BASIN ANALYSIS AND SEQUENCE STRATIGRAPHY : A REVIEW, WITH A SHORT ACCOUNT OF ITS APPLICABILITY AND UTILITY FOR THE EXPLORATION OF AURIFEROUS PLACERS IN THE WITWATERSRAND BASIN

By

Johan van Eeden

Dissertation submitted in the partial fulfilment of the requirements for the degree of

MASTER OF SCIENCE

in

EXPLORATION GEOLOGY

in the

FACULTY OF SCIENCE

RHODES UNIVERSITY

Supervisor : Prof John Moore

January 1996

This dissertation was prepared in accordance with specifications laid down by the University and was completed within a period of ten weeks full - time study.

ABSTRACT

The Witwatersrand basin is unique in terms of its mineral wealth. The gold in the Witwatersrand basin is mainly concentrated in the placers and two types of unconformities are associated with the placer formation.

This paper attempts to quantitatively describe the origin and depositional process of placers within the context of basin analysis, geohistory and sequences stratigraphic framework.

Several tectonic models have been proposed for the evolution of the Witwatersrand basin and it seems as if a cratonic foreland basin accounts for many of the observed features observed the Central Rand Group basin. The tectonic subsidence curve generated for the Witwatersrand Basin clearly implies foreland basin response which was superimposed an older, deep seated extensional basin. These compressive tectonics can be superimposed on extensional basins, where the shift from extensional to compressional tectonics lead to inversion processes.

The critical issues about the Witwatersrand basin which were addresed in this review, is the validity of basin wide correlation of placer unconformuties and whether sequence stratigraphy is applicable to fluvial systems of the Witwatersrand sequence. It is believed that the Central Rand Group was deposited as alluvial - fan deltas by fluvially dominated, braidplain systems with minor marine interaction which had a considerable impact on the preservation of economically viable placers.

Most important to the exploration geologist is the recognition of stacking patterns of the fluvial strata to determine change in the rate at which accommodation was created. Identifying sequence boundaries and other relevant surfaces important for identifying these stacking patterns of the sequences, depends entirely on the recognition of a hierarchy of stratal units including beds, bedsets, parasequences, parasequence sets and the surfaces bounding sequences.

Placers are closely associated with the development of disconformities and therefore become important to recognise in fluvial strata. If these placers are to become economic, the duration of subaerial exposure of the unconformities that allowed the placers to become reworked and concentrated must be determined. In order to preserve the placer, a sudden marine transgression is necessary to allow for minimal shoreline reworking and to cap the placer to prevent it from being dispersed.

The placers in the Witwatersrand basin occur in four major gold - bearing placer zones in the Central Rand Group. Accordingly they can be assigned to four supercycles, which are cyclical and therefore predictive. It is the predictive nature of these rocks and the ability of sequence stratigraphy to enhance this aspect, which is a pre - requisite for an effective exploration tool in the search for new ore bodies or their extension in the Witwatersrand basin.

TABLE OF CONTENTS

LIST OF FIGURES:

LIST OF TABLES:

1 INTRODUCTION

The Witwatersrand gold - uranium placer deposits in South Africa have produced over 50 % of all the world's gold (Tainton, 1994). Other fluvial placers fields elsewhere in the world, such as Tarkwa (Ghana), Moeda and Jacobina (Brazil), and the Huronian (Ontario), have also been economically important. Therefore, fluvial placer deposits are clearly an important gold deposit type in global terms (Force, 1991b).

The Witwatersrand basin however, remains unique in terms of its mineral wealth. This study concentrates on three broad areas of basin analysis and sedimentary geology, to ascertain the controlling factors determining uniqueness of the Witwatersrand basin i.e. primary mechanics of basin formation, controls on basin sedimentology and stratigraphy, and tectonic subsidence and geohistory analysis (Klein, 1991a). Of particular interest is the predominantly fluvial nature of the Central Rand Group sediments, which contain most of the auriferous placers, and its geohistory. Special reference will be made in this study to the application of sequence stratigraphic concepts and their predictive capabilities in the exploration of further deposits.

1.1 The Role of Basin Analysis in Exploration

Basin analysis should form the basis of any exploration for placer deposits. In the current era, basin analysis integrates several fields of earth science, including geophysics, geochemistry, sedimentology, stratigraphy, precise biostratigraphy for geological age determination, structural geology and computer modelling (Klein, 1991b). Furthermore, it uses the systematic approach of both the quantitative and qualitative understanding of a sedimentary basin's response to regional and local geological variables, such as climate, water chemistry, biological activity, tectonism and volcanism (Haq, 1991). The ultimate role of interdisciplinary basin analysis is to ensure that the resource potential of a basin is fully understood and to define exploration targets (Eidel, 1991).

Sequence stratigraphy has become an important analytical tool for sedimentary processes and petroleum exploration. The exploration efforts of the petroleum industry, in particular the Exxon research group (e.g. Vail et al, 1977; Haq et al., 1987; Posamentier and Vail, 1988), revolutionised the manner of viewing stratigraphic relationships. The conceptual background to sequence stratigraphy is published in a series of key papers in SEPM Memoir 42 (Wilgus et al, 1988) and in Van Wagoner et al, (1990) and Haq (1991). Since the mid - 60`s sequence stratigraphy has been applied extensively to hydrocarbon exploration in Phanerozoic rocks, but it was only after the publication of the AAPG Memoir 26 in the mid - `70`s that the Vail et al. (1977) global sea - level model made such an impression on the methodology of sedimentary basin interpretation.

However, sequence stratigraphy has yet to be extensively applied to mineral exploration, and in particular to the Witwatersrand basin. The application of sequence stratigraphy is essential for the understanding of the origin and depositional processes of placers in the Witwatersrand and ultimately to provide some predictive guidelines for their exploration. The appeal to the explorationist is to appreciate the predictive nature of the sequence stratigraphic concept models. The strength of sequence stratigraphy lies in its ability to implicitly enhance the cyclic nature of stratigraphic successions and the use of the chronostratigraphic framework to enhance lithologic prediction (Posamentier and James, 1993).

Since the advent of high - quality seismic data, sequence stratigraphy has been applied to regional seismic data, high resolution seismic data, outcrop and subsurface geology, flume scale data and to modern systems. The emphasis in sequence stratigraphy has shifted from an age - model prediction to a lithologic prediction since the publications of Jervey (1988), Posamentier et al. (1988), Posamentier and Vail (1988), Sarg (1988) and Van Wagoner (1990).

The concepts of sequence stratigraphy are sometimes misunderstood and applied as a rigid template. If the fundamental principles of sequence stratigraphy are correctly applied, the user can appropriately utilise the information to build a suitable model for the Witwatersrand basin sedimentary processes. As with any model, caution must be taken not to idealise or over generalise these concepts, but to consider sequence stratigraphy as an additional way of looking at and ordering geologic data. According to Posamentier et al. (1988) sequence stratigraphic concepts should be applied as an approach or tool, rather than as a template. It is important to remember that sequence stratigraphy deals with the stratigraphic response to the interaction of sedimentary influx vs the space created on the shelf for the sediments to fill. These two parameters are essentially space and time independent (Posamentier and James, 1993).

There are, however, two issues regarding sequence stratigraphy which still remain contentious. One is the issue of global synchroneity of unconformities, in spite of all the recent publications with more comprehensive information based on sequence stratigraphy and sea level changes. In terms of the Witwatersrand basin the issue of correlating unconformities laterally within the basin is important and even correlating sequence boundaries with other basins on a global scale. The other issue is the question of whether sequence stratigraphy can be applied effectively to fluvial systems. It is felt that the effect of sea - level changes on fluvial systems may be swamped by climatic and autocyclic changes such as local tectonics, sedimentary flux variations and changes in fluvial discharge upstream (Posamentier and James, 1993). This issue is very relevant to the Witwatersrand sediments, in particular to the economically important fluvially dominated sequences within the Central Rand Group.

1.2 Previous Work in the Witwatersrand Basin

The huge gold deposits of the Witwatersrand basin have become probably the most studied mineral deposits since the day when gold was discovered on the farm Langlaagte near Johannesburg in 1886. Since then many papers have been published on the clastic sediments of the Witwatersrand Supergroup. Key papers include Mellor (1913, 1915, 1916, 1917), Brock and Pretorius (1964) etc. A comprehensive review of earlier work was presented by Pretorius (1975). Detail geological accounts of the history of each major goldfield within the Witwatersrand basin were given by Antrobus, (1986).

However, only very few studies have been undertaken to unravel the geohistory of the Witwatersrand basin. It is only recently that subsidence analysis has been applied to the Witwatersrand basin by Beukes et al. (1995), Nelson et al. (1995) and Maynard and

Klein (1995). The application of sequence stratigraphy with its ability to construct age models for a given stratigraphic succession and to predict lithology based on the interpretation of identifying cyclicity in the rock record, is an obvious and useful technique. However, thus far it has been applied mainly to Phanerozoic basins and only recently to Archaean and Early Proterozoic basins (eg. Grotzinger, 1986; Christie Blick et al., 1988).

Beukes and Cairncross (1991) have applied sequence stratigraphy to correlate the late Archaean age Mozaan Group of the Pongola Supergroup with the Witwatersrand Supergroup. Krapez (1993) has applied similar techniques to reassess the stratigraphic successions of Archaean supracrustal belts of the Pilbara Block.

The aforementioned regional sedimentological and stratigraphic studies in Southern Africa, Australia, and North America show unequivocally that sequence stratigraphic concepts are applicable to older basins despite the difficulties in calibrating and detecting hiatuses.

1.3 A Geological Overview of the Witwatersrand Supergroup

The known extent of the Witwatersrand basin measures approximately 300 x 150 km. The basin has a NE - SW elongated shape and is made up of several individual goldfields, including the Evander, East Rand, Central Rand, West Rand, Carletonville, Klerksdorp and Welkom goldfields. By definition a goldfield comprises several gold producing mines, each perhaps belonging to different companies. The goldfields do not form one continuous line, but have several gaps in between, either due to structural complexity or declining gold grades i.e. Potch and Bothaville "gaps" (Figure 1.1).

1.3.1 Dominion Group

Thick volcanics both underlie and overlie the clastic sediments of the Witwatersrand Supergroup ie. Dominion Group and Klipriviersberg Subgroup lavas. The Dominion Group represents a proto - basinal phase of the Witwatersrand consisting of 2710 m of lavas and sediments which rest unconformably on Archaean basement granites

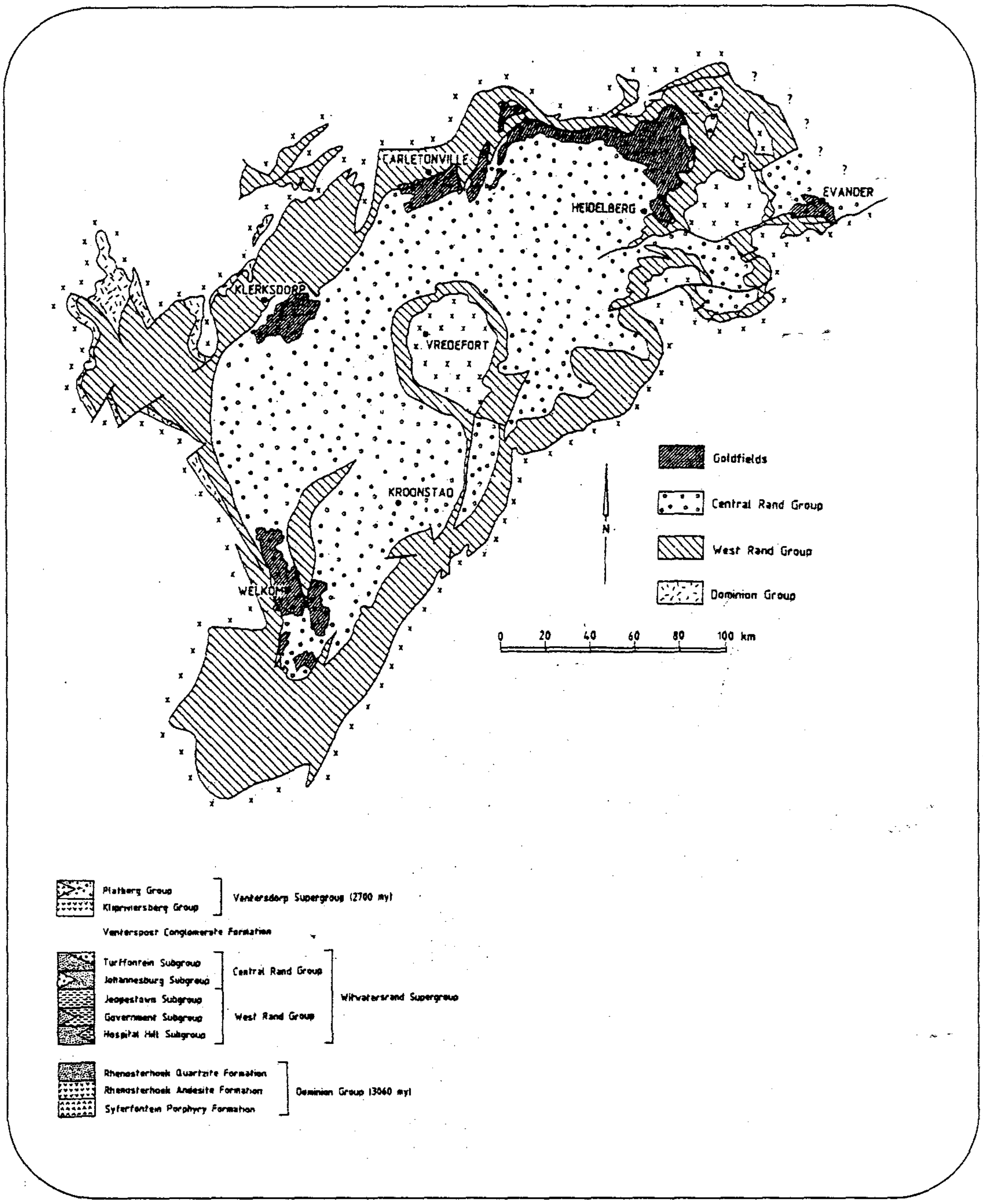


Figure 1.1 - Sub - Transvaal Supergroup geology in the area of the Witwatersrand Basin showing the distribution of Witwatersrand Supergroup and certain middle Ventersdorp basins. Below is a generalized stratigraphic column from the Dominion Group to the Transvaal Supergroup (From Myers et al., 1990).

The Dominion Group can be subdivided into the Renosterspruit Formation at the base consisting of up to 60 m sandstone, minor conglomerates and argillaceous horizons overlain by interbedded volcanosedimentary sequences (Figure 1.1). Conformably overlying this unit is the Renosterhoek Formation comprising 1100 m basaltic andesites and tuffs. The Syferfontein Formation represents the upper most unit comprising acid lavas, subordinate tuff layers, andesitic lava, volcanics breccias and quartz - feldspar porphyries (Jackson, 1992).

1.3.2 Witwatersrand Supergroup

The Witwatersrand Supergroup itself contains minor lavas ie Crown (West Rand Group) and Bird lavas (Central Rand Group) (Figure 1.2). The Witwatersrand basin is an elongated structure filled predominantly with clastic sediments of West Rand and the overlying Central Rand Groups (SACS, 1980).

