Hallesches Jahrb. Geowiss.	B 18	1 - 21	Halle (Saale) 1996

Sedimentation and basin development of the Middle Proterozoic Doornpoort and Klein Aub Formations, Central Namibia

Sedimentation und Beckenentwicklung der mittelproterozoischen Doornpoort- und Klein Aub Formationen, Zentral-Namibia

With 10 figures

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Summary: The 1300-950 Ma old Sinclair Sequence at Klein Aub, central Namibia, is preserved between the Congo and Kalahari Cratons. It commenced with bimodal pyroclastic rocks and lava flows which partly filled a variety of graben-like basins. Siliciclastic and carbonate sedimentation dominated basin infill, developing an overall fining-upward sequence of sedimentary rocks. Deposition of conglomerates and compositionally immature arenites was initially controlled by fault activity but ultimately overstepped graben shoulders. Younger basin-wide sedimentation comprised quartz arenites and carbonaceous mudstones with subordinate limestones deposited in association with waning fault activity.

The coarse sandstones and conglomerates represent red-beds deposited on debris flow-dominated fans next to the graben fault scarps. Fan lenses are arranged in an en échelon fashion which suggest a strike-slip component to the marginal fault system. These fans passed axially into sandy braided stream systems and localised aeolian dunes and playa lakes. Evaporite pseudomorphs confirm arid climatic conditions.

The finer clastic rocks and carbonates are organized in up to seven fining-upward cycles passing from subtidal quartz-arenites to finely interlaminated and flaser laminated pyritic intertidal limestones. This mixed siliciclastic/carbonate sequence developed in a series of prograding cycles averaging 4 m - 5 m thick, which could imply a similar maximum tidal range of macrotidal proportions.

The basin developed in an extensional crustal regime with thermal uplift, followed by mechanical rifting, associated with volcanism and the deposition of coarse clastics. During the thermal subsidence phase a marine transgression gave rise to the tidal deposits of the later fine grained clastic sediments. Both tectonically and sedimentologically the Sinclair Sequence heralds the subsequent development of early Damara rifting.

Kurzfassung
Zwischen den archaischen Schildregionen des Kongo und Kalahari Kratons ist im Gebiet von Klein Aub, Zentral-Namibia, die etwa 1300 - 950 Ma alte Sinclair Sequenz südlich des Damara Orogens aufgeschlossen. Die Ablagerung dieser vulkanosedimentären Sequenz begann mit der Eruption bimodaler Pyroklastika und Laven, die, zusammen mit geringeren Mengen grobklastischer Sedimente, frühe Gräben und Halbgräben füllten (Nückopf Formation und Grauwater Formation). Während Sedimente im Vergleich zu vulkanischen Gesteinen in der anfänglichen Beckenentwicklung eine eher untergeordnete Rolle spielten, dominierten klastische Einheiten mit einem generellen 'fining-upward'-Trend den höheren Teil der Beckenfüllung. Die räumliche Verteilung der frühen Konglomerate und kompositionell wie auch texturell immaturen Arenite innerhalb der Gräben war in dieser frühen Phase (Grauwater Formation und z.T. Doornpoort Formation) primär störungskontrolliert. Die klastische Sedimentation der späteren Phase weitete sich deutlich über die Grabenränder aus (Doornpoort Formation und Klein Aub Formation). In dem sich nun ausbildenden, erweiterten Sedimentationsraum kam die Störungsaktivität weitgehend zum Erliegen; es kamen vor allem Quarzarenite, kohlenstoff-führende Silt- und Tonsteine sowie in geringerem Umfang Karbonatgesteine zur Ablagerung.

Bei den grobklastischen Gesteinen der Doornpoort Formation handelt es sich um kontinentale Rotsedimente, die sich vor allem als alluviale Schuttfächer vor Grabenrandstörungen akkumulierten. Dabei sind die im heutigen Erosionsbild linsenartig angeschnittenen Schuttfächer 'en échelon' angeordnet, was auf eine laterale Verschiebungskomponente der Randstörungen hinweist. Diese alluvialen Schuttfächer gingen axial in sandige 'braided stream'-Systeme über, stellenweise auch in Playa-Seen und äolische Dünen. Die hierdurch angezeigten kontinentalen, ariden bis hyperariden Klimabedingungen werden zusätzlich durch Pseudomorphosen von Karbonat und Quarz nach Evaporitmineralen angezeigt.

Die feinklastischen Gesteine und Karbonatgesteine der Klein Aub Formation lassen sich in bis zu sieben rythmische 'fining-upward'-Zyklen untergliedern. Diese gehen jeweils von der Basis zum Top von subtidalen

Quarzareniten in feinlaminierte und flaserig laminierte Siltsteine und z.T. in pyrit-führende intertidale Siltsteine und laminierte Karbonatgesteine über. Diese gemischt siliziklastisch-karbonatische Sequenz entstand in jeweils etwa 4 m - 5 m mächtigen Zyklen. Die Mächtigkeit der einzelnen Zyklen gibt einen Hinweis auf den maximal für die Entstehung anzunehmenden Tidenhub, der vermutlich in der gleichen Größenordnung, d.h. in 'makrotidalen' Dimensionen, gelegen haben dürfte.

Das Sinclair Becken in der Region von Klein Aub entwickelte sich in einem Bereich gedehnter kontinentaler Kruste. Dabei kam es zu thermisch bedingter Krustenaufwölbung, gefolgt von mechanischem 'Rifting', bimodalem Vulkanismus, mechanischer Erosion der gehobenen Grabenränder und Sedimentation des grobklastischen Erosionsschutts.

Es folgte eine Phase thermaler Subsidenz und des Einsackens der ehemaligen Grabenränder, was eine Ausweitung des Sedimentationsraumes und eine marine Transgression zur Folge hatte. Im Gezeitenbereich dieses marinen Beckens kamen feinklastische tidale Sedimente zur Ablagerung. Sowohl in Bezug auf den tektonischen Bau- als auch Sedimentationsstil. kündigt sich mit der Klein Aub Formation bereits die Entwicklung des frühen Damarasedimentationsraumes an, dessen zeitliche und räumliche Abgrenzung weniger deutlich ist als bisher vielleicht vermutet.

1 Introduction

The tectonic setting of the Sinclair Sequence has been interpreted by a number of authors. WATTERS (1974) suggested an island arc setting, which he termed the Rehoboth magmatic arc. This was based on the investigation of the geochemistry of certain intermediate volcanics. KRÖNER (1977), MASON (1981) and COHEN et al. (1984) tentatively inferred that the Sinclair basins formed in an extensional crustal regime. RUXTON (1986) and RUXTON & CLEMMY (1986) interpreted the sedimentary sequence as "a zone of molassic red bed deposits on the southern foreland of the Damara Orogenic Belt". BORG (1988a, b) infers a rift basin setting for the Sinclair Sequence. However, there is a marked contrast between the lower part of the Sinclair Sequence (Nückopf and Grauwater Formations) and the upper part (Doornpoort and Klein Aub Formations) and an angular unconformity between these two major units is described by Schalk (pers. commun.). Nückopf and Grauwater Formations are characterized by pronounced volcanism and contemporaneous plutonism in the surrounding areas with minor sedimentary units in the upper part. In contrast, Doornpoort and Klein Aub Formations comprise predominantly sedimentray units with subordinate amounts of mafic volcanics in the basal part.

None of the authors mentioned above have attempted to constrain in detail their tectonic models by the information derived from sedimentary facies analysis. It is the aim of this paper to describe the facies and to interpret the palaeoenvironmental changes, that will lead to an improved understanding of the basin evolution.

The study area is situated 180 km SSW of Windhoek (Fig. 1 A, B) in the vicinity of Klein Aub village. Here the Sinclair Sequence is a late Middle Proterozoic volcano-sedimentary succession, preserved over a strike length greater then 1000 km between the Congo and Kalahari Cratons on an Early to Middle Proterozoic igneous and metamorphic basement (S.A.C.S. 1980). Radiometric ages of the sequence vary from approximately 1300 Ma to 1000 Ma (S.A.C.S. 1980). The area has been affected by greenschist facies metamorphism, with temperatures not exceeding 350 °C during the peak of the Damaran Orogeny at 530 ± 10 Ma (AHRENDT et al. 1978; SCHNEIDER & BORG 1988). Sediment-hosted stratabound Cu-Ag mineralization, predominantly of epigenetic origin, occurs in dark pyritic fine-grained sedimentary rocks near a redox interface with underlying red beds (BORG 1988a, b, 1991, 1994; BORG & MAIDEN 1989; WALRAVEN & BORG 1992).

The study area has been previously mapped by DE KOCK (1934), HANDLEY (1965) and SCHALK (1970). THOMAS (1978) mapped equivalent beds in Botswana. SCHALK (1970) established the local stratigraphy and RUXTON (1981, 1986) and RUXTON & CLEMMEY (1986) undertook sedimentological studies of the upper portions of the sequence. BORG (1988a, b, 1994) and BORG & MAIDEN (1986a, b, 1987, 1989) integrated the tectonic, volcanic and sedimentary patterns. The present paper extends these tectono-sedimentary relationships and analyses the sedimentology of the Sinclair Sequence in greater detail.

