vendian microfossil assemblage resembling samples from the Kuibis Subgroup (Knoll in Dolby, 1990). The 1060.8-m-thick interval of sandstone between those Kuibis rocks and the base of the Karoo is interpreted to represent the upper portion of the Nama Group: the Schwarzrand Sub-

Period	Physiographic record	Tectonics
Quaternary to Cretaceous	Kalahari deposits; Nossob River; Erosion of Karoo and upper Nama rocks (3000 to 4000 m)	NE/SW normal faults, in continuation of the East African Rift system
Jurassic to Carboniferous	Depositon of Karoo Sequence Dwyka tillite in proto-Black Nossob valley	
Cambrian to Namibian	Closure of the Damara ocean, thrusting of the rocks of the SMZ over the foreland	D3 of the SMZ: Doming of the foreland in large-scale antiforms and synforms. D2: Minor phases of folding deflects the axes of the first phase folds; slaty cleavage; folding of the fault planes. Late D1 or early D2: Reverse faulting along rejuvenated post-Kamtsas/pre-Witvlei faults, accompanied by formation of fracture cleavage. D1: Open, symmetrical and locally overturned folds in the western part of the foreland.
	Depositon of the Schwarzrand and Fish River Subgroups	
Zaris Formation	Basin widens; fluvial at the margins to marine towards the centre	Development of a foreland basin
Dabis Fm.	Fluvial to shallow marine Nama transgression	Active margin
Okambara Memb.	Regression; intertidal to supratidal	Passive margin, drifting stage
La Fraque Memb.	Subtidal-lagoonal	Eustatic subsidence
Bildah Member	Intertidal Transgression from W to NW	
Simmenau Memb.	Filling of basin with fining-upward fluvial arenites; local erosion and alluvial fans	Minor tectonic pulse in lower part with uplift in the central basin
Constance Gobabis Members	Filling of basin; sediments change from "deep-water" carbonate to shallow-water fine clastic	Tectonic quiescence
Pre-Witvlei/ post-Kamtsas	Deep erosion and peneplanation	Upwarp of the southeastern margin of the Damara trough; blockfaulting in the foreland; local weak folding and fracture cleavage
Blaubeker Fm.	Patchy tillite and mixtite in the S	
Kamtsas Fm.	Mainly fluvial sandstone	Rifting stage
pre-Kamtsas, post-Eskadron		Deformation (moderate folding and tilting in the W)
Eskadron and Doornpoort Formations	Mainly clastic sediments in different environments (alluvial fan, fluvial, lacustrine, marine)	Main rifting stage of passive margin

Table 8.1: Depositional and tectonic history and morphogenesis of the study area

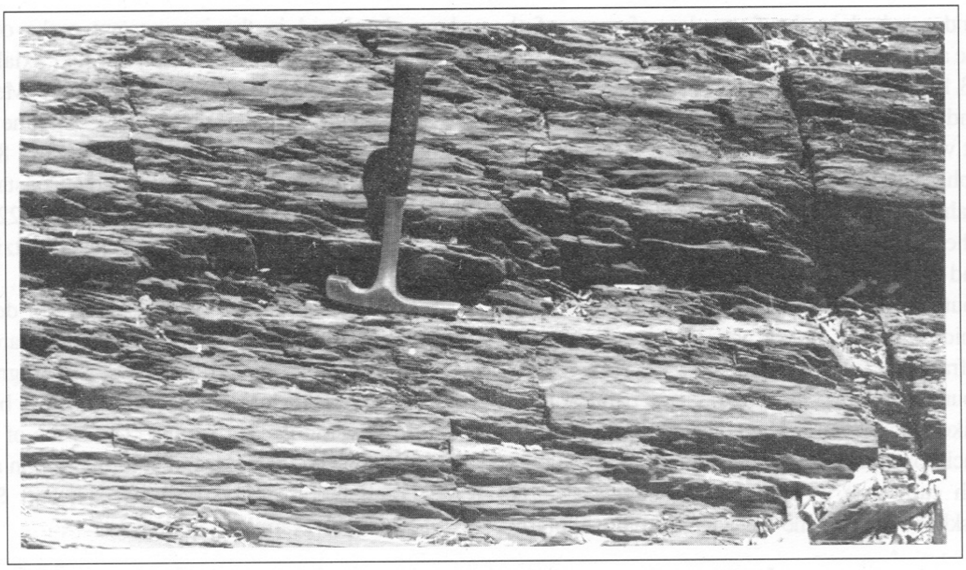


Plate 8.1: Thin-layered, calcareous siltstone of the La Fraque Member dipping 20° away from observer. It is cut by a cleavage which runs parallel to the axial planes of megascopic folds. The cleavage dips steeply towards the viewer and intersects bedding at an angle of 20°. Bildah 220

group, in a sandy, marginal facies and the Fish River Subgroup, the latter probably accounting for most of the thickness. Masetlheng Pan-l well was drilled near the landward margin of the Nama basin where all Nama sediments occur in an arenaceous facies. No trace of the Witvlei Group was found in the borehole as the passive margin deposits did not extend so far onto the Kalahari Craton. Sediments of that group, however, can be expected to have been deposited along the southeastern margin of the Damara trough in northern Botswana and might be preserved in the Passarge Basin. It can be expected that in the Witvlei-Gobabis area, which is further basinward than the Masetlheng borehole, the Kuibis Subgroup was overlain by rocks of the Schwarzrand Subgroup in basin facies and by thick Fish River Group arenites.