The West Rand Group represents the lower portion of the Witwatersrand Supergroup and has an average thickness of 4650 m (Figure 1.2). It comprises a sequence of predominantly marine shelf shales and shallow marine shelf orthoquartzites interbedded with fluvial braid plain quartzites and conglomerates, with a major unconformity separating it from the underlying Dominion Group (Tankard et al., 1982) (Figure 1.2). The West Rand Group varies in thickness and is laterally extensive throughout the basin. There is an apparent for the West Rand Group to thin towards the Evander and the South Rand goldfields in the southeast. The West Rand Group attains a thickness of up to 850 m in the Evander area, but faulting renders these average thicknesses questionable. Economically minor gold - bearing conglomerates occur in the Jeppestown and Government Subgroup within the West Rand Group (Beukes and Nelson, 1995; Watchorn and O` Brien, 1991). Some of these placers include the historically famous Bonanza, Promise, Coronation, Government, Buffelsdoorn and Veldschoen reef horizons, which were mined on small scale at the turn of the century.

The upper part of the Witwatersrand Supergroup is represented by the Central Rand Group which has a maximum cumulative thickness of 2880 m in the Central Rand goldfield (Figure 1.2). The Central Rand Group is essentially more arenaceous,

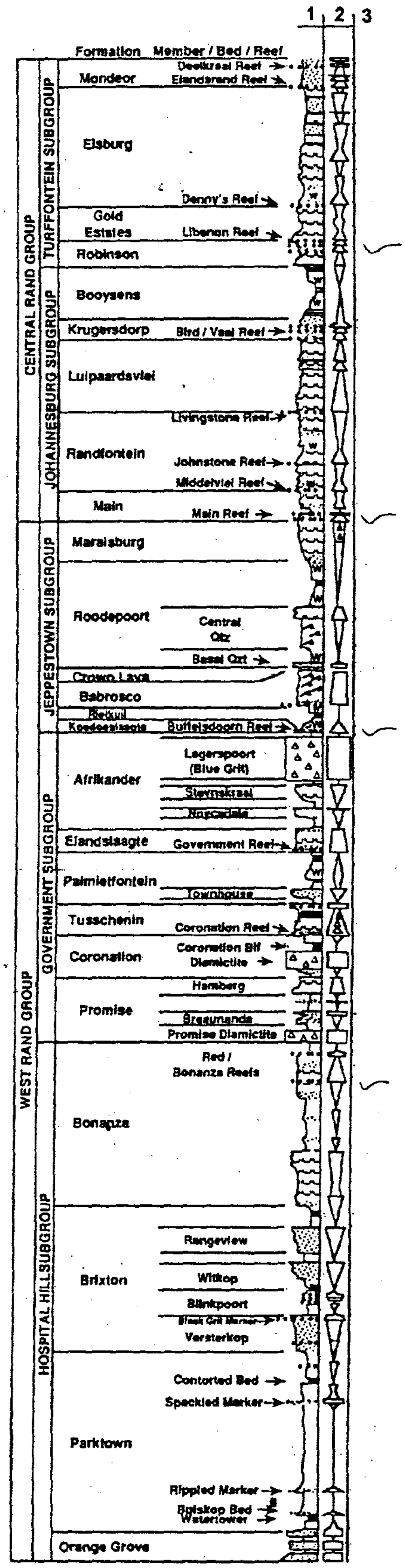


Figure 1.2 - General stratigraphic reference profile for the Witwatersrand Supergroup in the Carletonville - Krugersdorp area. Column 1: Lithology - (w = wackstone, clear = argillite, black dots = conglomerate, troughs = argillaceous quartzite, triangles = diamictites). Column 2: Genetic succession. Column 3: Incised valleys. (Modified from Beukes and Nelson, 1995).

comprising fluvial quartzites and conglomerates, alternating with shallow marine inner shelf orthoquartzites and minor middle to outer shelf quartz wackestones, siltstones and mudstones (Tankard et al., 1982) (Figure 1.2). It is believed that the Central Rand Group was deposited mainly as alluvial - fan-deltas by fluvially dominated, braid plain systems with minor marine interaction of considerable economic significance (Beukes, 1990; Karpeta et al., 1991; Karpeta, 1994; Beukes and Nelson, 1995). The economically significant auriferous placers within the Witwatersrand Supergroup, are confined to the Central Rand Group. The placers apparently formed in a foreland basin setting.

The thickness of the Central Rand Group strata increases towards the basin centre, as the foreland depositional axis migrates with renewed sedimentation, resulting in the thickest strata along the depositional axis. Both the Central Rand and West Rand Groups individual and cumulative thicknesses decrease towards the southeastern margin. The Pongola Supergroup is now considered to be part of the Witwatersrand basin and originated as an entity in the same depository (Beukes and Cairncross, 1991). The southwestern and northeastern edges of the Witwatersrand basin still remain ill defined due to thick sequences of younger cover rocks and structural complexity. The dominant palaeocurrent directions are southeasterly but southwesterly components have been recorded west of the Klerksdorp goldfield. Palaeocurrent directions into the basin from a southwest source area have been recorded in the Evander goldfield (Minter, 1991) (Figure 1.3). Another important feature regarding the basin is the West Rand sea with a relatively straight northeast to southwest shoreline and a general southeasterly palaeodeclivity (Mayer and Albat, 1988).

1.3.3 Ventersdorp Supergroup

The Ventersdorp Supergroup occupies a large elliptical basin which exceeds 200 000 km^2 (Figure 1.4). In most case it overlies the older sequences with a distinctive angular unconformity developed at the base of the Klipriviersberg Group lavas (Figure 1.1). The Klipriviersberg Group attains a thickness of 1830 m, consisting of voluminous continental tholeiitic basalts, which filled half graben structures controlled by post - Witwatersrand extensional rift faulting.

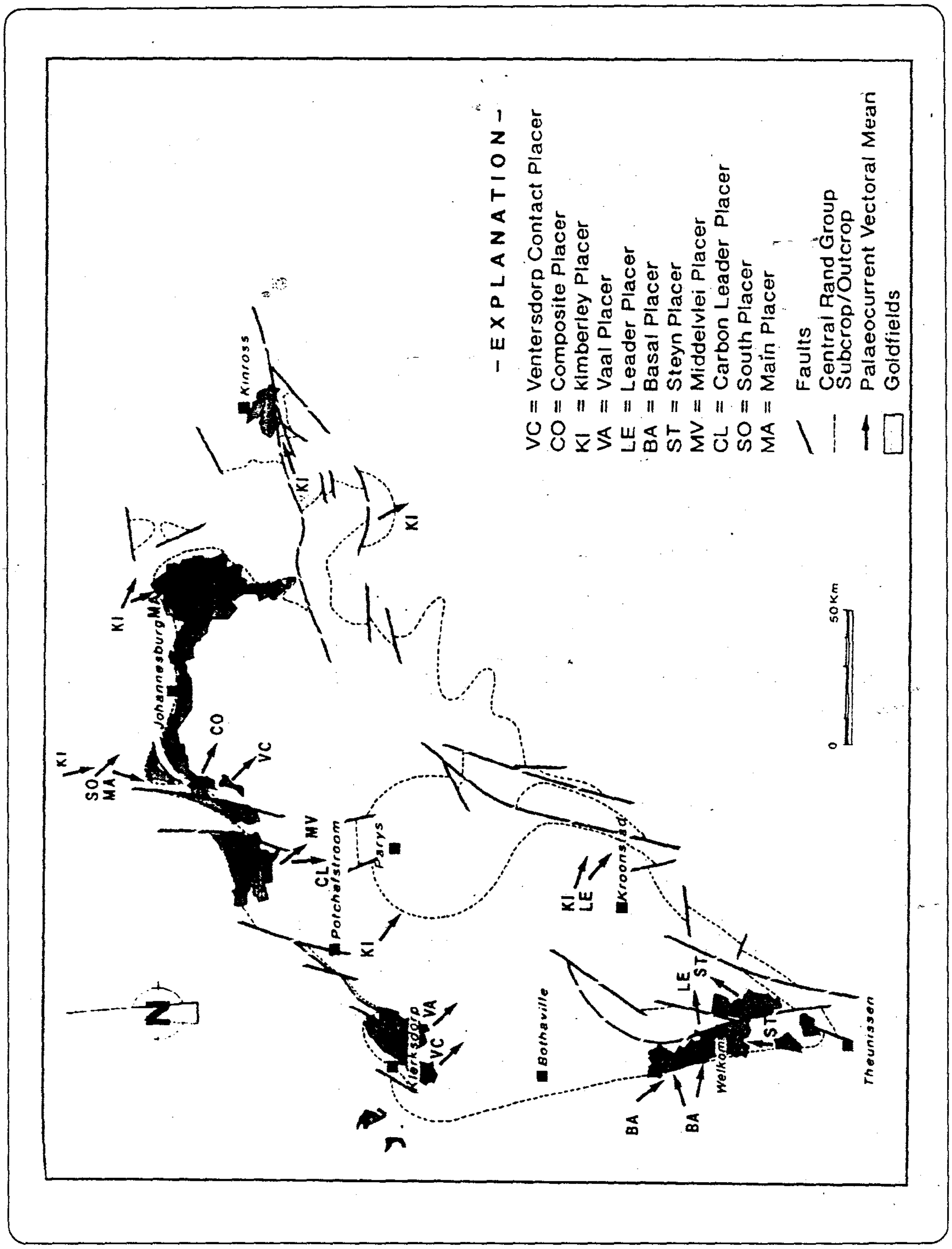


Figure 1.3 - Simplified map showing the preserved outline of the Central Rand Group and the localities of the various goldfields along the western and northwestern periphery. Palaeodispersal of major placers deposits are added, modified after Minter and Loen (1991) (From Tainton, 1994).

The Witwatersrand basin was deposited on older basement rocks of the Kaapvaal craton (Hartnady and Stowe, 1991). The age of the Witwatersrand basin has been constrained by dating the underlying Dominion Group lavas (Armstrong et al., 1990) at ~3074 ± 6 Ma and the upper limit at ~2714 ± 8 Ma for the mafic sequence of the overlying Klipriviersberg Group (Figure 1.4). Thus a period of 360 Ma (~3075 Ma to ~2700 Ma) is implied. The Crown lava of the Crown formation in the Jeppestown Subgroup has been dated at ~2914 ± 8 Ma. (Armstrong et al., 1990). Beukes and Nelson (1995) bracket the age of the Witwatersrand and Mozaan sequences between ~2970 Ma and ~2820 Ma, allowing 140 - 150 Ma for sedimentation.

1.3.4 Tectonic setting of the basin

Several tectonic models have been proposed for the formation of the Witwatersrand basin. The pioneering work of Borchers (1964) led to a map showing the surface and subsurface geology of the basin and became a benchmark contribution to the definition of the shape, configuration, boundaries and geometry of the basin. The precedent set by Borchers (1964) influenced all subsequent models. This map was modified twenty five years later by Pretorius, Brink and Fouché (1986).

In the mid -'70's Vos (1975) and Hutchinson (1975) favoured the idea of an intracratonic, alluvial plain, lacustrine model for the Witwatersrand basin. Van Biljon (1980) was the first to propose a plate tectonic model for the Witwatersrand basin involving continent - continent collision. Pretorius (1981) proposed a taphrogenic basin model, based on a tectonic framework controlled by vertical tectonics, giving rise to syneclises and anteclises in a pattern of superimposed interference folds.

Bickle and Eriksson, (1982) and Clendenin et al. (1988) proposed rifting - only models (Figure 1.5). The Dominion Group was identified as the main phase of rapid mechanical subsidence with the West Rand and Central Rand Groups as the later part of the slower thermal subsidence phase (Bickle and Eriksson, 1982). Bickle and Eriksson, (1982) proposed a plate tectonic setting with extensional tectonics, in which a downwarped Witwatersrand basin, without rifting, was followed by a rifted

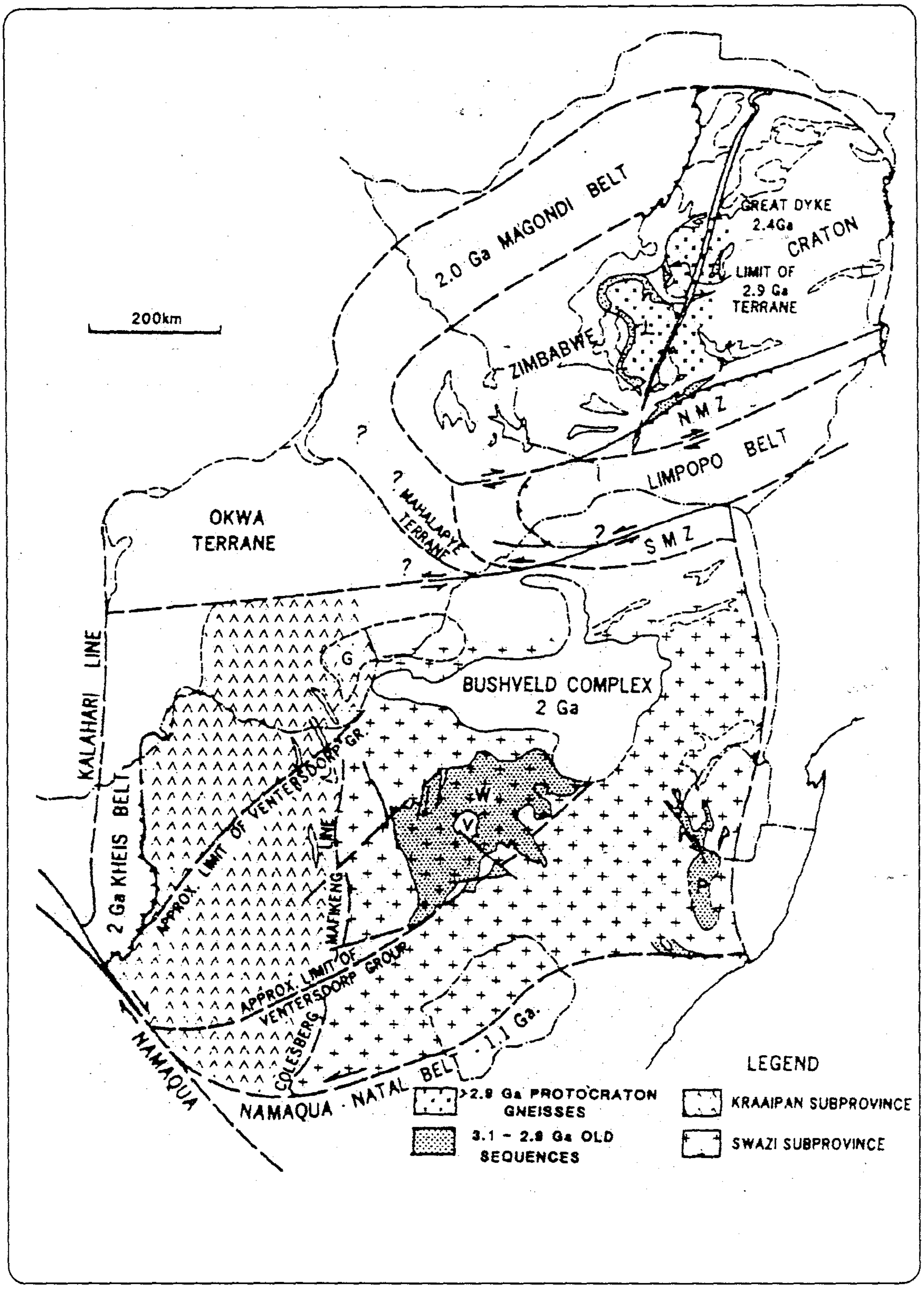


Figure 1.4 - Archaean to Early Proterozoic tectonic features on the Kaapvaal and Zimbabwe cratons of Southern Africa: G - Gaberone Granite; L - Lower Bulawayan (. 2.9 Ga) Group; NMZ - Northern Marginal Zone, Limpopo Belt; P - Pongola Supergroup; SMZ - Southern Marginal Zone, Limpopo Belt; V - Vredefort Structure; W - Witwatersrand basin (From Hartnady and Stowe, 1991).