2 Geological setting

The Klein Aub basin in which the Sinclair Sequence accumulated is an intercratonic basin along a zone of crustal weakness which developed between the Kalahari and Congo Cratons (MASON 1981; CAHEN et. al. 1984). The sequence is preserved on the northwestern margin the Kalahari Craton to the south of the Damara Orogenic Belt (Fig. 1 A, B). Crustal extension may have been caused by an underlying mantle plume (BORG et al. 1987; BORG 1988b). Caused by the subsequent Damaran Orogeny the rocks have been faulted and folded. The succession has been tilted, folded, has been the subject of thrust faulting, and dips at angles between 20° and 75° to the south (Fig. 1 C), exposing the entire sequence so that basement/cover relationships can be established in detail. The stratigraphy of the Sinclair Sequence is shown in Fig. 2.

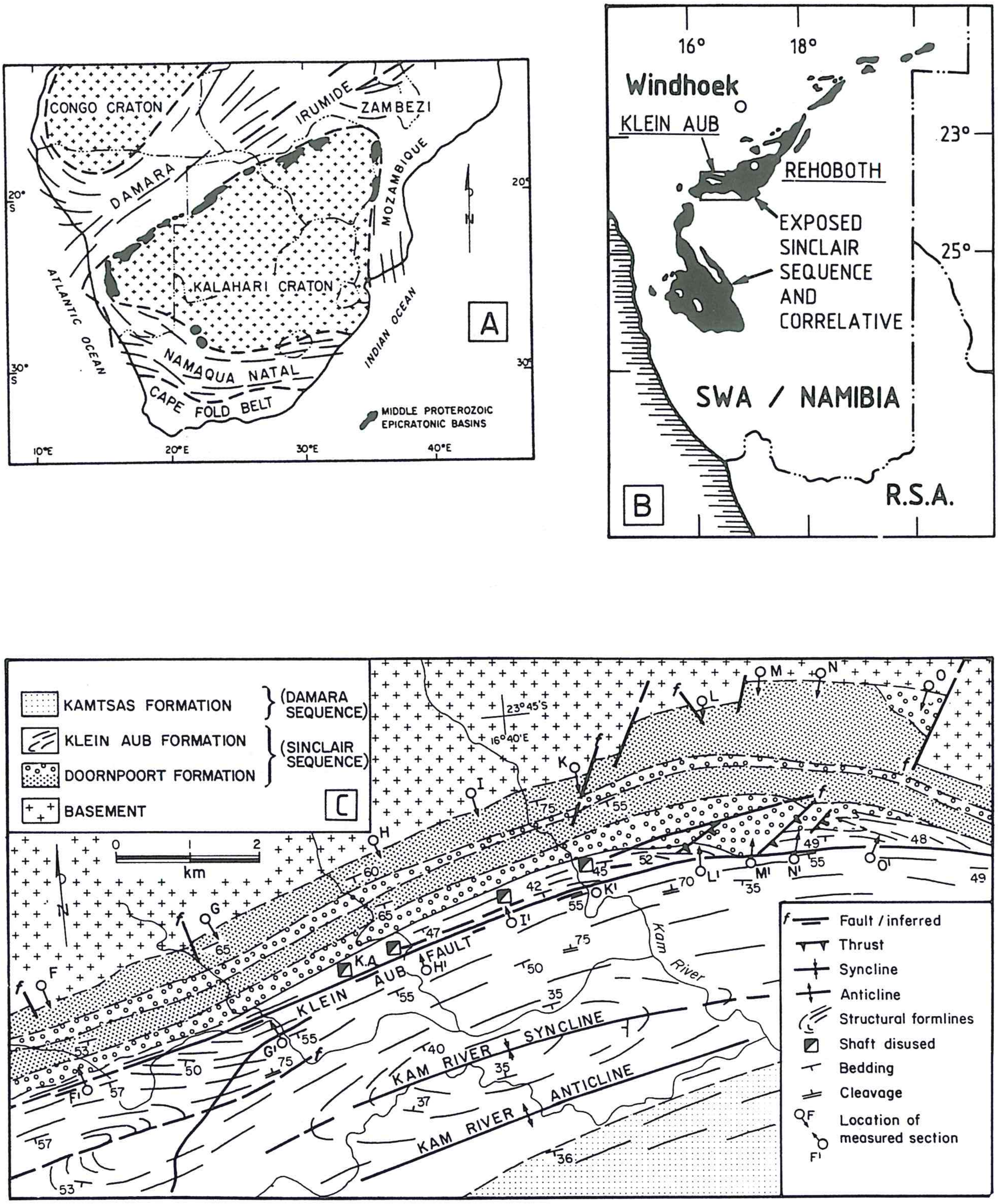


Fig. 1:
Geological location of the study area.
A: Distribution of Middle Proterozoic basins (Sinclair Sequence and correlatives) in southern Africa.
B: Exposed Sinclair Sequence in Namibia, with study area indicated.
C: Geology of the area around Klein Aub Mine. In this area the Doornpoort Formation overlaps the Nückopf and Grauwater Formations which are therefore not exposed.

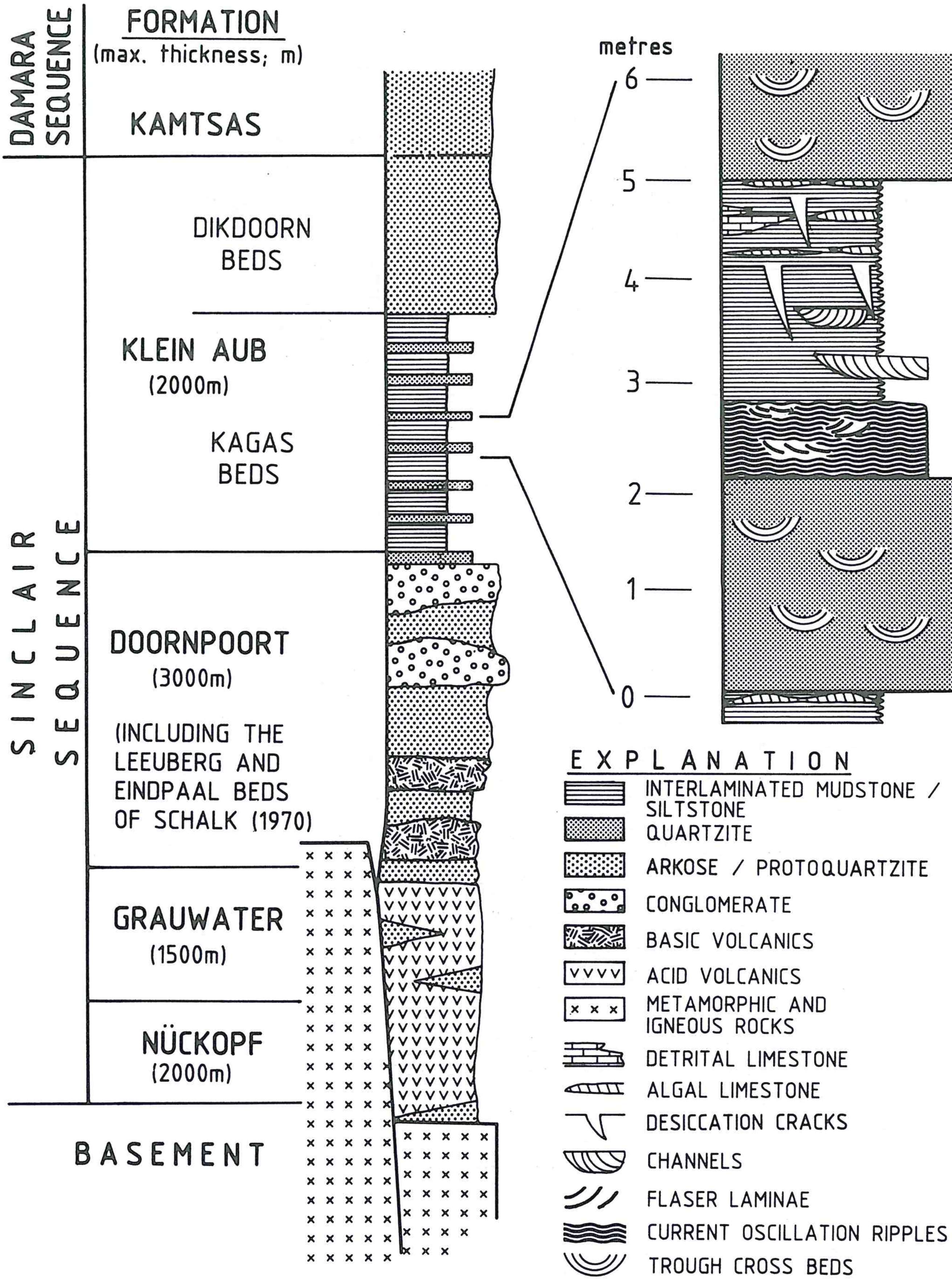


Fig. 2: Stratigraphy and lithology of the Sinclair Sequence in the study area.

2.1 The alluvial fan to Braid Plain Depo System of the Lower Sinclair Sequence (i.e. Nückopf and Grauwater Formations)

Interbedded Conglomerate and Arkose Facies

This facies comprises conglomerate beds 0.5 m - 2.5 m thick and interbedded arkose beds 0.1 m - 10 m in thickness. The conglomerates are poorly sorted with sub-angular to sub-rounded clasts varying up to 70 cm in diameter, averaging 8 cm (Fig. 3 A). They are mainly matrix-supported and polymictic in character. The clasts are derived from the following sources: Extrabasinal clasts include granite, granodiorite, vein quartz, subordinate dolorite and minor banded iron formation clasts derived from a mixed igneous and metamorphic Proterozoic basement. Intrabasinal clasts include red quartz prophyry and quartz-fedspar porphyry, derived from the Nückopf and Grauwater Formations which represent the volcanic infill of the initial rift. To a lesser extent intraformational clasts of protoquartzite and arkose are also present. The matrix is arkosic and is mineralogically and texturally immature. The conglomerates are generally massive and show no obvious imbrication features. Some units show both normal and inverse grading with largest clasts in the stratigraphic centre of the unit.