Ductile deformation during D_1 in the Southern Margin Zone, just prior to or during final closure of the Damara ocean, produced open folds in the foreland region bordering the southern boundary thrust. In the western portion of the study area, the folds are symmetrical and doubly plunging, open to tight and some are overturned to the southeast. Parasitic folds occur on the limbs of the first-order structures (Plates 3.1 and 3.2). In the east, the structures are asymmetrical but overturned folds are absent. Compared with the western area, their dimensions decrease. The different form of the folds in both areas may be enhanced by variations in types and thickness of the rock units present. Originally the fold axes were aligned southwest-northeast but during a subsequent phase of less intense deformation (probably D_2), the axes in the northern Gobabis Synclinorium were deflected in a more easterly direction (Fig. 1.2). Rotated fold axes in the central Witvlei Synclinorium suggest a second phase of folding in this area as well (Plates 3.1 and 3.2).

Several of the antiforms and synforms are disrupted by reverse faults which formed during thrusting in the Southern Margin Zone. This suggests a late D_1 or early D_2 age for these faults. Generally, they are upthrown toward the southeast leading to duplication of strata. The zone of the northwestern boundary fault of the Witvlei Synclinorium reflects at least two phases of deformation. The oldest phase belongs to the post-Nosib-pre-Witvlei deformation mentioned above. The fault was rejuvenated by post-Zaris deformation (Table 8.1). This is especially obvious in the Kehoro area (2218 BA; Plate 3.3), where Buschmannsklippe rocks which overlie the Kamtsas Formation with a paraconformable contact and the Eskadron Formation with an unconformable contact, are fault-bounded against the Eskadron Formation and have been eroded from the block uplifted during post-Nama times. Faulted contacts between the Eskadron and Kamtsas Formations and between the Eskadron and Buschmannsklippe Formations attest to the two phases of movement. Near the boundary between the farms Nudom 161 and Kehoro 939, however, the post -Zaris throw becomes insignificant and the Buschmannsklippe rocks transgress over the pre-Witvlei fault from Kamtsas onto Eskadron Formation.

The northwestern boundary fault of the Witvlei Synclinorium, as well as parallel fault zones which divide the synclinorium into several fault blocks, can be followed for more than 150 km. The northwestern boundary fault converges with the southern boundary thrust of the Southern Margin Zone of the Damara Orogen for some distance (Fig. 3.1). There is no indication that the shorter (1- to 7-km-long) and more irregular reverse faults with throws not exceeding 100 m may be rejuvenated pre-Witvlei structures (a few of them are indicated in the western portion of Plate 3.1). Several are arranged *en echelon* in 10- to 15-km-long fault zones.

Several of the faults were probably folded during late D_2 deformation. The northwestern Kehoro Anticline (2218 BA; Plate 3.3) is bounded by a fault running parallel to the folded contact of the strata. On Nautabis 268 and Smalhoek 236 (2317 BB), reverse faults are probably folded around tight folds. Four boreholes which intersect the northwestern boundary fault of the Witvlei Synclinorium between Christiadore 104 (2218 AD) and Okasandu 158 (2218 BA) indicate that the fault surface changes its dip from 45°N to NW to 45°S to SE, probably by folding of the fault plane.

Large-scale doming in the Southern Margin Zone during D_3 deformation (Miller, 1983) extends onto the Southern

Foreland and resulted in the formation of most of the macroscopic antiforms and synforms in the study area (Fig. 1.2; Plates 3.1 to 3.3).

Slaty and fracture cleavages are not extensively developed but where present can in most cases be related to D_1 and D_2 phases of deformation. In the Okambara Member on Bildah 220 (2218 CA) a vertical cleavage is “axial planar” to the small-scale wavy lamination of domal stromatolites and was caused by D_2 movement on a thrust fault situated about 200 m further east. At the same locality the quartzitic middle unit of the Okambara Member is intersected by a fracture cleavage dipping 60°W.

A pronounced fracture cleavage nearly parallel to the regional strike is observed in the Kamtsas quartzite in the Witvlei Synclinorium. It does not continue into the overlying Bildah carbonate. Although this suggests a pre-Buschmannsklippe age for the cleavage, it may simply not have developed in the incompetent rocks of the Buschmanns-klippe Member. A well-defined axial-planar slaty cleavage is developed in the La Fraque Member of the central Witvlei Synclinorium, intersecting the siltstone and mudstone at a high angle to bedding (Plate 8.1). A weak cleavage was found in mudstone of the Constance Member on Kanabis 54 at the Black Nossob River (2218 DC), and on Styria 52 (2218 DC) the Simmenau quartzite is cut by a wavy cleavage which dips about 45°NW.

That there is a much younger phase of deformation is apparent by the occurrence of fine-grained Karoo sediments in area 2218 AD. According to aerial photograph interpretation they occur in a half-graben which is bounded by a fault on its northwestern edge. The strike of the structure parallels that of the older faults in the region and could be the continuation of the East African Rift System from Botswana into Namibia (Table 8.1).

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