Ventersdorp succession. Clendenin et al. (1988) proposed a three stage rift system with a number of superimposed basins developed on the Kaapvaal craton, which represented a fully evolved rift system. Clendenin et al. (1988) assigned the entire series of Archaean basins (Dominion - Witwatersrand) to what they called pre - graben protobasins, followed by the main graben development during the Ventersdorp Supergroup deposition and terminated with a post graben development of the Chuniespoort/Ghaap depositional basins.

In the mid - `80`s a cratonic foreland basin became a favoured basin model concept (Figure 1.6). Burke et al. (1986) reviewed the geological history of the Witwatersrand basin and concluded that the basin formed in a retro - arc, foreland setting resulting from subduction of oceanic crust beneath the Kaapvaal craton causing the development of the continental volcanic arc along the craton margin of the hinterland of the Witwatersrand basin. The Dominion and West Rand Groups were assigned to this phase of basin development. Subsequent to this event, the Kaapvaal craton collided with the Zimbabwe craton at about ~2.7 Ga, causing the uplift of the source area represented by the continental volcanic arc, which became the gold - rich source of the Central Rand Group. Winter (1987) independently came to a similar conclusion as Burke et al. (1986) that the Witwatersrand basin is reconcilable with a back - arc, foreland setting. (Figure 1.6)

Stanistreet and McCarthy (1991), Robb (1991) and Jackson (1992) suggested a more complex history for the development for the Witwatersrand basin i.e. an impactogenal model (Figure 1.7). They envisaged the Dominion Group as an early rift stage of the basin development, followed by the West Rand Group and its Pongola equivalent belonging to the thermal subsidence phase of a cratonic basin. A foreland basin developed accompanied by oceanic crust subduction beneath the northern margin of the Kaapvaal craton during which time the upper West Rand Group and lower Central Rand Group sediments were deposited. The collision stage followed and led to continent escape tectonics and the development of strike - slip faults bounding discrete blocks controlling the synsedimentary deposition of the upper Central Rand Group. This was followed by the extrusion of the Klipriviersberg Group lavas and deposition of the Platberg sediments as a result of the impactogenal rifting due to the indentation of

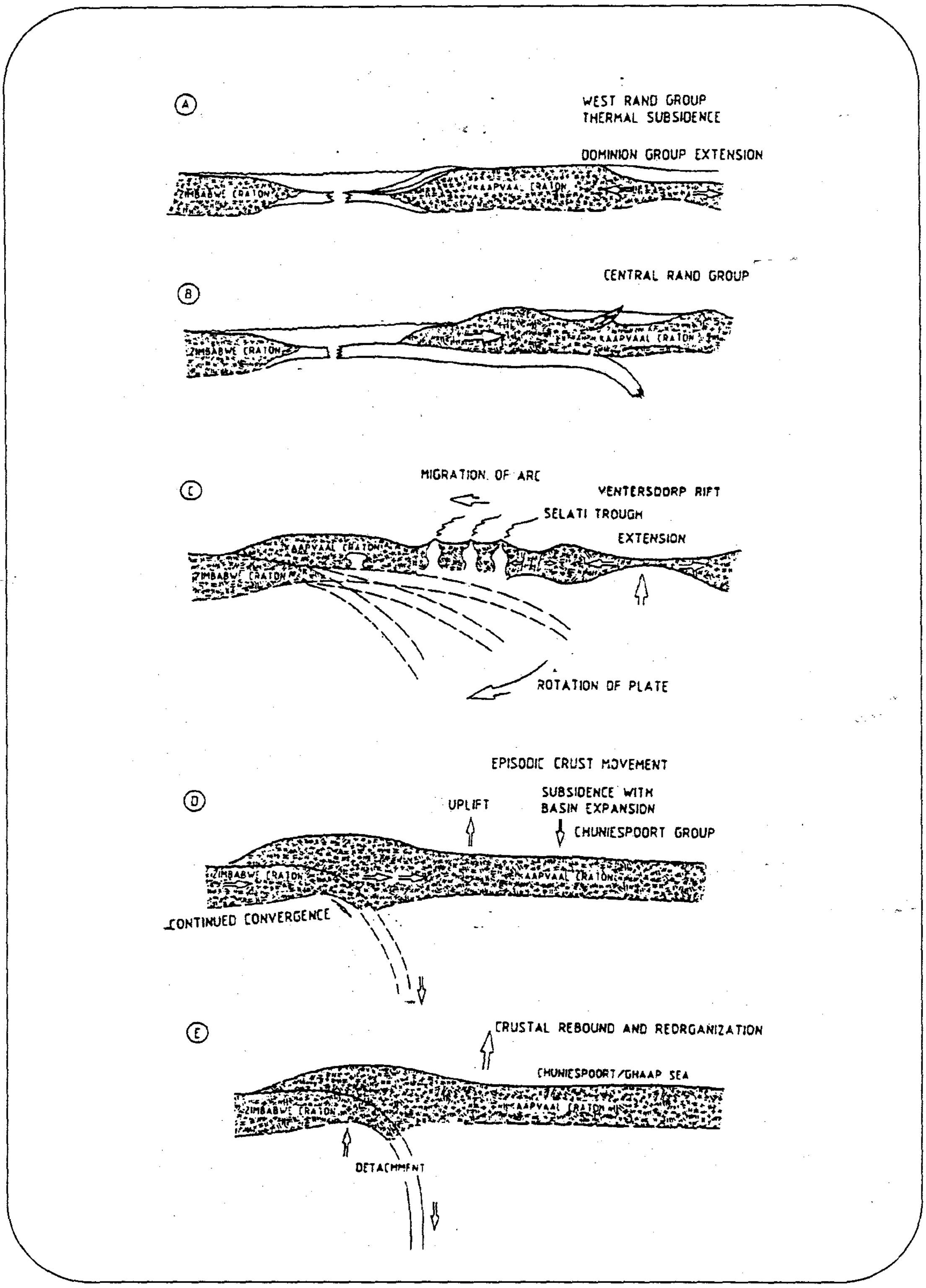


Figure 1.5 - Cartoons of the relationships of successor basins to proposed subduction and collision between the Kaapvaal and Zimbabwe Cratons. **A**. Intraplate deformation during Dominion and West Rand Group time. **B**. Low - angle subduction and intraplate deformation during Central Rand Group time. **C**. Roll - back of subducted slab following collision and resulting intraplate deformation during Ventersdorp Supergroup time. **D**. Continued convergence and effects of terminated inplane stresss during Chuniespoort/Ghaap Group time. **E** Crustal rebound and reorginazation leading to the pre - Pretoria Group unconformity following detachment of subducted slab (From Clendenin et al., 1988).

the Zimbabwe craton into the Kaapvaal craton (Limpopo Orogeny ~2700 Ma). The likelihood of a Central Rand Group representing a time of escape tectonism resulting in strike - slip dominated transtensional and transpressional basins was seriously considered as an alternative model (Myers et al., 1990; McCarthy et al., 1990).

According to Hartnady and Stowe (1991) and De Wit et. al. (1992) the latest accepted age dating of ~2.9 Ga for the Central Rand Group is inconsistent with the previous hypothesis (Burke et al., 1986 and Stanistreet and McCarthy, 1991) that a foreland basin developed during Central Rand times as a result of the Limpopo Orogeny (~2700 Ma).

Hartnady and Stowe (1991) have modified the latest favoured cratonic, foreland basin model based on new geochronological and geochemical evidence, by proposing a Kraaipan arc - collision model (Figure 1.8). They subdivided the Kaapvaal craton into two distinct subprovinces ie an eastern, older *Swazi* - and a western, younger *Kraaipan* subprovince (Figure 1.4). These two subprovinces are separated by a low - angle, west - dipping structure, equated with the Colesburg - Mafikeng geosuture, which formed as a result of a collision between the Swazi continental nucleus and the east - facing Kraaipan arc during late West Rand times. Subsequent to this event, the Central Rand basin formed as a peripheral foreland basin on the subducted plate, accompanied by arc polarity reversal, leading to later inception of a east - to southeastward subduction of lithosphere beneath a newly accreted Kraaipan active margin. The roll - back of the trench axis on the underriding plate coupled with the absolute divergent motion of the overriding plate, may have led to the Klipriviersberg volcanism in a retro - arc environment (Figure 1.8).

De Wit et al., (1992) concluded, after following the same model as Burke et. al., (1986) and Stanistreet and McCarthy, (1991), that the northward thrusting and age dating of 2.9 Ga of the Witwatersrand basin is inconsistent with a simple foreland model. Therefore, they suggest an accretionary orogeny caused by the collision of a series of oceanic and continental fragments, with sediments being shed southward into piggy - back basins, overprinted by out - of - sequence northward thrusting. However, the northward verging thrust faults are not of Witwatersrand age as suggested by De Wit et

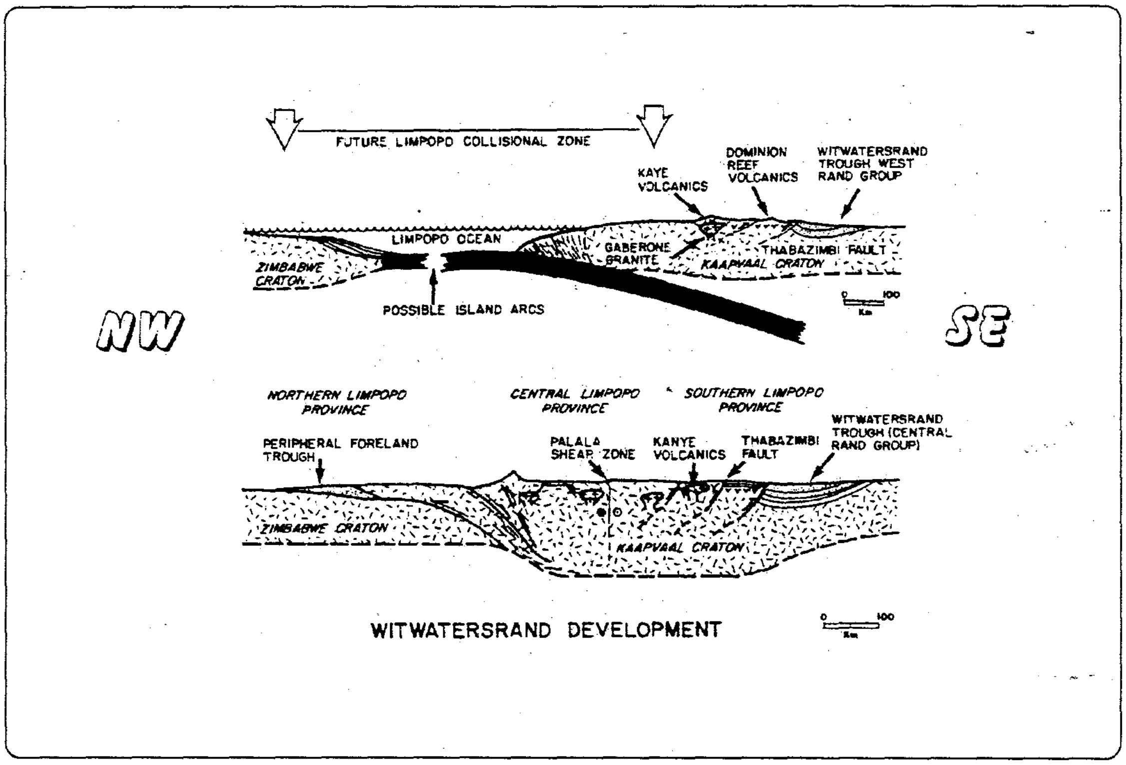


Figure 1.6 - Intepretation by Burke, Kidd and Kusky (1986) of possible plate - tectonics events: (above) the West Rand Group as a retro - arc basin, (below) the Central Rand Group as the product of continent - continent collision (From Winter, 1987).

al., (1992), but rather syn - Ventersdorp to post - Pniel age (Pitts, 1990; Vermaakt, 1994).

Maynard and Klein (1995) recently attempted a quantitative subsidence analysis of the Witwatersrand basin based on the latest publications of high resolution U - Pb dates from single zircons (Barton et al., 1989; Robb et al., 1990a and Armstrong et al., 1991). It appears that the results of the subsidence curve are more consistent with a pull - apart basin rather than a foreland or rift basin, however the geochronological data provides too poor a constraint to produce definite conclusions.

Maynard and Klein (1995) envisaged a basin initiated as a simple rift during the early - rift phase of the Dominion conglomerate deposition, followed by the extrusion of large volumes of lava. A long period followed with no obvious basin subsidence for at least a 100 m.y., reflecting a shift to a transpressional regime. With the onset of the West Rand deposition, it abruptly changed to an extensional phase, reflecting a shift to transtension, and pull - apart basin formation, accompanied by rapid subsidence and subsequent marine transgression. The rate of subsidence waned, resulting in a slow accumulation of non - marine Central Rand Group sediments, with a reduction in the basin size. The presence of reverse faulting during this period, shows the return to a phase of transpression (Stanistreet and McCarthy, 1991).

Of all the models, the cratonic foreland model is presently favoured as a tectonic model for the Witwatersrand basin since it accounts for many of the observed features of the basin and its lithological infill such as:

- the northwestern, western and northern margins of the basin are interpreted as thrust fault bounded, marginal to a contemporaneous fold - thrust belt and an open southeast margin, at least for some part of the basin development with probable uplift during the waning phase of basin development (Olivier, 1965; Winter, 1994)
- an asymmetric profile with thicker strata and steeper dips occurring toward the fault - bounded northwestern, western and northern margins

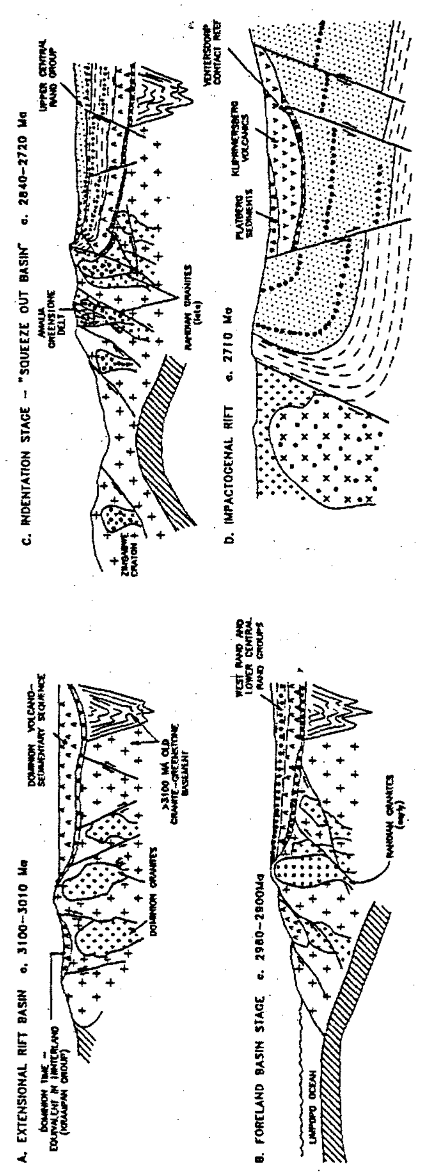


Figure 1.7 - Tectonic and depositional framework for the Witwatersrand Triad as envisaged by Stanistreet and McCarthy (1991), with time constrains imposed by recent U - Pb zircon ages (From Robb et. al., 1991).