The interbedded arkose beds are 15 cm to 100 cm thick and comprise immature coarse sandstones to granulestones. They commonly contain scattered pebbles mainly of vein quartz and show planar bedding, rare trough cross-bedded sets averaging 80 cm thick and also plane bedding (Fig. 3 A, B). Palaeocurrents are directed to the south and south-west which is in agreement with measurements of RUXTON (1981, 1986). The arkose units are lensoid and of limited lateral extent.

The interbedded conglomerate and arkose facies as a whole vary in aggregate thickness from 100 m to 800 m (Fig. 4). The conglomeratic bodies display lensoid geometries with lenses extending 10 km - 20 km along strike and of about 500 m maximum thickness on average. The conglomerate and arkose facies occurs as a minor component of the Nückopf Formation and makes up almost 50% of the Grauwater Formation (Fig. 2). In both cases these facies are interbedded with felsic volcanic rocks. Conglomerates and arkoses are is most extensively developed in the Doornpoort Formation, where they are interbedded with the protoquartzite facies. The Doornpoort Formation overlaps older rift-fill and locally covers horst blocks (Fig. 4 B).

Within the Doornpoort Formation of the study area there are three major conglomeratic lenses developed, one of which is a basal conglomerate directly overlying basement rocks (Fig. 4 B). Another lens is directly related to an underlying subordinate graben (Fig. 5). The lenses are arranged in an en échelon fashion with their maximum thicknesses shifting progressively towards the east (Fig. 4 B). The mapped and measured sections in Fig. 4 A (from A to U) represent a longitudinal profile parallel to the edge of the rift basin. In these sections the interbedded conglomerate and arkose facies is well represented.

Protoquartzite/Arkose Facies

These medium- to coarse-grained arenites have feldspar contents varying between 15% and 40%. They are also texturally immature with poor sorting and angular to sub-angular grains. Erosional channel bases are developed within the facies and planar and trough cross-bedding are both common. Plane bedding also occurs. Palaeocurrents are unimodal to the south and south-west (RUXTON 1981, 1986 and author's observations). The facies geometry is tabular, with bodies generally 200 m - 300 m thick. Were they are underlain by local graben structures, however, their thicknesses increase to 700 m. Towards the basin centre the proportion of the protoquartzite/arkose facies increases considerably. The protoquartzite/arkose facies is extensively developed in the Doornpoort Formation and higher in the sequence in the upper part of the Klein Aub Formation as defined by SCHALK (1970).

Large Scale Cross-Bedded Quartz Arenite Facies

Locally developed, this facies occurs within protoquartzite/arkose facies in the lower portion of the Doornpoort Formation, where this facies comprises 30 m thick units of mineralogically and texturally supermature quartz arenite. The sandstone is well sorted, with well rounded grains and has a characteristically deep red colour. The only sedimentary structure present is planar cross-bedding with a maximum set thickness of 4 m. The planar cross-beds have a consistent thickness of approximately 10 cm and commonly display internal lamination. The facies was identified in an otherwise poorly exposed area. However, the limited exposure indicate that palaeocurrent directions appear to have been quite variable. No complete dune foresets are exposed and thus no information on the angle of the aeolian planar cross-beds has been obtained. Due to a lack of lateral continuity, a lens-shaped geometry of the boundary of this facies is implied.

Fine-Grained Sandstone/Siltstone Facies

This facies occurs as thin sheet-like lensoid bodies intercalated within the protoquartzite/arkose facies of the Doornpoort Formation. It consists of interbedded quartz-rich siltstones and fine sandstones with subordinate mudstones. The rocks are light brown in colour and well sorted grain size distributions. The facies of fine-grained sandstone/siltstone rarely contains calcite-filled vugs, which are blade-shaped (Fig. 6 A) and sometimes show swallow-tail style re-entrant angles.

Symmetrical ripple marks (including micro-ripples, Chandler 1986) and cross-lamination are the common sedimentary structures together with plane lamination. The symmetrical ripple crests generally trend ENE. Polygonal desiccation cracks are occasionally preserved in fine grained layers (Fig. 3 B).

2.2 Palaeoenvironmental interpretation and synthesis of the alluvial fan to Braid Plain Depositional System

Rapid basinward facies changes characterize this portion of the sequence (Fig. 4). The interbedded conglomerate and arkose facies is the most proximal, which we interpret as lensoid alluvial fan bodies similar to those described by Ruxton & Clemmy (1986) elsewhere. Conglomeratic units represent debris flow deposits similar to those described by Bull (1972) from recent equivalent environments.

The interbedded arkoses represent sheetflood and braided stream deposits. At the base of the sequence the facies interfinger with both felsic and mafic volcanics. Higher in the sequence the alluvial fan deposits interfinger with the more distal protoquartzite/arkose facies. They represent braided stream sediments (Smith 1970; Miall 1978) and sheetflood deposits derived locally from the neighbouring fans and from axial westward flowing braided streams parallel to tectonic strike. A similar facies distribution in a graben-like pull-apart basin was inferred for the Midland Valley of Scotland (Bluck 1978, 1980) and in the Menderes graben of SW-Anatolia (Russel 1954). Locally the protoquartzite/arkose facies was reworked to produce the large scale cross-bedded quartz arenite facies interpreted here as aeolian dune deposits in a setting similar to that described by Hubert & Mertz (1984). The fine sandstone/siltstone facies was affected by unidirectional currents and wave processes, indicating deposition in localised ephemeral lakes or ponds. The predominance of debris flow on alluvial fans, development of aeolian dunes and the occurrence of evaporite minerals are evidence of a semi-arid to arid climatic regime in the basin although the source area may have been humid and montane..

2.3 The sub- to supratidal Depo System of the Upper Sinclair Sequence (i.e. Doornpoort and Klein Aub Formations)

Grey Quartz Arenite Facies

This facies is represented by laterally extensive sheet-like tabular thickbedded sandstone bodies 150 cm to 200 cm thick, with remarkably persistent thickness within any unit (Fig. 7 A). The bodies occur within units, together with other facies units, which are cyclically repeated up to seven times within the Klein Aub Formation (Figs. 2 and 7 A). The lithology is texturally and mineralogically mature and shows good sorting, but angular to subrounded grain shape. Small scale trough cross-bedding is the major sedimentary structure, in sets 5 cm - 10 cm thick, but much of the facies is apparently massive. Intergranular areas contain interspersed chlorite (5% - 10%), either disseminated or as a cement, and muscovite; however, silica and calcite predominate as cementing materials. Pyrite cubes up to 7 mm in diameter are disseminated throughout the rock.

Interlaminated and Flaser Laminated Siltsone and Mudstone Facies

This facies comprises predominantly coarse siltstone/fine sandstone with subordinate mudstone laminae. Sedimentary structures include cross-lamination, formed by both symmetric and asymmetric ripples, plane lamination and some planar cross-bedding. Where mud laminae occur with ripples, flasers are developed. The cross-laminae show numerous reactivation surfaces, scour surfaces and channel structures (Fig. 7 B and 7 C). This facies is also developed within each of the cycles immediately above the grey quartz arenite facies and shows laterally persistent thicknesses of 70 cm - 100 cm.

Interbedded Grey Mudstone and Siltstone Facies

In this facies type, grey mudstone predominates and is locally interlaminated with thin siltstone layers (Fig. 8 A). The thin-bedded to plane laminated and cross-laminated fine-grained sandstones are a minor component, locally exhibiting flaser lamination and normal grain size grading. Grains are well sorted but poorly rounded. Pyrite cubes 1mm across are also communly found disseminated within this facies. The beds described are laterally very persistent within outcrop limits. Other important features of this facies are large desiccation cracks. They are 10 cm to 40 cm deep, up to 5 cm wide and all of them taper conspicuously downwards (Fig. 8 A). In all cases the filling is a grey massive coarse grained to fine grained sandstone. Soft sediment deformation is absent from this facies. The facies is also repeated in each cycle and is on average 2 m thick.

Stromatolitic to Detrital Limestone Facies

Limestones occur as lenses and layers which are commonly associated with the grey mudstone and siltstone facies. Thin- to medium-bedded plane laminated or cross-laminated detrital limestone units occur in layers up to 40 cm thick extending metres to tens of metres along strike. Stromatolitic limestone units are fine-grained (recrystallized) and form discontinuous layers 1-2 cm thick, commonly lensoid over tens of centimetres (Fig. 8 A). They are internally laminated and may display "cellular" or "spongy" structure (Fig. 8 A and 8 B). The laminae sometimes form small, close, laterally-linked hemispheroidal domes (LLH of LOGAN et al. 1964). Elsewhere in the basin, in N. W. Botswana, such stromatolitic limestones are interbedded with detrital limestones. Here the stromatolitic limestones are thicker and more extensive and show discrete vertically stacked hemispheroidal dome features (SH of LOGAN et al. 1964) 20 cm high and 30 cm wide (Fig. 8 C). Overall, in Botswana there is a higher proportion of limestone in the stratigraphic equivalents of the Sinclair Sequence.