- it comprises a lower flysch - like facies (West Rand Group) and an upper mollasse - like facies (Central Rand Group)
- there is a paucity of magmatic activity in the basin
- the basin margin nearest to the fold - thrust belt shows many unconformities related to erosion and cannibalisation of the underlying placers which lose their conformable relationships southeasterly into the more distal basin (Winter, 1994)
- compressional tectonics were clearly, in part, synsedimentary, younging and becoming weaker away from the active margin
- there was a definite basinward migration of the margin nearest to orogen and the depositional axis
- there are indications that the basin was decreasing in size during Central Rand Group times. This can be attributed to viscoelastic response to loading of the lithosphere (Burke et al., 1986).

Many aspects of the taphrogenic model of Pretorius (1981) can be discarded, since they do not accord with observed facts. Depositional patterns were not related to a long - lived pattern of interference folding. The goldfields are not defined by a contiguous fan deltas of haphazard distribution, but are defined by the marginal infilling of a marine basin, either open ended or with a sea - way linked to an ocean, hence the continuity of sequences laterally and the increase marine aspects in the distal parts of the basin (Winter, 1994).

The Stanistreet - McCarthy (1991) impactogenal model presents innovative exploration target opportunities but to maintain a general southeasterly palaeoslope so characteristic of the Central Rand Group sedimentation, seems unlikely with rotational blocks in a strike - slip tectonic regime. So far intra - basinal exploratory drilling adjacent to certain of these structural blocks has not substantiated this model (Tainton, 1994).

A quantified subsidence analysis of the Witwatersrand basin has long been outstanding requirement for Witwatersrand studies. The application of sequence stratigraphy and basin subsidence analysis should be able to provide some insight into the tectonic

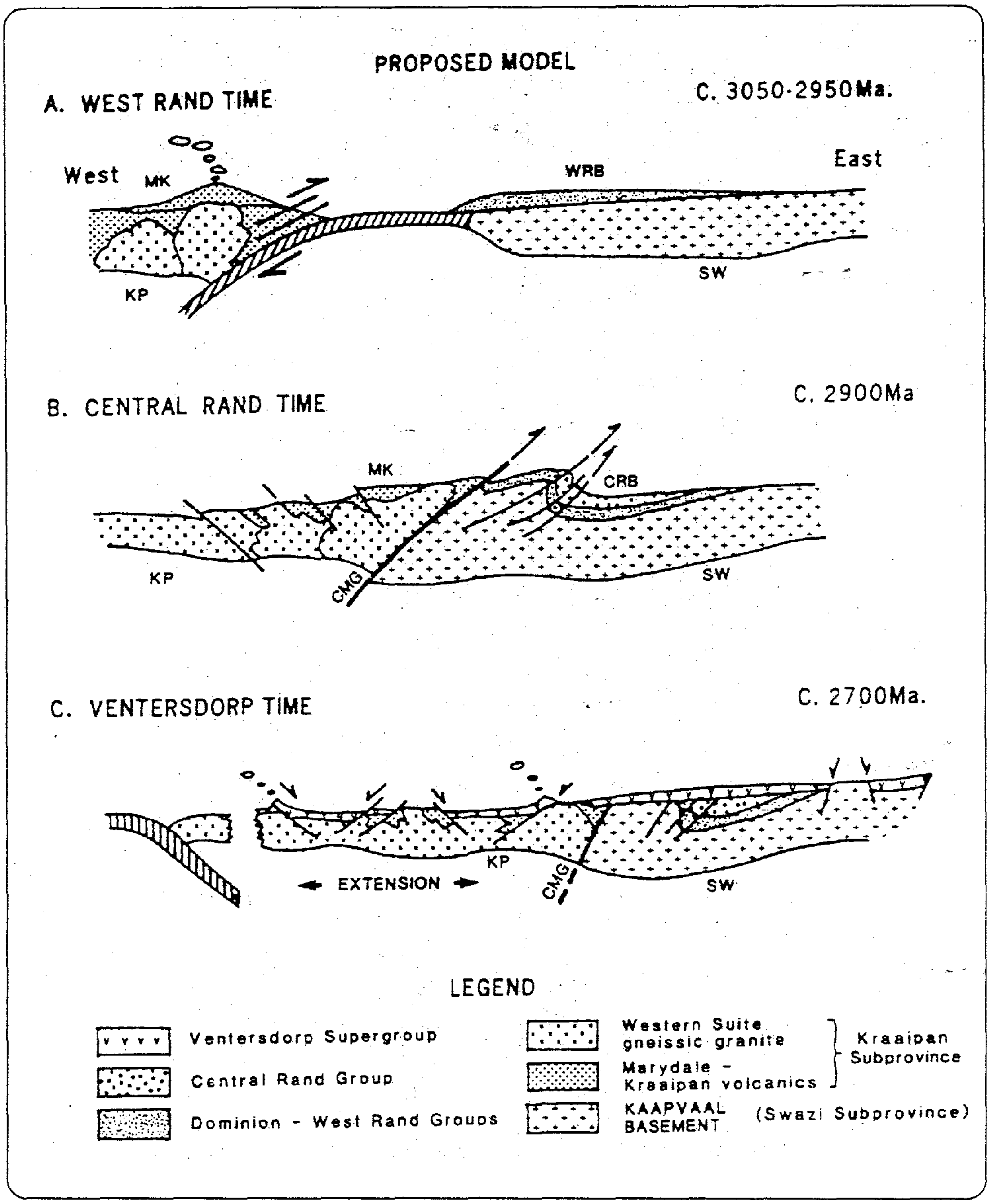


Figure 1.8 - Proposed crustal evolution model for the western Kaapvaal Craton (**above**) West Rand time (3.1 - 3.0 Ga) convergence of Kraapan Subprovince (KP) Arc greenstones, forearc basin deposits and granites, with the Swazain Subprovince (SW) basement and West Rand basin (WRG) (**middle**) Central Rand time (~2.9 Ga), collision, thrusting, granitic plutons and formation of Colesburg - Mafikeng gravity anomaly (CMG) at suture. Central Rand Group (CRG) deposited in a foreland basin (**below**) Platberg time (~2.7 Ga), new subduction zone with reversed dip direction and eruption of lower Ventersdorp sequences in an extensional setting (From Harnady and Stowe, 1991).

evolution of the Witwatersrand basin and substantiate or refute some of the proposed models.

1.3.5 Gold mineralization in the Witwatersrand

A remarkable feature of the Witwatersrand stratigraphy, particularly the Central Rand Group, is the richness in gold concentration, the number of placers developed and their lateral continuity throughout the basin. The Central Rand Group can be subdivided into the basal Johannesburg and overlying Turffontein Subgroups, each containing several important economic placers. In the Johannesburg Subgroup there are two main formations have exploitable reef zones. For instance, the Carbon Leader and Middelvlei Reefs of the Main Conglomerate Formation, are extensively mined in the Carletonville goldfield. On the Central Rand the Main Reef was the important gold producer, whereas the South/Nigel Reef was the main producer in the East Rand goldfield. In the Klerksdorp goldfield the Vaal Reef is mined for its gold content, which can be correlated with Bird Reef (Krugersdorp Formation) in the West Rand and Central Rand goldfields where it is also exploited for gold. In the Welkom goldfield mainly the Basal/Steyn Reef is exploited. The Kimberley reefs occur in the Turffontein Subgroup with the Kimberley Reef/UK9a or May Reef being the most important producer of gold and uranium higher up in the Turffontein succession two more reef packages have been exploited on a localised scale in several goldfields (SACS, 1980) ("Reef" in this context is equated with an ore body and not a coral reef) (Figures 1.2 and 1.3).

During the 1980's, the solution to the problem of the genesis of the Witwatersrand ores was thought to be found in the "Provenance" of the gold source, rather than the "process" of gold and uranium concentration. It is time to review the "processes" that concentrate gold and uranium from a sequence stratigraphic point of view. Sequence stratigraphy enables one to understand the environment in which these placers develop to determine their temporal and spatial relationship are with the underlying and overlying sequences.

The following discussion elaborates on some of the models that were proposed for the origin of the gold and uranium in the Witwatersrand placers. Historically, a placer (Ramdohr, 1958) and a hydrothermal model (Davidson, 1955) have been proposed for the origin of gold in the Witwatersrand gold deposits. A long standing idea was that the gold was derived from Archaean greenstone belts and the uranium from the associated granitic rocks. Observations favourable to a placer model are the close relationship between the gold distribution and sedimentary features, such as unconformities, pyritiferous quartzites, conglomerates and carbon seams. The placerist contradicted the hydrothermal origin by noted the lack of alteration in the deposit, low permeability resulting in poor postdepositional fluid movement, lack of vertical zoning and inability of fluids to transport gold (Whiteside et al., 1976; Minter, 1978; Hallbauer, 1986). Recently, the modified placer hypothesis has emerged as one of the more acceptable theories to explain some of the problems which the hydrothermalist so often underline. This model is based on the assumptions that most of the gold and other heavy minerals were fluvially transported from the hinterland to the basin edge. Here they were subjected to further mechanical and chemical processes of reworking and concentration. Subsequent diagenetic processes, modified the mineral spectrum further through solution, recrystallization and remobilisation of the gold and other components over a short lateral distances (eg. Tainton 1994).

Advocates of precipitation saw biogenic activity as an important factor in the concentration of gold. Gold was dissolved during the weathering process by cyanogenic micro - organisms and transported as organic - protected colloids. Precipitation took place in association with organic matter, virtually contemporaneous with the sedimentary deposition. The precipitated grains were then reworked mechanically, mobilised and reprecipitated during subsequent diagenesis and later metamorphism (Reimer, 1984).

The epigeneticists assert that metamorphism has played the major role in the genesis of the mineralising fluids (Phillips and Meyers, 1987; Phillips et al., 1987 and Phillips, et al. 1989a & b). The hydrothermal fluids were generated by dehydration of a thick accumulation of argillites in the lower part of the stratigraphic succession, during greenschist facies metamorphism. Indirectly it is implied that the gold was present as a primary constituent in the shales. Gold was transported as sulphur (thio) - complexes and precipitated when the sulphur was involved in the sulphidization of heavy oxide heavy minerals to form pyrite.

Hutchinson and Viljoen (1988) viewed the source of gold as endogenous rather than a exogeneous as postulated by most other investigators. They considered the source of the gold to originate from pyritic auriferous exhalites leached from andesitic - basaltic volcanism of the Dominion and West Rand Group, deposited proximal to fluid vents along the active marginal faults and becoming concentrated by mechanical reworking associated with fluvio - deltaic processes along a regressive basin edge.

Another group of researchers, e.g. Robb and Meyer (1985, 1987); Robb et al., (1990b); Robb et. al., (1991); Hallbauer (1986); Hallbauer and Barton (1987) and Klemd and Hallbauer (1987) studied the granitic components of the region west, north and northwest of the basin. They suggested that the "provenance" is important in understanding the source of the gold and concluded that hydrothermally altered granitic rocks to the northwest and west of the basin could be the source of the gold and uranium.

In all the above models, co - occurring uraninite was derived from erosion of peraluminous granites surrounding the basin, which have been hydrothermally altered and therefore was viewed as being exogenous.

A fourth model is magmatic back - arc version of the modified placer theory which is in fact a proposition attempting to incorporate all three schools of thought. It envisages the gold and uranium entering the conglomerate and other sediments as detrital particles, as dissolved constituents of fluvial waters, and as components of hydrothermal fluids at more or less the same time, in response to coeval tectonism, granitic magmatism, associated hydrothermal and metasomatic alteration, accompanied with highly charged surface waters, erosion of uplifted margin, contemporaneous with the sedimentation of the Central Rand Group (Pretorius, 1991).

This paper attempts to quantitatively describe the origin and depositional processes of placers within the context of basin analysis, geohistory and a sequence stratigraphic framework. To facilitate this, a comprehensive review of basin analysis principles and methodology is initially presented. This is followed by application to the Witwatersrand sequence.

2 CONTROLS ON BASIN STRATIGRAPHY

2.1 Basins in their Plate Tectonic Environment

Sedimentary basins are the manifestation of prolonged subsidence at or near the surface of the earth, irrespective of the tectonic domain in which they occur (Allen and Allen, 1990). The processes within the relatively rigid, cooled thermal boundary layer of the earth`s lithosphere, are ultimately responsible for the driving mechanism of subsidence. The lithosphere is composed of several moving plates in which these sedimentary basins exist, where stratigraphy is the long - term response of the depositional surface to prolonged subsidence (Allen and Allen, 1990).

The earth`s interior is essentially made up of several compositionally and rheologically different zones (Figure 2.1). The main compositional zones consist of an outer lithospheric crust able to store elastic stresses over long time scales, a mantle and a core. The rheological boundary between the lihosphere and asthenosphere is fundamental in the formation of sedimentary basins. Of interest in basin analysis is the differentiation between the rheological zonation of the lithosphere and asthenosphere, because the vertical motions (subsidence and uplift) of sedimentary basins ultimately have a response to the deformation of the uppermost rheologic zone. The continental lithosphere displays a distinct strength profile with depth. A weak ductile zone exists in the lower crust, which separates a brittle upper crust from the upper mantle. The oceanic lithosphere lacks this low - strength layer and its strength increases with depth to the brittle - ductile transition in the upper mantle (Allen and Allen 1990).