2.4 Palaeoenvironmental interpretation and synthesis of the sub- to supratital depositional system

The association of facies types is vertically arranged in at least seven fining and thinning upward asymmetric cycles (Fig. 2). This trend is caused by an overall increase in the mudstone component and a reduction in the thickness and frequency of coarser sediment influx. The desiccation cracks in the upper part of each cycle indicate repeated emergence of the sediment surface. The high maturity and grey-green colouration contrast with the underlying red-coloured alluvial sequence and leads to the interpretation of a shoreline palaeonvironment. The sedimentary structures indicate that the depositional setting was tidally dominated. Reactivation surfaces can develop in response to current reversal (KLEIN 1971) (although such structures are not restricted to tidal settings, e.g. COLLINSON 1970). Thin interlamination of silt and mud and development of flaser laminae on a fine scale are typically developed on tidal flats, due to variation in tidal velocity through successive tidal cycles (REINECK & WUNDERLICH 1968). Lack of grading distinguishes these silt layers from rhythmites.

Desiccation cracks together with wave and current produced ripples are also well developed on tidal flats (EVANS 1965, 1975). Tidal flat deposits are commonly organised in fining upward cycles resulting from a repeated progradation of the facies (EVANS 1965; TERWINDT 1975, 1988). A tidal flat environment is favoured over previous interpretations of a lacustrine environment for this part of the sequence (RUXTON 1980, 1986; RUXTON & CLEMMEY 1986; BORG & MAIDEN 1986a, b, 1989).

The three terrigenous facies types present in each cycle are remnants of subtidal, lower tidal flat and upper tidal flat subenvironments. The grey quartz arenite facies represents deposition in the subtidal area with higher energy currents causing the development of dunes (megaripples) (DE RAAF & BOERSMA 1971; TERWINDT 1971). The massive nature of the quartz arenite is explained, et least in part, by the good sorting of this sediment. The interlaminated and flaser laminated quartz arenite facies represents deposition in a lower to middle tidal flat setting. Erosional and channel features represent tidal channels crossing this subenvironment (REINECK 1984). Reactivation surfaces show the punctuated migration of bedforms (BOERSMA 1969; VISSER 1980), wave and current processes were both important in transporting sand. The interbedded grey mudstone and siltstone facies represents deposition on the middle to upper tidal flat, where deposition of fine sediment was at a maximum. Silt and mud interlaminae resulted from tidal cyclicity. Graded sand layers were produced by infrequent storm events moving fine sand into this subenvironment. Desiccation cracks are indicative of the repeated emergence of the tidal flat. The carbonate facies in the study area tends to developed in the upper tidal flat, where algal stromatolitic domes developed on a variety of scales. Detrital limestones represent the carbonate equivalent of the storm transported sands developed in the terrigenous facies.

It has been previously suggested that the thickness of the fining upward cycles in a tidally generated sequence can possibly be used as an index of the palaeotidal range (KLEIN 1971; TERWINDT 1988), although other authors have pointed out that this would depend on the degree of synsedimentary subsidence (BUTTON & VOS 1977; CALLOWAY & HOBDAY 1983) and therefore the index may act as a minimum measure. It is interesting to note that the cycles in the present study range from 4.5 to 5 metres. This would place the tidal range just above the mesotidal/macrotidal boundary, associated with an unbarred shoreline (HAYES & KANA 1979) as envisaged in this paper. Modern and ancient examples of such mixed siliciclastic-carbonate tidal environments of micro-, meso- and macro-tidal range have been described by various authors such as BEUKES (1977), BRUNN & MASON (1977), BUTTON & VOS (1977) and FLEMING (1977). BUTTON & VOS (1977) gave a good ancient analogue to the facies described and the (palaeo-) meso- to macrotidal setting interpreted in this paper. However, TERWINDT (1988) critically reviewed previous estimations of palaeotidal ranges thus casting some doubt on the appliccability of such classifications.

2.5 Possible evaporites within the sequence

No evaporite minerals are preserved in the sediments of the Sinclair Sequence. Nevertheless, blade-shaped vugs, filled with quartz, calcite and locally sulphide minerals, have been found (BORG & MAIDEN 1989). RUXTON (1981) described vugs, calcite nodules and lath-shaped calcite aggregates from the Witvlei aea, which he interpreted as pseudomorphs after gypsum.

Locally, fine-grained red quartzite of the fine sandstone/siltsone facies of the Doornpoort Formation contains calcite filled vugs (Fig. 6 A). The vugs are often blade-shaped and show rare swallow-tail style re-entrant angles.

The interbedded grey mudstone/siltsone facies of the Klein Aub Formation, contains aggregates consisting of calcite and Cu-sulphide (BORG & MAIDEN 1989). The aggregates are angular to lensoid (Fig. 6 B, C, D), but original shapes are partly obscured by cleavage and pressure shadows filled by quartz, calcite, and/or sulphide minerals. Rocks of the same facies type from the Chanzi Formation of the Lake N'Gami area (a Sinclair Sequence equivalent in N.W. Botswana) contain similar quartz-calcite-sulphide aggregates which also display conspicuously angular shapes (Fig. 9 A, B). The sedimentary layering of the host rocks commonly wraps around the aggregates (BORG & MAIDEN 1989), which often show an orientation normal to bedding (Fig. 9 B).

In agreement with RUXTON (1981) the more or less angular aggregates are interpreted as pseudomorphs after evaporite minerals such as gypsum or anhydrite. Evidence from the lake N'Gami area suggests that the evaporite minerals had formed within the soft sediment, prior to compaction (Fig. 9 B). The orientation of gypsum or anhydrite crystals, normal to the bedding planes, is commonly observed (VAI & RICCI LUCCI 1977). The occurrence of pseudomorphs after evaporite minerals supports both the interpretation of an arid playa environment for parts of the Doornpoort Formation and tidal flats in the Klein Aub Formation and its equivalents in the Lake N'Gami area.

2.6 Subsequent facies development

The upper contact of the facies just described is an erosional surface, on top of which lies a succession of up to 1500 m of purplish arkoses with subordinate 3 m - 4 m thick intercalations of pebble conglomerate, which SCHALK (1970) referred to as the Dikdoorn Beds (Fig. 2). We interpret this sequence as a return to continental alluvial conditions, as a braid plain environment. This heralds the deposition of the Damara Sequence whose basal units are similar in facies development to the Dikdoorn Beds, and which were deposited in essentially the same depositional basin.

2.7 Basin evolution

The facies described in previous sections document two distinct phases of basin development. The lower part of the sequence (Nückopf and Grauwater Formations) is dominated by bimodal volcanism, decreasing in intensity with time. Fault activity directly controlled sedimentation. Sedimentary facies transitions are rapid (on the metre scale) both laterally (over hundreds of metres) and vertically. The sediments are immature, derived from local basement and were deposited in subaerial, arid continental oxidising environments. Early in the basins history, sedimentation took place in individual sub-basins.

In contrast, volcanism is absent in most of the upper part of the sequence, i.e. Doornpoort and Klein Aub Formations. Facies types change rapidly in vertical sequence, but are laterally very persistent for tens of kilometres. These fine-grained, mature sediments were deposited in a variety of tidal marine sub-environments. Carbonate sediments are present and become an important component of this part of the sequence elsewhere in the basin. Sedimentation at this stage was not controlled directly by faulting and probably overlapped the initial tectonic margins.

We believe that these phases reflect different sedimentational responses to varying tectonic styles in the development of the basin. The lower part of the sequence represents a phase of active rifting dominated by block faulting and the development of sub-grabens. The upper part of the sequence results from thermal subsidence with erosion and gentle regional down warping of the basin flanks. Such phases in the history of intracratonic basins have been recognized by many authors as it applies to both modern and ancient basin development, e.g. modern North Sea basin (MCKENZIE 1978; ZIEGLER 1978, 1983; SCLATER & CHRISTIE 1980); and ancient examples (HOFFMANN et al. 1974). Fig. 10 summarizes stages in the development of the basin. Initially an extensional crustal regime is probably caused by an underlying mantle plume (BORG 1988a, b; BORG et al. 1987). This results in thermal uplift, followed by crustal thinning and block faulting to produce deep grabens (Fig. 10 A). Partial crustal melting resulted in the extrusion of mainly felsic pyroclastics and lavas. Varying proportions of tholeiitic intra-plate basalts were extruded in the vicinity of deeply penetrating faults (BORG & MAIDEN 1987). These extrusives were associated with contemporaneous intrusions (SCHALK 1970). Sedimentation was subordinate to volcanism, mainly comprising debris flow deposition.

The next stage in the study area was characterized by the cessation of igneous activity; sedimentation was initially still confined to the sub basins, but local horsts eventually buried, implying that the rate of sedimentation overtook the rate of fault movement. Individual alluvial fans are observable as conglomeratic lensoid bodies; one is visible directly in aerial photographs (Fig. 5) and three such bodies are identifiable as stratigraphic entities in the profile of Fig. 4. The latter shows that the lensoid bodies have a systematic arrangement, and are stacked vertically and laterally in an en échelon arrangement with the younger bodies stepping progressively eastwards. Although purely sedimentary reasons for this arrangement cannot be entirely ruled out, it seems more likely that tectonic causes were responsible. This arrangement would imply that the entry point was moving with respect to the depositional site. A similar relationship between the en échelon arrangement of alluvial fan bodies and a marginal strike-slip fault bas been documented by STEEL & GLOPPEN (1980) in the Devonian Hornelen Basin of western Norway. Since transport direction was from the north, this would imply that the bounding fault was a right-lateral oblique slip fault. This also might suggest that the rift development as a whole involved a strike-slip component.

In the final stage of basin development basaltic volcanism had ceased throughout the basin and block faulting had ceased as shown in the block diagram of Fig. 10 C. A marine transgression ensued which led to mature siliciclastic sedimentation under chemically reducing conditions and and to subordinate deposition of stromatolitic and detrital carbonates. This stage represents a sag or thermal subsidence phase which led to overstepping of the rift shoulders.

3 Conclusions

The Sinclair Sequence in the Klein Aub area was deposited in an extensional rift basin between the Congo and Kalahari Cratons. Thermal uplift, mechanical rifting and thermal subsidence phases can be distinguished on the basis of an evolution from volcanic dominated fill, to deposition of sediments in continental environental settings, to deposition of tidally dominated shallow marine sediments possibly under macrotidal conditions. A dextral strike-slip component to the rifting is indicated by the geometry and spatial distribution of successive alluvial fan bodies during the mechanical rift phase. Evaporitic pseudomorphs are recognised both from ephemeral lake deposits of the mechanical rifting phase and tidal flat deposits of the thermal subsidence phase. The association of debris-flow dominated fans and ephemeral lakes are indicative of semi-arid climatic conditions. The deposition of the Sinclair Sequence took place in a protobasin, which heralded the development of the Damara Trough. In fact, the uppermost facies types of the Sinclair Sequence are similar to those of the lowermost Damara Sequence in this region (PORADA 1985).

Acknowledgements

Funding of this research by the Geological Survey of SWA/Namibia and the Council for Scientific and Industrial Research (C.S.I.R.) is gratefully acknowledged. Geologists of the Geological Survey provided many useful discussions and overall logistic and scientific support. Dr. KARL SCHALK and CHARLIE HOFFMANN particularly gave of their detailed knowledge of the field area and always encouraged the progress of the research. The technical support of DI DU TOIT, ADELE WORKMAN-DAVIES, MELISSA KRAUSEY, MARK HUDSON, MITCH MILES, JOHN MOLOTSI and the late JETHRO MATLALA is gratefully acknowledged. Additional constructive comments by K. J. MAIDEN, H. HEINISCH, P. WYCISK, and A. CLAUSING on an earlier version of the manuscript were much appreciated.

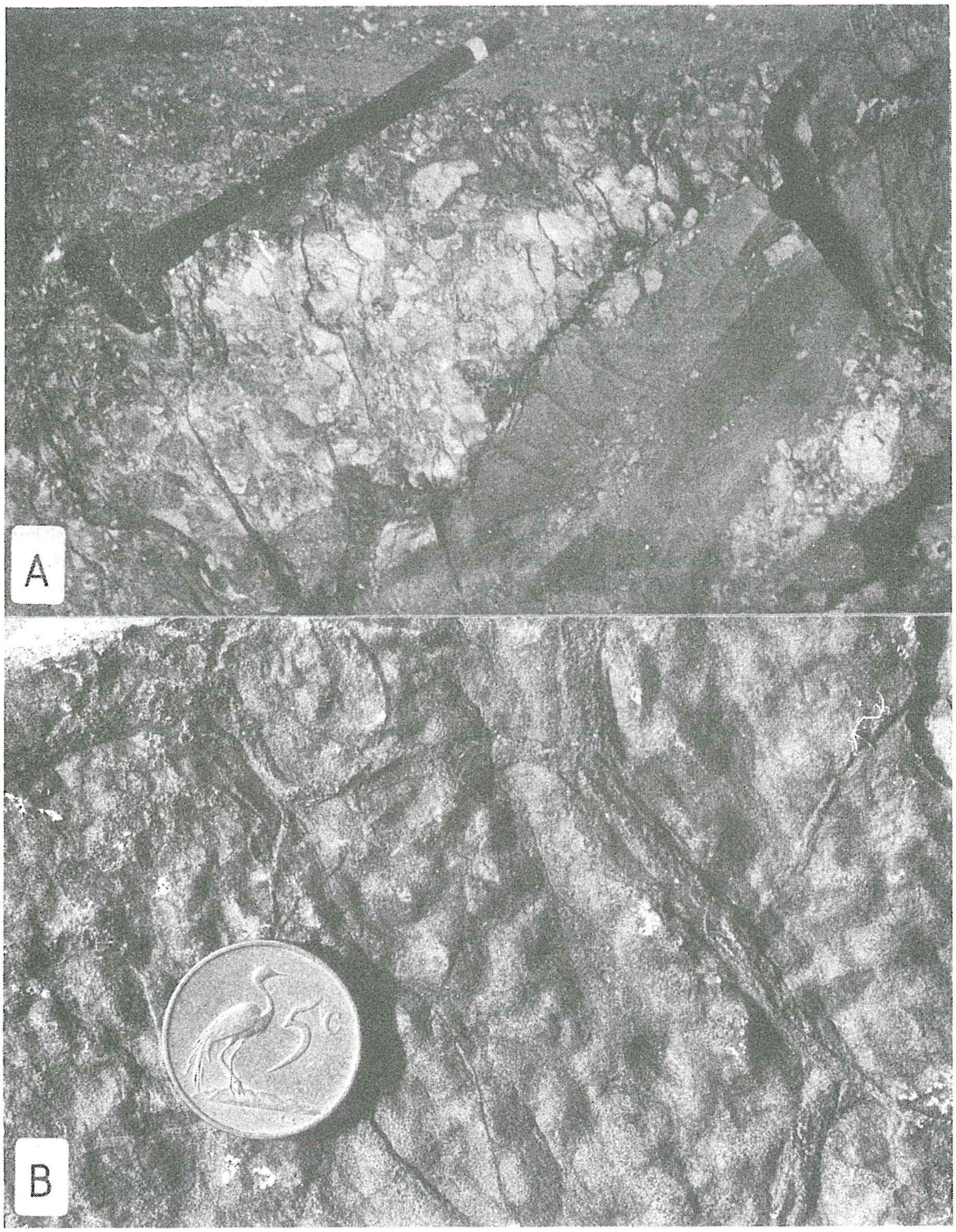


Fig. 3: Sedimentological features of the Doornpoort Formation.
A: Interbedded conglomerate and arkose facies. Note the massive nature of the conglomerate and the plane lamination of the arenite (hammer is 1m long).
B: Polygonal desiccation cracks within the fine sandstone/siltstone facies developed in playa lakes (diameter of coin is 17 mm).

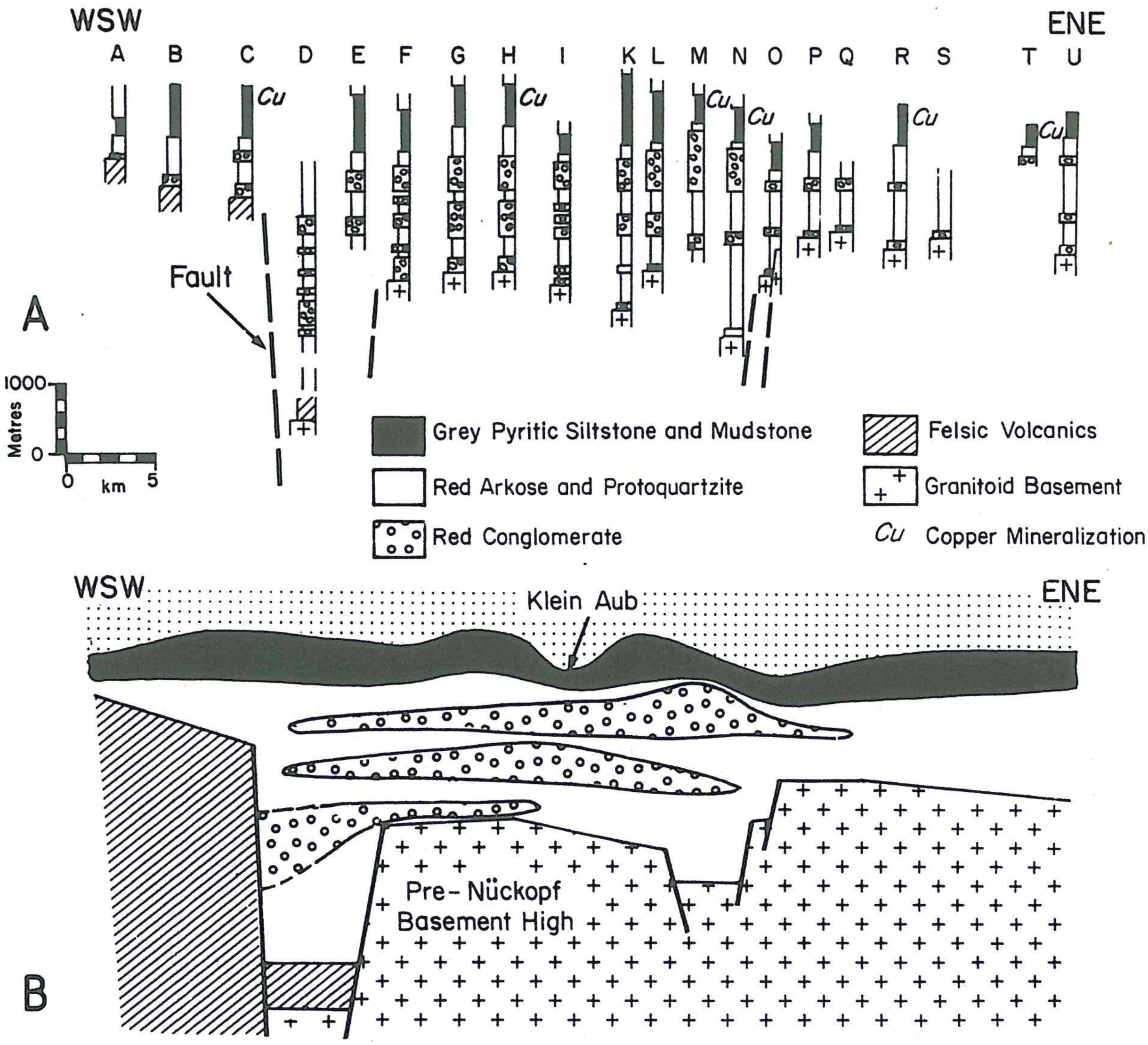


Fig. 4:
A: Measured sections A to U through the Doornpoort and Klein Aub Formations.
B: Correlation diagram showing distribution of lithotypes. NB: the en échelon arrangement of alluvial fan lenses.