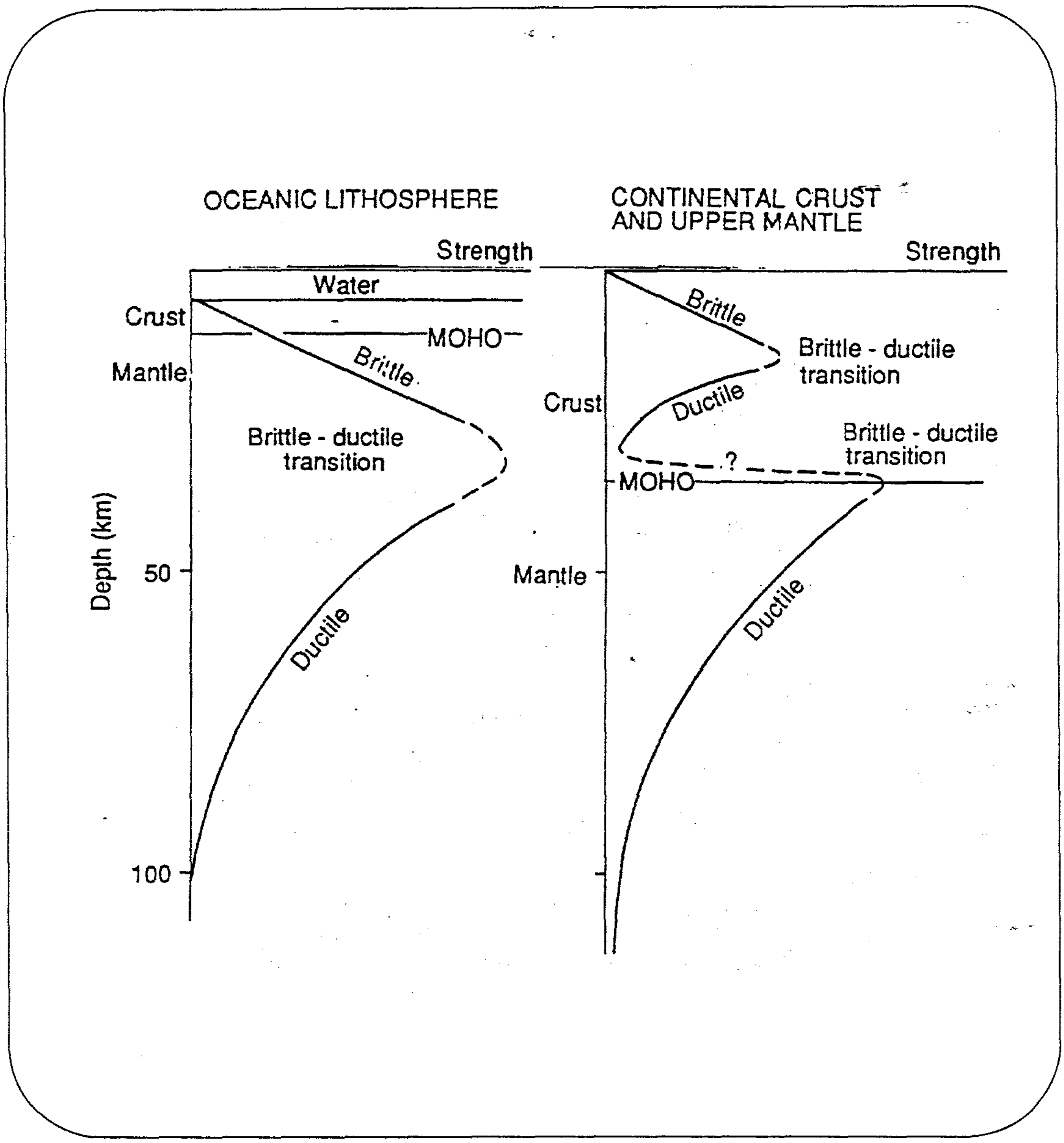


Figure 2.1 - The strength of the oceanic and continental lithosphere as a function of depth (Molnar, 1988). The oceanic lithosphere has a strong elastic core extending to depths of over 50 km, whereas the continental lithosphere appears to have a weak ductile layer in the lower continental crust. This gives a rheological layering like a jam sandwich. The elastic lithosphere is the upper portion that is able to store elastic stresses over geological periods of time. The base of the thermal lithosphere is a mechanical boundary separating the relatively strong outer shell of the lithosphere from the very weak asthenosphere (Allen and Allen, 1990)

Table 1.1 Basin classification of Bally and Snelson (1980)

1 **Basins located on rigid lithosphere, not associated with formation of megasutures**
- 1.1 Related to formation of oceanic crust
  - 1.1.1 *Rifts*
  - 1.1.2 *Oceanic transform fault associated basins*
  - 1.1.3 *Oceanic abyssal plains*
  - 1.1.4 *Atlantic-type passive margins (shelf, slope & rise) which straddle continental and oceanic crust*
    - 1.1.4.1 Overlying earlier rift systems
    - 1.1.4.2 Overlying earlier transform systems
    - 1.1.4.3 Overlying earlier backarc basins of (321) and (322) type
- 1.2 Located on pre-Mesozoic continental lithosphere
  - 1.2.1 *Cratonic basins*
    - 1.2.1.1 Located on earlier rift grabens
    - 1.2.1.2 Located on former backarc basins of (321) type

2 **Perisutural basins on rigid lithosphere associated with formation of compressional megasuture**
- 2.1 Deep sea trench or moat on oceanic crust adjacent to B-subduction margin
- 2.2 *Foredeep and underlying platform sediments,* or moat on continental crust adjacent to A-subduction margin
  - 2.2.1 Ramp with buried grabens, but with little or no blockfaulting
  - 2.2.2 Dominated by block faulting
- 2.3 *Chinese-type basins* associated with distal blockfaulting related to compressional megasuture and without associated A-subduction margin

3 **Episutural basins located and mostly contained in compressional megasuture**
- 3.1 Associated with B-subduction zone
  - 3.1.1 *Forearc basins*
  - 3.1.2 *Circum Pacific backarc basins*
    - 3.1.2.1 Backarc basins floored by oceanic crust and associated with B-subduction (marginal sea *sensu stricto*).
    - 3.1.2.2 Backarc basins floored by continental or intermediate crust, associated with B-subduction
- 3.2 Backarc basins, associated with continental collision and on concave side of A-subduction arc
  - 3.2.1 On continental crust or *Pannonian-type* basins
  - 3.2.2 On transitional and oceanic crust or *W. Mediterranean-type* basins
- 3.3 Basins related to episutural megashear systems
  - 3.3.1 *Great basin-type* basin
  - 3.3.2 *California-type* basins

The nature and rates of plate tectonic motion govern every aspect of the geodynamics involved in basin formation and its environment (Figure 2.2). This led to a classification scheme of sedimentary basin based on plate tectonics. The most widely used classification is the one by Bally and Snelson, (1980) (Table 1.1).

It is necessary to consider some of the mechanics of sedimentary basin formation. These fall into three classes. All three mechanisms may have played a vital role in the evolution of the Witwatersrand basin (Allen and Allen, 1990). (*1*) *Purely thermal mechanics* are important when the oceanic lithosphere is *cooling* and subsiding as it moves away from spreading centres. This which explains oceanic bathymetry. *(2) Changes* in the *crustal/lithospheric thickness* involves the *thinning* of the crust by mechanical stretching, accompanied by extensional fault - controlled subsidence and subsequent thermal doming of the lithosphere due to the thinning of the crust. *(3) Loading and unloading* of the lithosphere on a small scale (e.g. volcanoes and

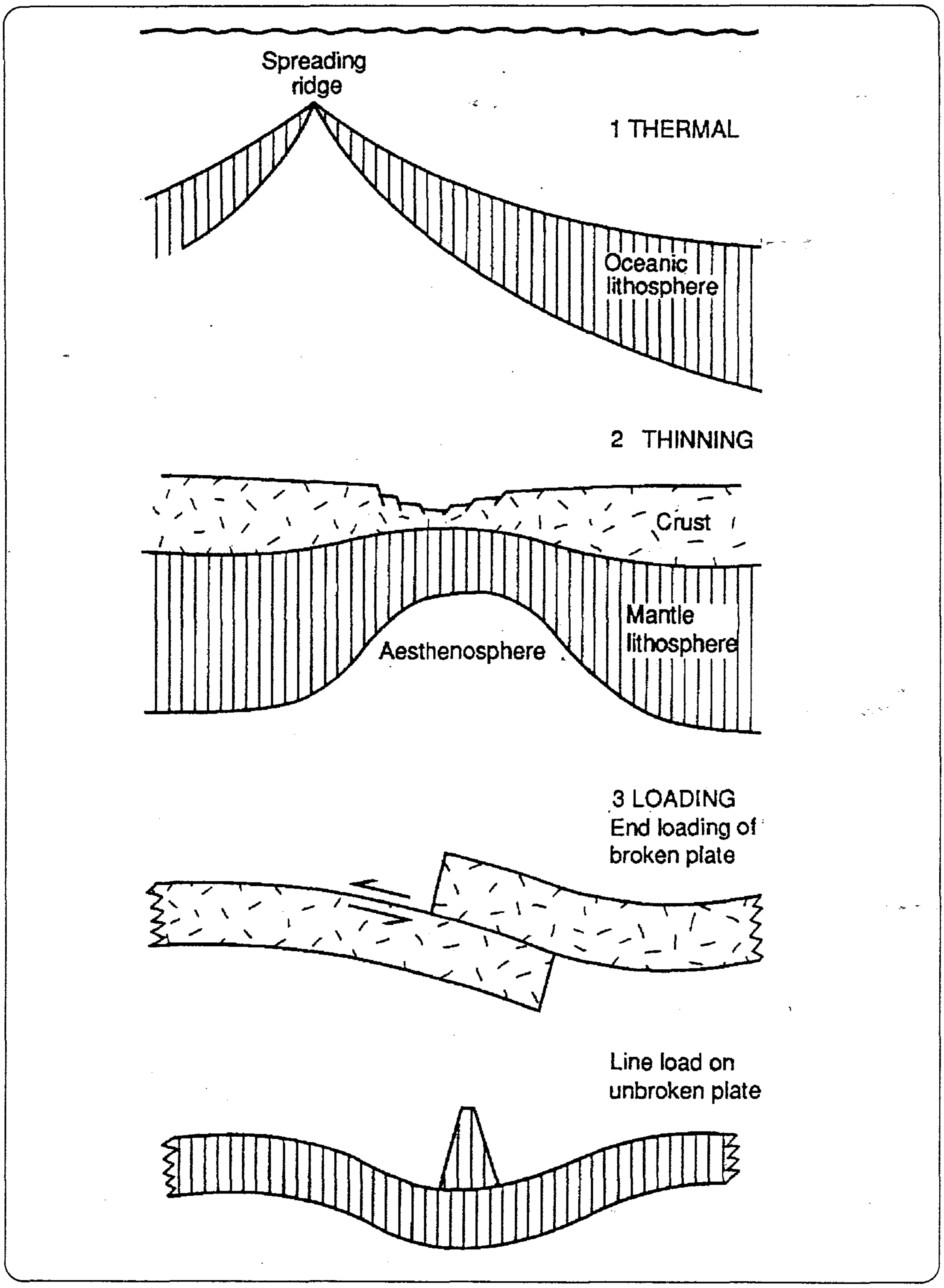


Figure 2.2 - The three basic mechanisms for basin subsidence (From Allen and Allen, 1990).

seamount chains) or large scale (e.g. mountain belts) causes a deflection or flexural deformation which can lead to subsidence eg. foreland basins (Allen and Allen, 1990).

2.2 Primary Mechanisms of Basin Subsidence

2.2.1 Isostasy

To understand the processes of basin formation, one has to consider the primary mechanisms that influence basin subsidence. These include isostasy, thermal effects, flexure and intra - plate stresses. The principle of isostasy states that continents are buoyed up by a force equal to the weight of the displaced mantle. Adjacent blocks with different thickness and/or density structures will have different relative reliefs (Klein, 1991).

Two vertical columns illustrated in Figure 2.3 display the different lithospheric structures beneath continents and oceans and demonstrate the principles of isostasy. At some depth below surface there is no density contrast between the adjacent columns (asthenosphere of equal density underlies both columns). This elevation below surface which controls isostatic balance is known as the depth of compensation. It is this compensation which controls the elevation of the crustal topography of continents and ocean basins. Therefore, one can interpret that the earth's topography is in close approximation to equilibrium with isostatic adjustments (Klein, 1991b).

This is only true for passive situations and needs modification because in reality, horizontal and vertical stresses exist and their changing magnitudes govern the geodynamics of basin formation processes (Angevine and Heller, 1987).

By applying the formulas derived from the model of isostasy (Klein, 1991b), one can calculate the relative relief (Z) between two adjacent continental columns with different structures (Figure 2.3). These calculations can be repeated for a basin filled with water, a basin filled with air and a basin filled with sediment. From this equation one can deduce that a basin filled with water will be about 1.5 times deeper than the same basin

$$\rho_w Z + (1)\rho_L + \rho_a X = \rho_s S + (1)\rho_L + \rho_a (Z + X - S) \quad [1]$$

$$= \rho_s S + (1)\rho_L + \rho_a Z + \rho_a X - \rho_a S \quad [2]$$

Solve for S:

$$\rho_a S - \rho_s S = (1)\rho_L - (1)\rho_L + \rho_a Z - \rho_w Z + \rho_a X - \rho_a X \quad [3]$$

Cancel and simplify:

$$(\rho_a - \rho_s) S = \rho_a Z - \rho_w Z \quad [4]$$

$$= (\rho_a - \rho_w) Z \quad [5]$$

$$S = \left(\frac{\rho_a - \rho_w}{\rho_a - \rho_s}\right) Z \quad [6]$$

If $\rho_a = 3.3$ g/cc; $\rho_s = 2.3$ g/cc; and $\rho_w = 1.0$ g/cc [7]

then $S = 2.3\,Z$. [8]

Column 1 (left) = Column 2 (right)

$30\,(2.8) + 90\,(3.4)$

$$= Z(\ 1) + 15\,(2.8) + 45\,(3.4) + (60 - Z)(3.3) \quad [9]$$

$$390 = Z + 42 + 153 - 3.3(Z) \quad [10]$$

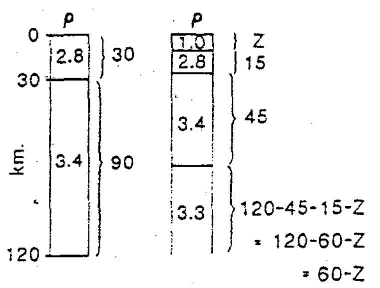
$$3.3Z - Z = 393 - 390 \quad [11]$$

$$2.3\,(Z) = 3 \quad [12]$$

$Z_w = 1.3$ km (depth of water in the basin) [13]

—Mathematical symbols used in this chapter.

β	Stretching factor
c	Coefficient of porosity reduction according to lithology
ϕ_N	Present porosity
ϕ_0	Original porosity
ρ_w	Density of water
ρ_L	Density of lithosphere
ρ_a	Density of asthenosphere
ρ_s	Density of sediment
X	Thickness of asthenosphere above depth of compensation
S	Sediment thickness
T_N	Present thickness
T_O	Original thickness
z	Coefficient of depth below surface
Z	Depth of basin
Z_a	Depth of basin filled with air only
Z_s	Depth of basin filled with sediment only
Z_w	Depth of basin filled with water only
ΔSL	Change in sea level



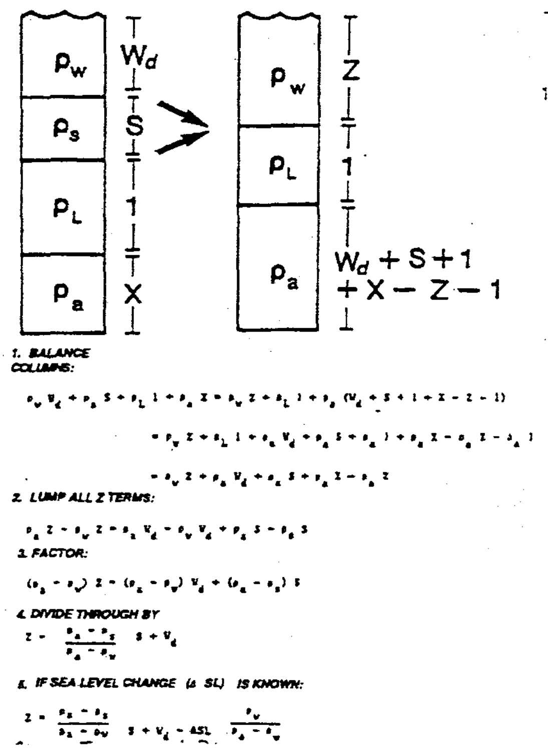


Figure 2.3 - Lithospheric profiles with sedimentary basins used for sample calculations. (**Left**) An example of the calculations for the two columns and derivation of the formula. The two columns are in isostatic balance. (**Above, right**) Mathematical symbols used in the calculations (From Klein, 1991) (Adapted from Angevine and Heller, 1987).

filled with air and also that the same basin filled with sediment will be about 2.3 times deeper than that filled with water (Angevine and Heller, 1987).

For typical sediment densities of 2.3 - 2.5 g/cm^3, the final thickness of a basin is in the order of 2 to 3 times the initial starting water depth. In reality this is not always valid, other factors are involved apart from sediment loading, since the majority of basins exceed the ratio of initial water depth vs final thickness of the basin (Angevine and Heller, 1987). Therefore, the principle of isostasy has demonstrated that there are other processes to explain the thicker sediments in the basin. These are tectonic subsidence (Angevine and Heller, 1987).

Other factors that need to be accounted for include:

- subsidence due to cooling. This can cause simple lithospheric stretching either by simple shear of the upper crust (block faulting) or pure shear (ductile necking) of the lower crust and lithospheric mantle
- heating of the lithosphere and its crust results in the uplift and subsequent crustal thinning due to surface erosion, with subsequent isostatic subsidence due to cooling of the attenuated crust and associated increase in its density.
- emplacement of dense material into the continental lithosphere by the intrusion of dense ultramafic dykes/diapirs or thrusted ophiolites

These other driving mechanism become important during the interpretation of tectonic subsidence curve of the Witwatersrand basin, after the effect of loading by sediments have been isolated from the total subsidence curve through a method called backstripping.

2.2.2 Flexure subsidence

Part of the Witwatersrand basins tectonic evolution is typical to the formation of a foreland basin in a retro - arc setting (Burke et al., 1986; Winter, 1987). Flexure subsidence therefore plays a role in the evolution of the Witwatersrand basin and an understanding of the ability of the elastic lithosphere to support significant bending stresses is necessary.

Flexure subsidence differs from isostasy in that it assumes finite strength for the lithosphere and relative rigidity, whereas isostasy assumes local compensation (Angevine and Heller, 1987).

The whole concept of rigidity implies that the lithosphere has the ability to transmit elastic stress which means that vertical movement of the earth's crust can be inhibited by the bending rigidity of the near surface rocks. The downbending of continental lithosphere that results from transmission of these elastic stresses is the main mechanism in the development of foreland basins (Grotzinger, 1990).

If a vertical load is applied to the lithosphere, the plate deforms by regional isostasy and regional compensation. Although the load may cause the greatest impact at its point of loading, it deforms the adjacent blocks which are linked over a broader terrane forming a basin. This whole process is driven by compressional tectonics involving thrusting and sediment deposition, amplifying the tectonic loading (Klein, 1991b).

The dimension of foreland basins depends on the magnitude of the overthrusted load and the elastic thickness of the lithosphere which is being depressed by flexure. The temperature largely controls the elastic thickness of the lithosphere and therefore the expected greater thickness of the cratonic regions and the thinner elastic thickness in newly rifted regions. Consequently, relatively deep and narrow foreland basins will develop on relatively weak (hot) lithospheric plates, and whereas shallower, broad foreland basins developed on a strong (cold) lithosphere (Grotzinger, 1990).

The lithosphere behaves like a elastic beam of some assumed or calculated rigidity, which also has an important influence on the variation in the dimensions of the flexure of the basin. If the crust has a high rigidity, the foreland basin will be broader but shallower. If the crust has a low rigidity, the foreland basin will become narrower but deeper (Angevine and Heller, 1987) (Figure 2.4).

2.2.3 Thermal subsidence

Another primary mechanism that may have an influence on the early part of the evolution of the Witwatersrand basin is thermal subsidence. This means that a part of the Witwatersrand basin thermal subsidence history can be influenced by deeply buried rift basins beneath the passive margin during the early phase of the basin's development. This fact is manifested by the presence of a preserved Dominion Group occurring immediately beneath the Orange Grove Quartzite Formation with an apparent conformity, suggesting a close affinity of the Hospital Hill Subgroup with the Dominion Group (Tainton, 1994). Therefore it is necessary to understand the evolution of oceanic - and continental rift basins that may evolve into thermally - subsiding post - extensional basins.

Thermal effects on crustal blocks that are in isostatic balance can lead to crustal doming, because heated rocks expand and become less dense, while maintaining isostatic balance. While the lithosphere cools it is restored to its original thickness and density. The crest of the updoming can be eroded and the crust is thinned forming a simple sedimentary basin. The lithosphere heats up rapidly but cools more slowly by conductivity. Cooling of by means of conductivity is a function of the square - root of time ($t^{½}$) and after tens of millions of years it cools exponentially ($e^{-t/t}$) (Klein, 1991b; Angevine and Heller, 1987).

McKenzie (1987) proposed a model for crustal extension and thermal subsidence as a response to heat addition into the lithosphere by a rising asthenosphere (Klein, 1991b; Angevine and Heller, 1987). In figure 2.5 an example starts prior to time 1 with a layered crust (C), lower lithosphere (L), asthenosphere (A) and a normal geothermal gradient. No subsidence has occurred yet between time 0 and 1. If the crust is

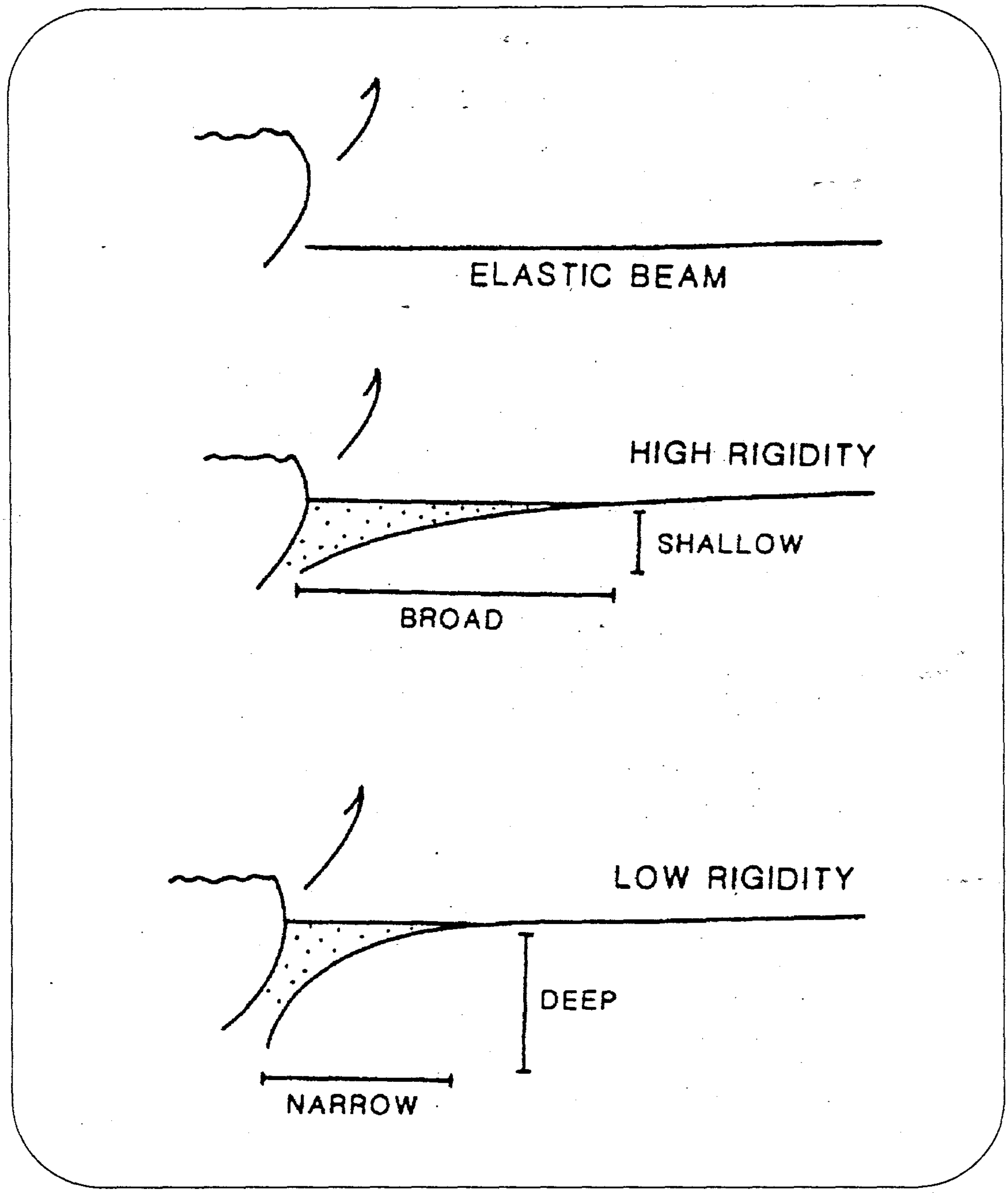


Figure 2.4 - Elastic beam model for flexural load with changing crustal rigidity (From Klein, 1991b) (Adapted from Angevine and Heller, 1987).

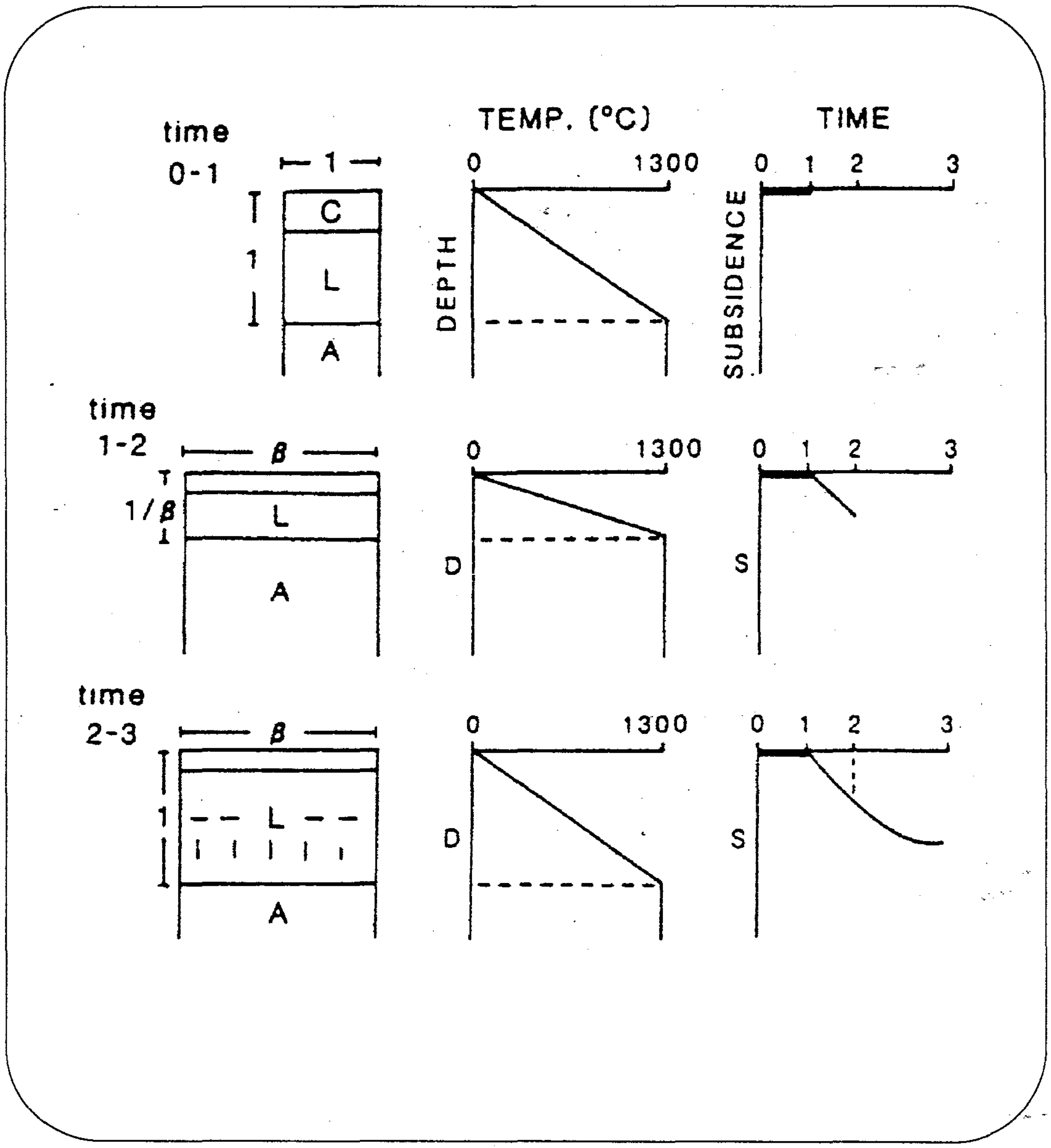


Figure 2.5 - Thermal subsidence as per stretching model of McKenzie (1978). (From Klein ,1991b) (Adapted from Angevine and Heller, 1987). (Abbreviations : C - Crust; L - Lower Lithosphere; A - Asthenosphere).

stretched by factor ß between time 1 and 2, then the lithosphere thickness will reduce to 1/ß. The geothermal gradient will become steeper, because the asthenosphere rises closer to surface in response to the stretching. This subsidence is not thermal subsidence but the local isostatic compensation to the thinning of the lithosphere. Between time 1 and 2 the mechanical subsidence rate is rapid changing as indicated by the steeper curve. After time 2 the lithosphere cools and thickens, the basin subsides by thermal decay and the density increases as warm asthenosphere is converted to cool lithosphere. The original geothermal gradient is restored at the end of this process. Subsidence will continue as the less dense asthenosphere is converted to more dense material until such a time as the original geothermal gradient is restored. The rate of cooling and subsidence will be exponential. In McKenzie`s example there are two stages of subsidence ie an initial phase of subsidence that occurs during extension of the lithosphere and a second phase of thermal subsidence of the lithosphere at an exponential rate, which follows once the cooling of the lithosphere and the extension is complete (Angevine and Heller, 1987; Klein, 1991b). Several considerations emerged from these findings which led to several models for explaining extensional basins.

White and McKenzie (1988) proposed a two layer model with both the lithosphere and asthenosphere being stretched differentially. They were modelling the "steers head" shape basin which is defined by a central basin overlain by a saucer - shaped basin - fill extending over a larger area than the underlying rift (Klein, 1991b).