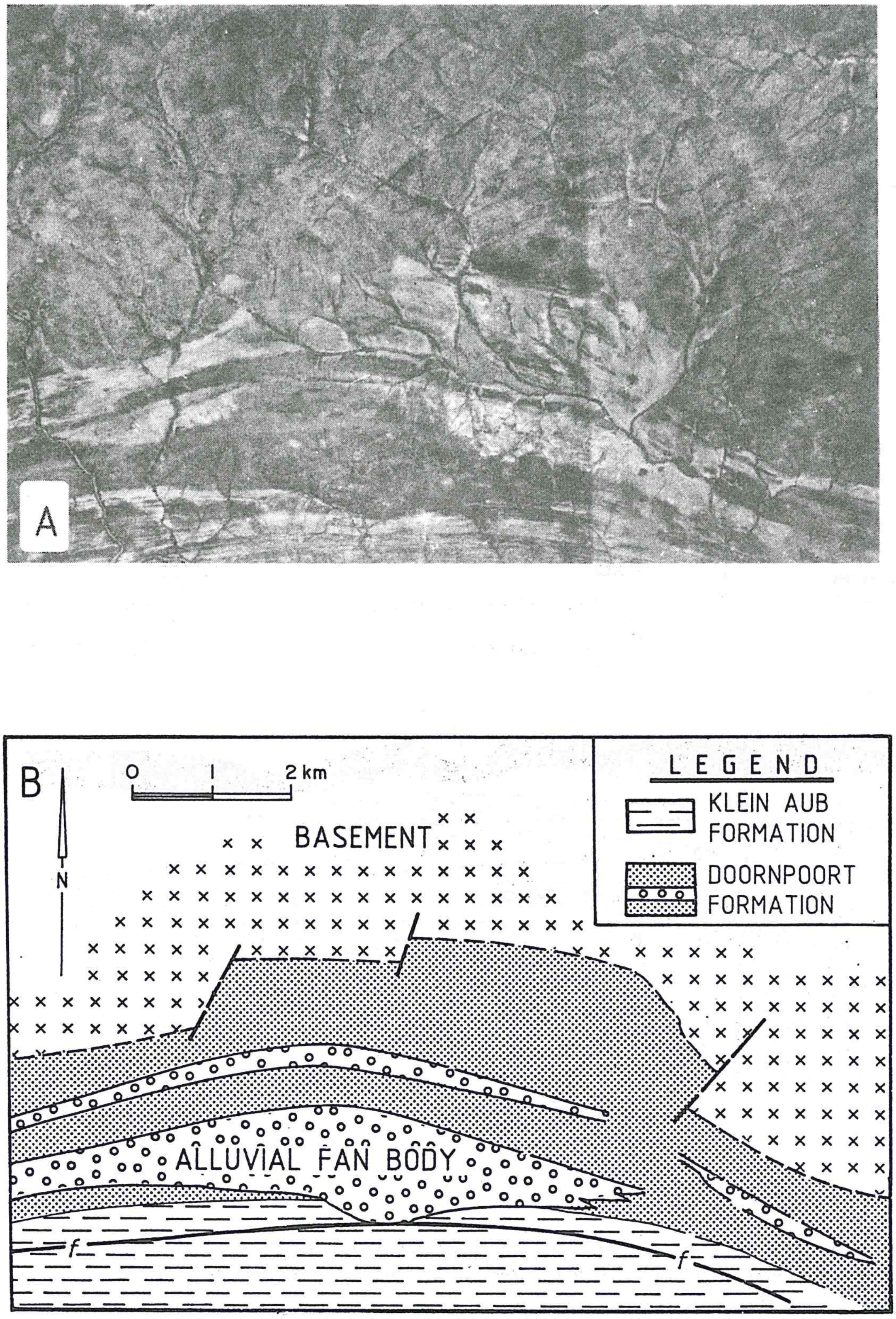


Fig. 5:
A: Aerial photograph.
B: Mapped interpretation of the same photograph displaying several discrete alluvial fan bodies. The sedimentary sequence dips 45° south and the present erosion surface has dissected an alluvial fan lobe. Note how fan position relates to a marginal feeder graben in the underlying basement. Also note the onlapping of the tidal Klein Aub Formation over the dormant fan and braided stream deposits.

Fig. 6:
(Page 14)

Pseudomorphs after evaporites, almost certainly gypsum, in the Klein Aub area.

A: Rhomb-shaped and lath-shaped vugs from the fine sandstone/siltstone facies of the Doornpoort Formation.

B, C, D: Rhomb-shaped pseudomorphs after gypsum, developed in muddy siltstone of the interbedded grey mudstone and siltstone facies of the Klein Aub Formation. The pseudomorphs consist of quartz, copper sulphides, chlorite and calcite. All shapes are obscured by pressure shadow development (indicated by arrows).

B: Well preserved pseudomorph under crosed polars.

C: Deformed equivalent also under crossed polars.

D: Highly deformed pseudomorph under plane polarized light.

Fig. 7:
(Page 15)

A: Fining-upward cycles of the Kagas Beds in the Klein Aub Formation in a river section 4 km ENE of Klein Aub Mine. Base and top of one complete cycle are arrowed. One of the authors is sitting on the grey quartz arenite facies which represents deposits from subtidal to lower tidal flat environment. Above this are facies representing lower to upper tidal flat deposits.

B: Shallow sand-filled channel (Basal scour (I) within the interlaminated and flaser laminated facies (same locality as Fig. 7 A). Laminae may represent small-scale epsilon cross-stratification filling this channel, with reactivation surfaces developed (II).

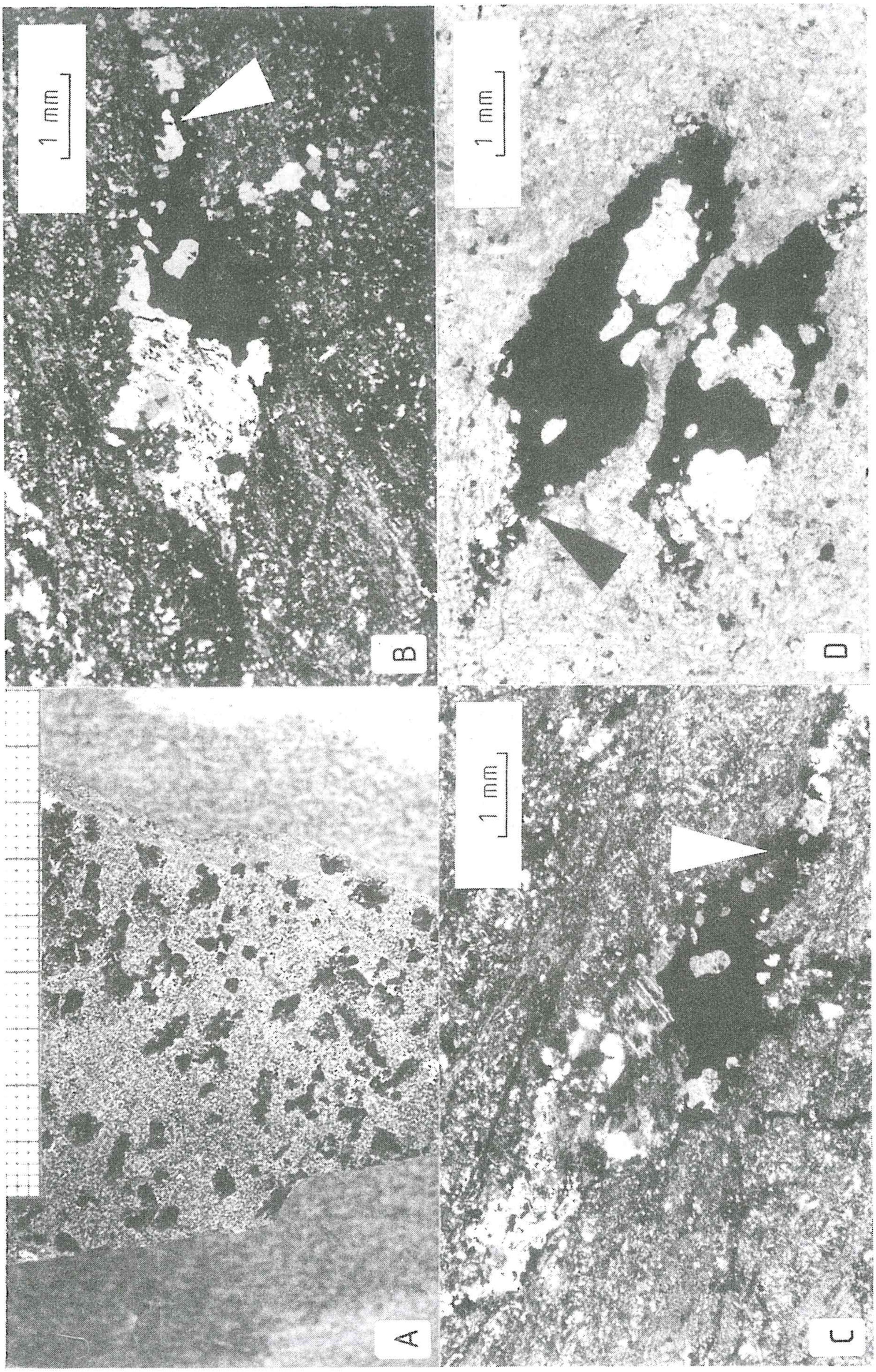
C: A progression from interlaminated subfacies (I) to flaser laminated subfacies (II) developed on the mid tidal flat (Same locality as Fig. 7 A). The more easily eroded flaser laminated subfacies is eroded to form a mid tidal flat channel (III) which is sand-filled. This area subsequently became a zone of active deposition of sand beds (IV).

Fig. 8:
(Page 16)

A: Sand-filled large desiccation crack (upper arrow) developed in the interlaminated facies (same facies as Fig. 7). A thin lensoid limestone is also indicated (lower arrow). It dislays a cellular structure and is interpreted as intertidal algal mat (coin is 17 mm in diameter).

B: Detail of algal laminated facies intercalated with the mid-tidal flat deposits (Same locality as Fig. 7). Plane laminated limestone passes upwards into poorly developed stromatolitic domes.

C: Well developed domal stromatolite (SH of Logan et al. 1964) interbedded with plane laminated stromatolites and detrital limestone. Locality: Lake N`Gami area of Botswana. Pencil is 14 cm long.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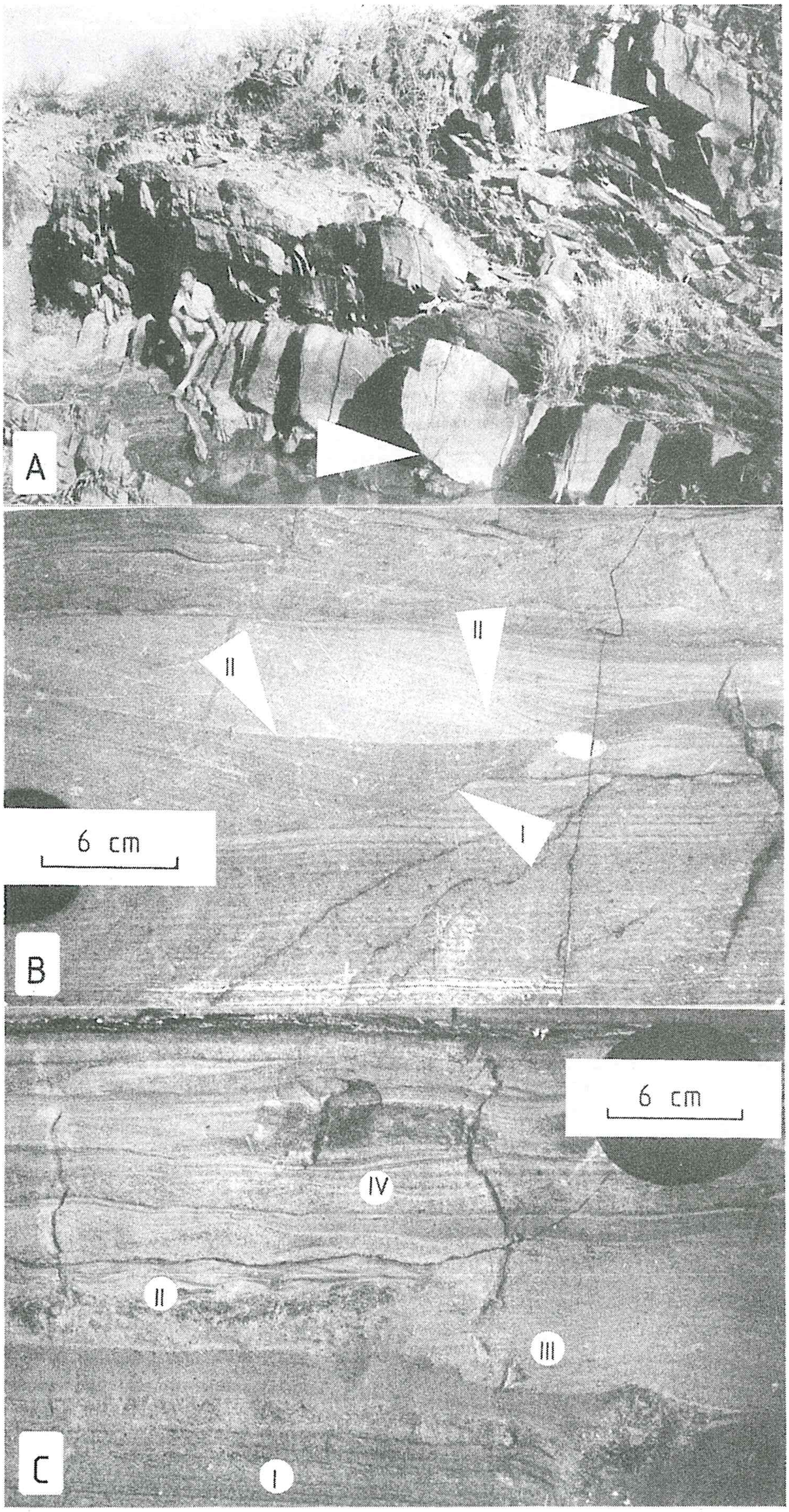
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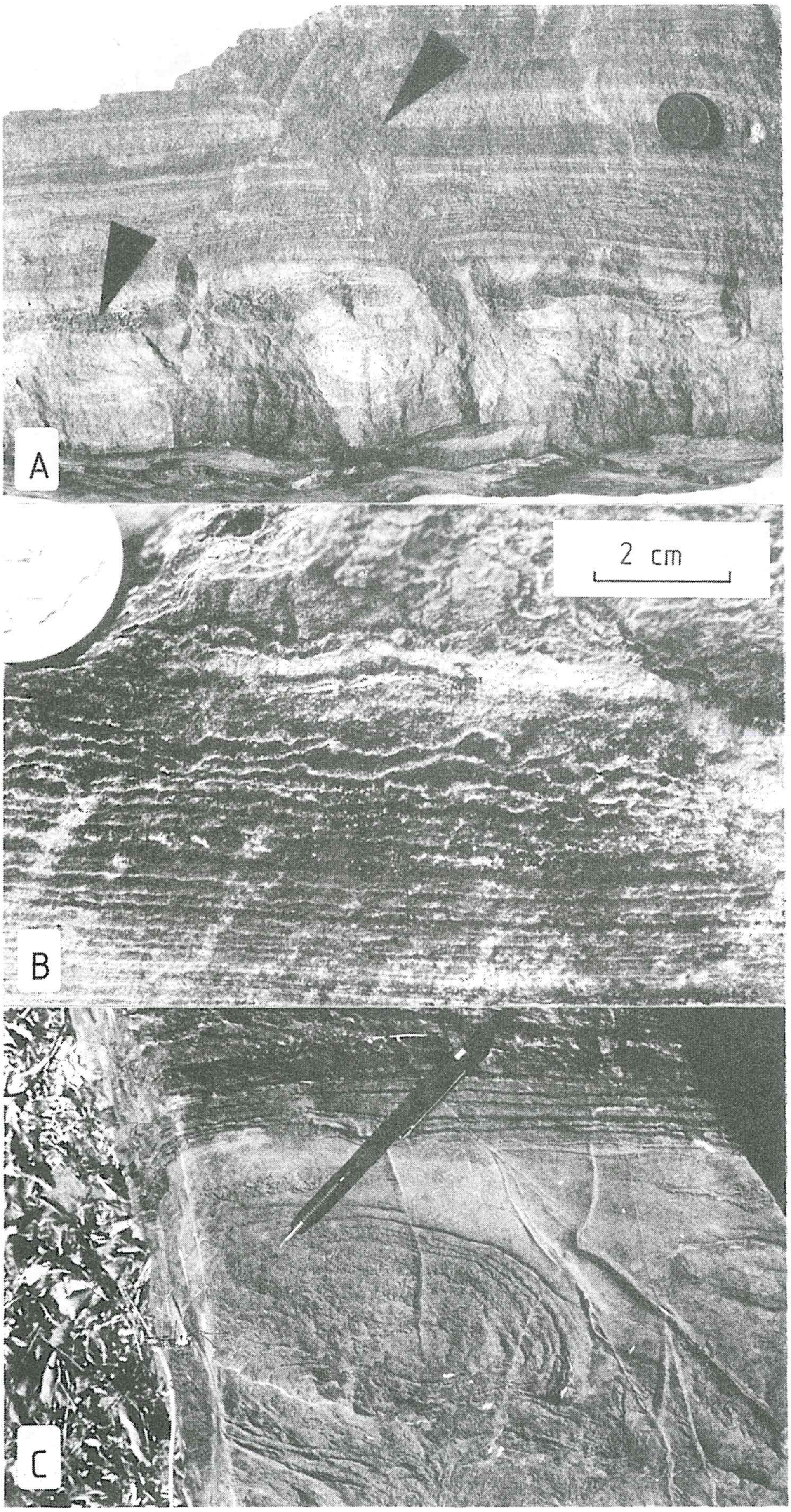
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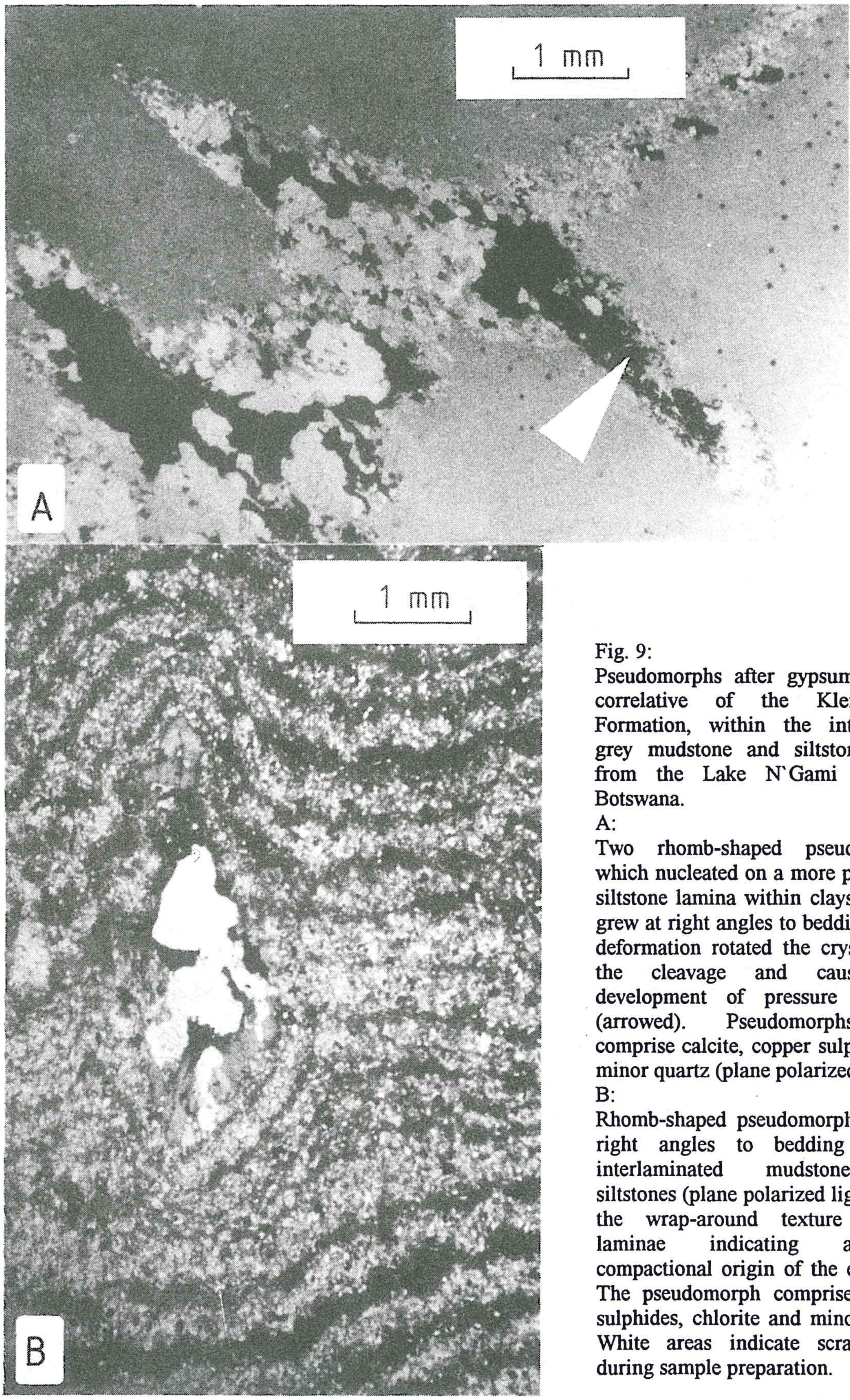