Several other explanations for thermal subsidence and crustal extension have been proposed. Royden et al. (1980), proposed that similar thermal perturbations can be developed by intrusion of extensive dyke sheets. Thermal subsidence will then occur with minimal stretching in response to a thermal event, without any significant extension (Klein, 1991b).

A pure shear model involves uniform stretching where the principal stress field is focused underneath the axis of the rift zone. A series of listric faults are developed in the brittle zone of the earth`s crust which flange into a brittle - ductile zone below it and merge in the centre of the basin (Klein, 1991b) (Figure 2.6a).

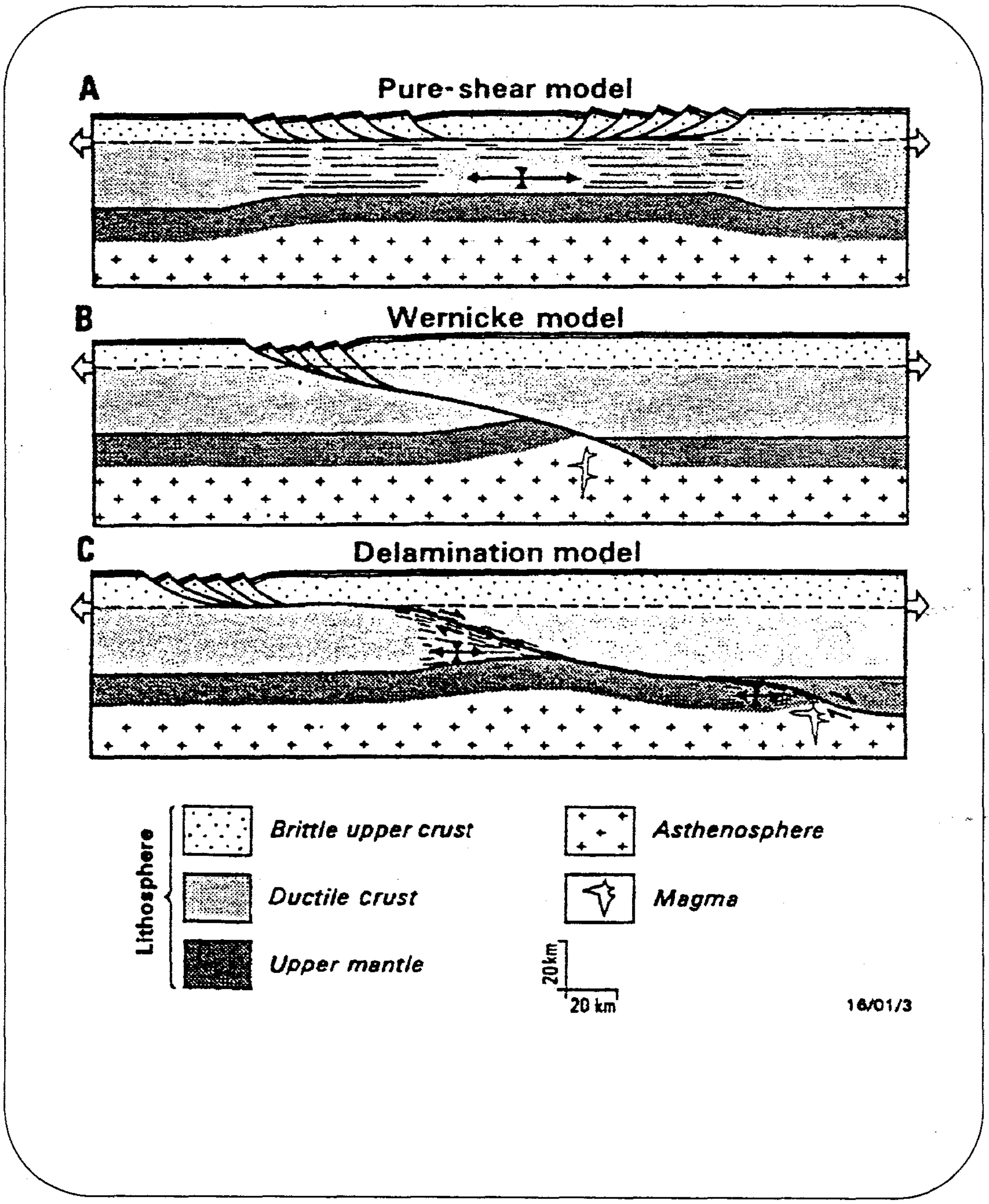


Figure 2.6 - Three models for continental extension (Lister et al., 1986) (From Klein, 1991b).

Wernicke (1985) challenged this model, arguing that a simple shear detachment surface cuts through the crust to the upper mantle, displacing the crustal thinning significantly away from the basin centre (Klein, 1991b) (Figure 2.6b & c).

2.3 Tectonic Mechanisms

2.3.1 Intraplate stress

Another aspect regarding the theoretical background of basin formation processes is the phenomenon of plate stresses and changing plate stresses. Cloetingh, (1988) established a correlation between flexural wave length and levels of flexural rigidity. For instance, if the crust is subjected to positive tension and negative compression, contrasting stress levels are produced which can cause significant vertical subsidence. Whereas, if only under compression, one would expect a large degree of buckling depending on the age of the lithosphere, but the predicted load stress will exceed the load strength of the continental crust.

Intra - plate stresses and stresses occurring at plate margins can be distributed over long distances and large areas far into plate interiors. These far - field tectonic effects appear to be common during recent basin formation but may have been overlooked in the past since many Precambrian basins developed substantial distances away from obvious zones of loading i.e. Witwatersrand basin. Another consequence of these observations is the superimposed effects of compressive tectonics on extensional basins, which was observed by Ziegler, (1987). This led to the introduction of inversion tectonics where the shift from extension to compressional represented the inversion process. This phenomena is expressed as renewed basin subsidence in response to the reversed stress regime. Usually this inversion effect is likely to be preserved in sedimentary basins with weakened zones, prone to amplify the tectonic stress change by preserving a thicker sedimentary sequence (Klein, 1991b). If an in - plane stress is applied to intracratonic basins, passive margins or foreland basins, it exploits original deformations of the lithosphere, such as the Moho, the sediment/basement interface or any other rheological boundary (Allen and Allen, 1990). Many features of the tectonic evolution of the Witwatersrand could easily be related to such conditions.

The effects of changes of in - plane stress are better preserved along the basin margins where the changes are not swamped by rapid subsidence rates. In - plane compression causes foreland basins, to have amplified basin subsidence rates and uplift along the margin, whereas in - plane tension causes basin uplift and marginal subsidence (Allen and Allen, 1990) (Figure 2.7). This situation is clearly illustrated and observed in the tectonic and total subsidence curve generated for the Witwatersrand basin (Figure 5.9).

The effects of inversion can be observed in the tectonic subsidence curves (Figure 2.7). Cloetingh (1988) interpreted the increase in slope angle as compressional events following thermal subsidence. These compressional events are short term events which disrupt the overall subsidence trend. Similar short term events disrupting the general long term subsidence trend are clearly illustrated in the Witwatersrand basin subsidence curve (Figure 5.9). Other interpretations suggest that these changes in the slope of the subsidence curve are represented by a sudden drop in sea - level followed by a rise in sea level with sediment accumulation rates following the sudden sea - level rise. Watso and Klein (1989) interpret these changes to represent repeated mechanical extensional events coeval with thermal subsidence (Klein, 1991b). Inversion tectonics could well have played a significant role in the early and late stages of the tectonic evolution of the Witwatersrand basin.

2.3.2 The effects of intraplate stress on the stratigraphy of basins

Changing plate stress has certain implications for the preservation of the eustatic record in a sedimentary basin. One of these implications is that with no plate stresses and subsidence being thermal, onlap signatures will develop and through time, sediment accumulation rates will decrease. A second alternative is that if one increases the extensional stress, basin margin subsidence will occur, promoting a stratigraphic onlap, with the rates of sediment accumulation increasing as accommodation space increases. In this scenario the time stratigraphic boundaries become further apart (Klein, 1991b).

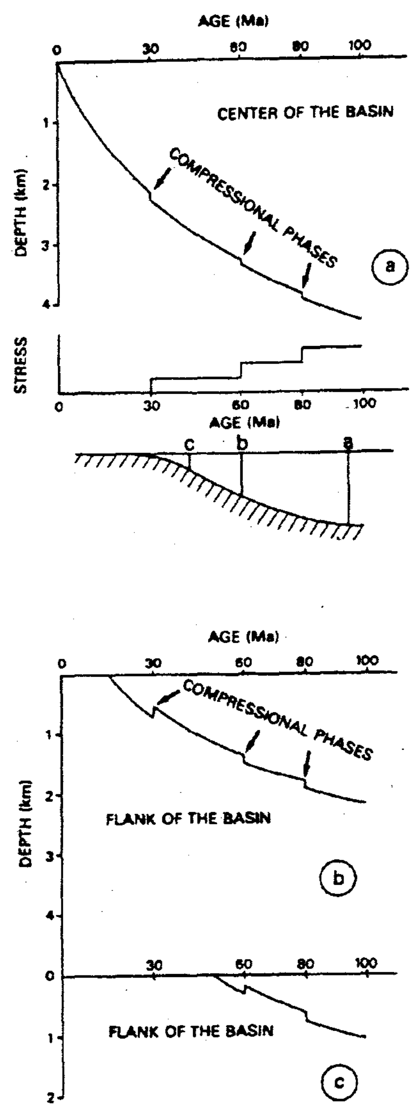


Figure 2.7 - Effects of intraplate stress on tectonic subsidence curves on an evolving sedimentary basin. **A.** Basin centre. **B.** Lower curves show effect of compressional stress on subsidence on the flank, and **C.** Closer to the flexural node of basin (Cloetingh,1988) (From Klein, 1991b).

A compressional stress on the other hand will result in basin margin uplift, and a corresponding relative sea - level drop, causing the formation of unconformities on the basin flank, which may be traced laterally into a correlative conformity or a drop in the amount of stratigraphic onlap. Usually these events are temporary and as soon as stress is restored to thermal subsidence, onlaps will develop again (Klein, 1991b). It is important to be able to recognise the effects of intraplate stress on basin stratigraphy because one or more of these scenarios may have occurred in the Witwatersrand basin.

Cloetingh (1986, 1988) argued that the Vail et al. (1977) onlap curve, can be used as a stressmeter just as much as it is an indicator of eustatic sea - level changes, because changes in tectonic stress can also occur in short - term recurring intervals (Klein, 1991b).

2.3.3 Effects of flexure on stratigraphy in basins due to stretching

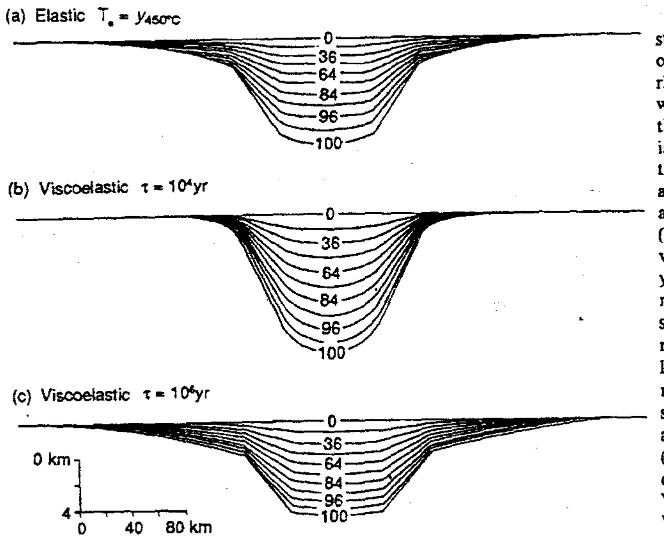
Basins such as rifts and passive margins initially become stretched, causing fault - controlled subsidence with the lithosphere in a state of purely local (Airy) isostasy followed by thermal subsidence caused by the cooling of the asthenosphere. The magnitude of the stretching event determines the initial depth and size of the basin, whereas during the thermal subsidence event, flexuring of the lithosphere becomes the determining factor influencing the depositional sequences during the post - rifting stages. There are two end members to flexural lithospheric responses, i.e. elastic or viscoelastic response (Figure 2.8). Therefore one can expect two corresponding stratigraphic styles. If the lithosphere responds elastically, the basin stratigraphy will reflect a progressive overstepping of the younger strata at the margin of the basin (onlap), whereas a lithosphere which has responded viscoelastically will reflect a progressive basinward shift of the depositional limit (offlap) with the youngest sediments restricted to the basin centre (Allen and Allen, 1990). These aspects are relevant during the early stages of basin development of the West Rand Group.

2.3.4 The role of flexure in generating foreland basins

Since a cratonic foreland basin model for the Witwatersrand widely accepted. It is appropriate to discuss the role of flexure in generating foreland basins. A foreland basin has a typical wedge - shape geometry, thickest close to the orogenic load and thinning onto the foreland in a "feather edge". It is the movement of the orogenic load which is responsible for the onlap onto the flexed plate with a time constant flexural rigidity (Figure 2.9). The flexuring of the lithosphere plays an important role in controlling the stratigraphy of the foreland basin, which is evident from the very existence of the basin. The secular evolution of the flexural rigidity of the basin plate is difficult to assess (Allen and Allen, 1990).

It becomes vital to consider the Witwatersrand foreland basin stratigraphy in conjunction with the previous geological history of the lithosphere. This involves the lifespan of a Wilson cycle, which consists of a cycle of rifting, drifting, subduction and collision, implying that the foreland basin could be superimposed on an inherited passive margin basin. This implies that the lithosphere should possess a rigidity, reflecting its previous history of heating and thinning. It should also reflect its first orogenic loads emplaced on a pre - existing oceanic bathymetry. These two aspects allow extremely thick overthrust thrust wedges to develop on stronger unstretched lithosphere with little topographic expression.

Two other features that become prominent during the formation of foreland basins, are the passage of the flexural forebulge, (which causes complex unconformities to develop) and progressive overthrusting of the plate, on the other hand, causes the forebulge and depocentres to migrate (Figure 2.9). Progressive uplift on the forebulge causes unconformities to merge and the intersection of the merged unconformities tends to migrate systematically towards the orogen (ie. Appalachain foreland basin systems). This is a clear indication of the contemporaneous uplift and migration of the forebulge toward the orogen and sedimentologically these effects are recorded as shoaling upwards sequences on the forebulge. The rate of the migration of the depocentre and pinch - outs on the forebulge gives an impression of the mobility of the distributed load and variations in the lithospheric response. The orogenic front causes the uplift of the



Calculated stratigraphy of stretched basins with $\beta = 2.0$, overlying lithospheres of contrasting rheology. (a) *Elastic* lithosphere where the equivalent elastic thickness is the depth to the 450 °C isotherm. The equivalent elastic thickness increases with thermal age, causing stratigraphic onlap and a steers-head geometry to the basin. (b) *Viscoelastic* lithosphere with a viscous relaxation time of 10^4 years. The lithosphere weakens rapidly following loading, causing stratigraphic offlap and a gradually narrowing basin. (c) *Viscoelastic* lithosphere with a viscous relaxation time of 10^6 years. The slower viscous relaxation produces a basin geometry similar to that in (a), but differs in being characterized by stratigraphic offlap. Vertical exaggeration of × 10. After Watts, Karner and Steckler (1982).