Fig. 9:
Pseudomorphs after gypsum from a correlative of the Klein Aub Formation, within the interbedded grey mudstone and siltstone facies from the Lake N`Gami area of Botswana.
A:
Two rhomb-shaped pseudomorphs which nucleated on a more permeable siltstone lamina within claystone and grew at right angles to bedding. Later deformation rotated the crystals into the cleavage and caused the development of pressure shadows (arrowed). Pseudomorphs now comprise calcite, copper sulphide and minor quartz (plane polarized light).
B:
Rhomb-shaped pseudomorph, also at right angles to bedding in the interlaminated mudstones and siltstones (plane polarized light). Note the wrap-around texture of the laminae indicating a pre-compactional origin of the evaporite. The pseudomorph comprises copper sulphides, chlorite and minor quartz. White areas indicate scraping off during sample preparation.

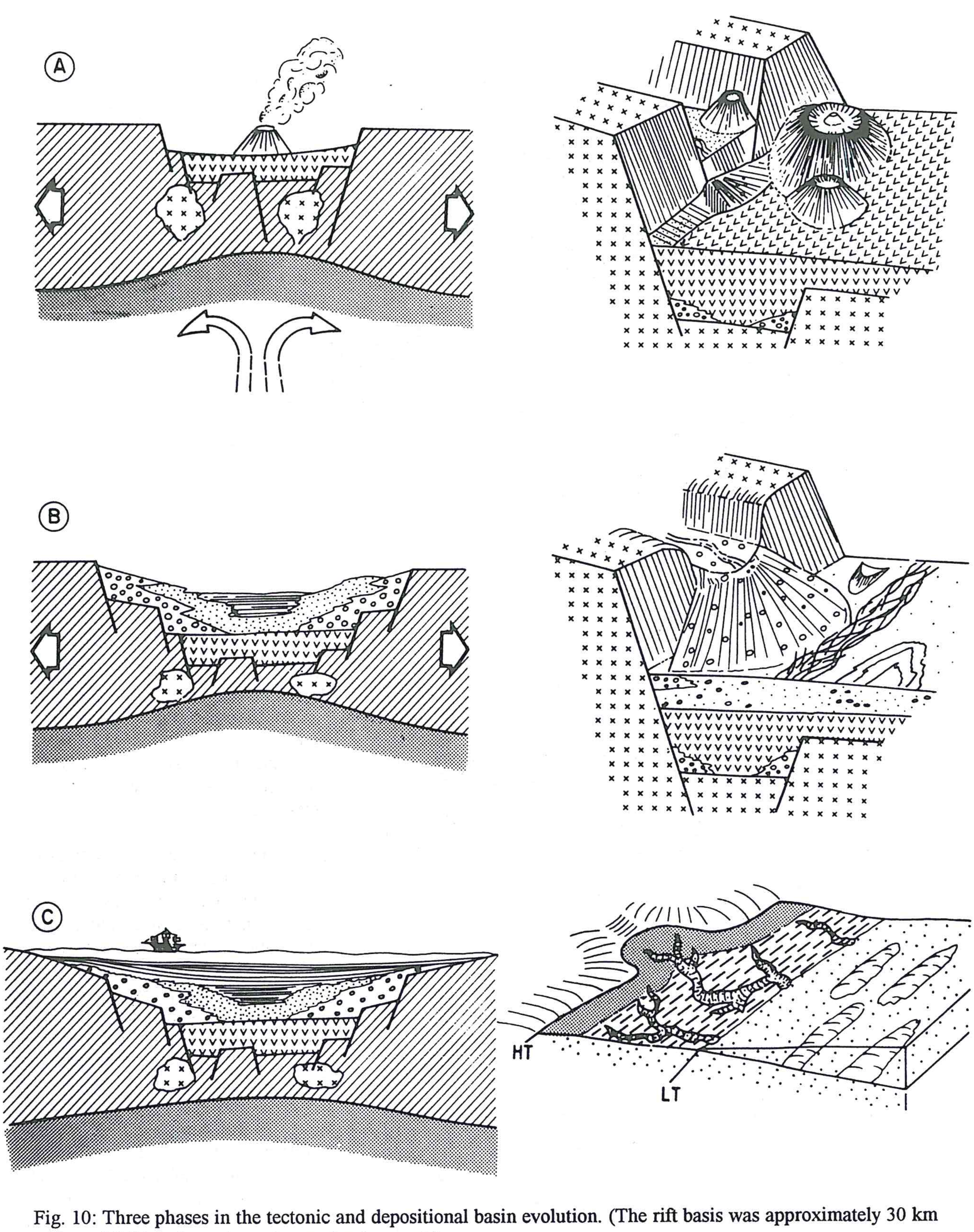

Fig. 10: Three phases in the tectonic and depositional basin evolution. (The rift basis was approximately 30 km wide and 8.5 km deep at maximum development.)

A: Crustal extension, related to an underlying mantle plume, causes major volcanism and minor sedimentation related to block faulting.

B: Block fault movement continues and sedimentation of continental red beds predominates over volcanism. Alluvial fan and braided stream deposits are reworked by aeolian processes and grade into playa lake deposits.

C: Volcanism and block faulting cease and the basin sags and widens due to thermal subsidence. The resulting marine transgression led to the development of tidal flats associated with an unbarred macrotidal shoreline. (HT = high tide position; LT = low tide position).

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