Figure 2.8 - Elasitic flexure vs viscoelastic flexure of the lithosphere (from Allen and Allen, 1990).

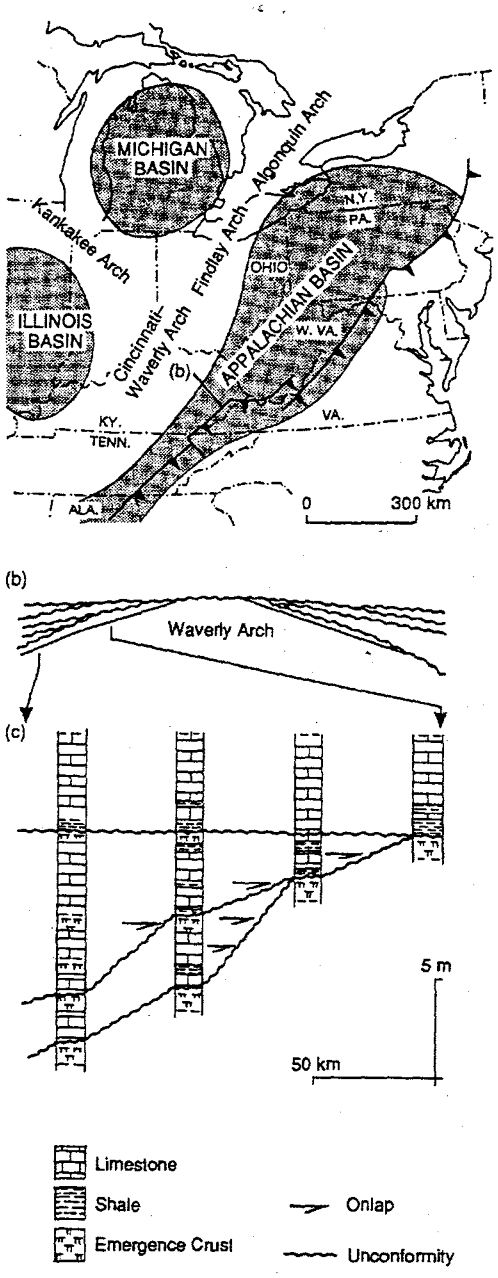


Figure 2.9 - The peripheral arches of the Appalachian Foreland basin system (Tankard, 1986). (a) Palaeozoic thrusting and flexure produced a system of swells or arches seperating the Appalachian foreland basin from the Michigan and Illinois basins of th American interior. (b) Stratigraphic relations during late Acadian (Early Carboniferous or Mississippian) across the Cincinatti - Waverly arches in Kentuky (Ettensohn, 1981), showing wedges seperated by unconformities fanning out from the crest of the arch. (c) Stratigraphic columns on the western edge of the arch show merging unconformities and evidence for periodic shoaling interpreted to be caused by forebulge uplift (Allen and Allen, 1990).

older stratigraphy, which becomes eroded and cannibalized to provide detritus to fill the foreland basin (Allen and Allen 1990). So far few facts are available about the actual development of unconformities along the forebulge of the Witwatersrand foreland basin, which is expected to be present along the southeastern margin of the basin, due to the lack of surface exposure caused by younger cover rocks. Therefore the interpretation is dependant on expensive exploration deep diamond drilling information, which is not available for public scrutiny.

3 BASIN SEDIMENTOLOGY AND STRATIGRAPHY - BASIN FILL

3.1 Sequence Stratigraphy and Eustatic Changes in Sea - Level

Sequence stratigraphy is an outgrowth of seismic stratigraphy and has since become a multidisciplinary approach to stratigraphic analysis, and therefore to basin analysis. It is based on information obtainable from well - logs, borehole core, outcrop data and seismic profiles. The concepts of sequence stratigraphy have stimulated an extraordinary amount of research/debate and revitalised stratigraphic analysis to such an extent, that sedimentary environments and facies are now being discussed as components of parasequences and sequences, which in turn form the building blocks of systems tracts and depositional systems (Walker, 1992).

The recognition of unconformities and bounding discontinuities caused by major changes in sea - level has become an important tool in allowing the subdivision of the geological record. This can be done either by purely descriptive *allostratigraphic* units or by interpretative *sequences* and *parasequences* of sequence stratigraphy. The most widely accepted discontinuities with stratigraphic significance are:

- regressive surfaces of erosion
- transgressive surfaces of erosion
- and maximum flooding surfaces.

3.1.1 Principles and definitions of depositional sequences

The sequence is the most fundamental unit of sequence stratigraphy and can be defined as " *a relatively conformable succession of genetically related strata bounded at its top and base by unconformities and their correlative conformities (Vail et al. 1977). It is composed of a succession of systems tracts and is interpreted to be deposited between eustatic - fall inflection points*" (Posamentier et. al., 1988).

Before the basic principles of sequence stratigraphy can be reviewed, it is necessary to understand the fundamental sedimentological principles that form the foundation of a depositional *sequence*, since this also forms the basis of sequence stratigraphic analysis. The term *facies sequence* in a "non - sequence stratigraphic context", implies that certain facies properties (e.g. abundance of sand, grain size, or sedimentary structures) change progressively in a specific direction, be it vertical or lateral. In a vertical succession, a gradual transition from one facies to the other implies that the two facies environments were once adjacent laterally. If the facies associations are bounded by surfaces of erosion or a hiatus indicating non - deposition, either at the top or the bottom, it is impossible to determine if two vertically adjacent facies represent environments that were once adjacent laterally (Walker, 1992). Therefore, in terms of sequence stratigraphic studies, the term *sequence* has recently been given a very specific definition, and the term *facies successions* has replaced the older *facies sequence* term (Walker, 1992). It is important to realise that the contacts bounding a particular sequence have chronostratigraphic significance.

The ultimate goal in sequence analysis is to identify these stratigraphic units with relatively conformable successions of genetically related units, which are known as *depositional sequence*. These genetically related coherent packages of strata can facilitate local, regional and interregional correlations (Haq, 1991; Allen and Allen, 1990). It is important to realise that depositional sequences are formed by the interaction of tectonics, thermal history, sea - level changes and sediment supply.

The advent of high - quality seismic reflection data which allowed Vail et al., (1977) to recognise some of these depositional sequence boundaries, contact relationships and the variety of geometrical relationships to the depositional boundary. Distinct boundary classes of sequence can be defined based on the upper and lower boundary relationships (Figure 3.1). Depositional sequence boundaries are used for correlation and definition. The angularity of the strata above and below the unconformity are important in defining terms such as nonconformity, disconformity and paraconformity. More important to depositional sequence analysis is the relation of the strata to the unconformity itself.

Table 1.2

Glossary Of Terms In This Chapter And Throughout The Dissertation (From Walker, 1992)

Allostratigraphy — subdivision of the stratigraphic record into mappable rock bodies "defined and identified on the basis of their bounding discontinuities" (NACSN, 1983, p. 865).

Architectural Element — a morphological subdivision of a particular depositional system characterized by a distinctive assemblage of facies, facies geometries, and depositional processes.

Bounding Discontinuity — a laterally traceable discontinuity; can be an unconformity, ravinement surface, onlap or downlap surface, condensed horizon or hardground.

Depositional Environment — geographic and/or geomorphic area

Depositional System — "three dimensional assemblage of lithofacies, genetically linked by active or inferred processes and environments" (Posamentier *et al.*, 1988, p. 110). It embraces depositional environments and the processes acting therein.

Downlap — the situation where "an initially inclined layer terminates downdip against an initially horizontal or inclined surface" (Mitchum *et al.*, 1977, p. 58).

Eustasy — a world-wide change of sea level relative to a fixed point such as the centre of the earth. Eustatic changes result from variations in the volume of water in the ocean basins (glacial control), or a change in the volume of the basins themselves (related to rates of ocean ridge building and rates of seafloor spreading). The eustatic sea level curve describes cyclic changes in sea level.

Facies — a body of rock characterized by a particular combination of lithology, physical and biological structures that bestow an aspect ("facies") different from the bodies of rock above, below and laterally adjacent.

Facies Association — "groups of facies genetically related to one another and which have some environmental significance" (Collinson, 1969, p. 207).

Facies Succession — a vertical succession of facies characterized by a progressive change in one or more parameters, e.g., abundance of sand, grain size, or sedimentary structures

Facies Model — a general summary of a particular depositional system, involving many individual examples from recent sediments and ancient rocks.

Genetic Stratigraphic Sequence — "the sedimentary product of a depositional episode" (Galloway, 1989, p. 125), where a depositional episode "is bounded by stratal surfaces that reflect major reorganizations in basin paleogeographic framework" (Galloway, 1989, p. 128). These stratal surfaces are maximum flooding surfaces, not the unconformities used to define stratigraphic sequences.

Lithostratigraphy — "a defined body of sedimentary...strata which is distinguished and delimited on the basis of lithic characteristics and stratigraphic position" (NACSN, 1983). It is internally lithologically homogeneous.

Marine Flooding Surface — "a surface separating younger from older strata across which there is evidence of an abrupt increase in water depth" (Van Wagoner *et al.*, 1990, p. 8).

Maximum Flooding Surface — a surface separating a transgressive systems tract (below) from a highstand systems tract (above). It is commonly characterized by a condensed horizon reflecting very slow deposition; markers in the overlying systems tract downlap onto the MFS.

Onlap — the situation where "an initially horizontal stratum laps out against an initially inclined surface" (Mitchum *et al.*, 1977, p. 57-58).

Parasequence — "a relatively conformable succession of genetically related beds or bedsets bounded by marine flooding surfaces and their correlative surfaces" (Posamentier *et al.*, 1988, p. 110).

Ravinement Surface — an erosion surface produced during marine transgression of a formerly subaerial environment.

Seismic Stratigraphy — "a geological approach to the stratigraphic interpretation of seismic data" (Vail and Mitchum, 1977, p. 51).

Sequence — "a relatively conformable succession of genetically related strata bounded at its top and base by unconformities and their correlative conformities...it is composed of a succession of systems tracts and is interpreted to be deposited between eustatic-fall inflection points" (Posamentier *et al.*, 1988, p. 110).

Sequence Stratigraphy — "the study of rock relationships within a chronostratigraphic framework wherein the succession of rocks is cyclic and is composed of genetically related stratal units (sequences and systems tracts)" (Posamentier *et al.*, 1988, p. 110).

Systems Tract — "a linkage of contemporaneous depositional systems" (Posamentier *et al.*, 1988, p. 110).

Unconformity — "a surface separating younger from older strata, along which there is evidence of subaerial erosional truncation...or subaerial exposure, with a significant hiatus indicated" (Posamentier *et al.*, 1988, p. 110). This is an extremely restricted definition; Posamentier (personal communication, 1990) now accepts that the "evidence" may be inferred rather than real.

Two types of lapouts can be recognised along the *lower boundary*:

- An *onlap* occurs where "an initially horizontal stratum laps out against an inclined surface", or where "an initial inclined stratum laps out against a surface with a greater inclination" (Mitchum et al., 1977).
- A *downlap* occurs when "an initially inclined stratum terminates against an initial horizontal, irregular or inclined surface" (Mitchum et al., 1977). The general term baselap is used when it becomes impossible to discriminate between onlap and downlap (Figure 3.1)

Lapout against the *upper boundary* of the depositional sequence can be recognised as a toplap:-

- A toplap occurs when "an initially inclined strata, such as clinoforms and delta foresets, terminate updip at the depositional boundary" (Mitchum et al., 1977) (Figure 3.1).

Onlap, downlap and toplap are evidence of non - depositional hiatuses. Toplap may be associated with sediment bypassing the physiographic shelf - break. *Erosional truncation* occurs when the initially inclined stratum is terminated by erosion along the upper boundary and is evidence an erosional hiatus (Allen and Allen, 1990).

There are two additional geometric patterns that can be recognised i.e.:-

- an *offlap*, due to progressive basinwards migration of the stratigraphic units and
- a *basinwards shift* displaying a discrete basinwards movement instead of a progressive one (Figure 3.1).

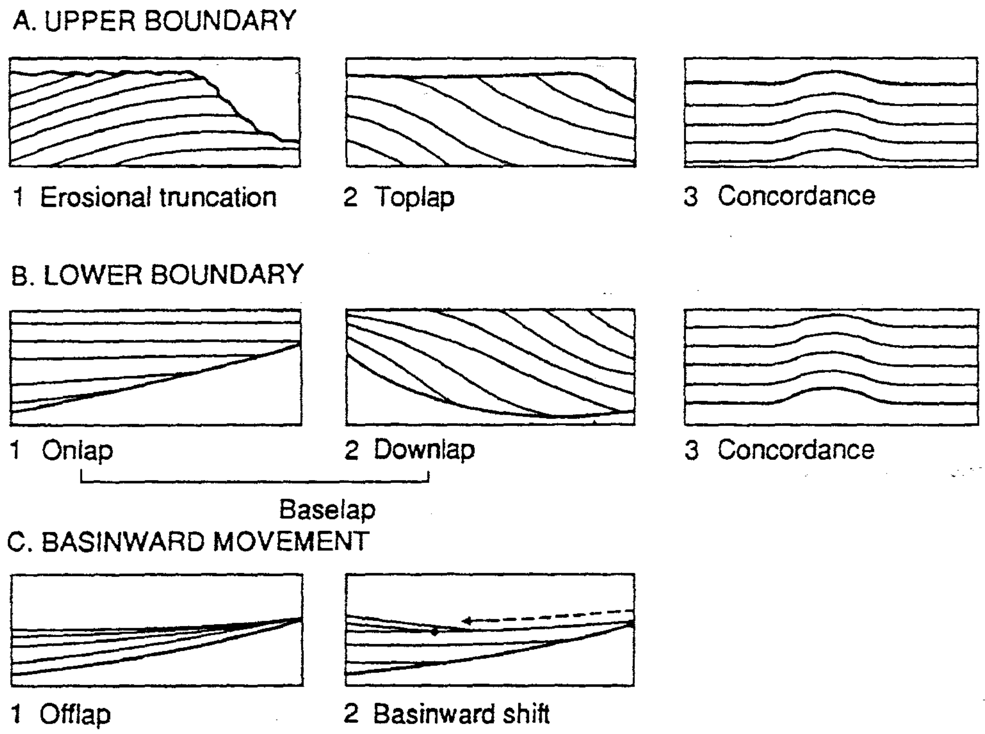


Figure 3.1 - The geometrical relationships of strata to a depositional sequence boundary or to any other surface within a depositional sequence. (A) Relations to upper surface, involving (1) erosional truncation, (2) toplap (commonly non - depositional rather than erosional) and (3) concordance. (B) relations to lower surface involving (1) onlap where the overlying strata are near - horizontal and the surface is inclined, (2) downlap where the overlying strata are inclined and (3) concordance. (C) Additional geometrical patterns of (1) offlap, where there is progressive basinward migration of the stratigraphic units and (2) basinward shifts, where the basinward movement is discrete rather than progressive (From Allen and Allen, 1990).

3.1.2 An Overview Of The Fundamentals Of Sequence Stratigraphy

A sequence can consist of several phases of sea - level cycles, known as systems tracts i.e. lowstand, transgression, highstand, and regression, which produce a distinct package of genetically related sediments. A system tract can be defined as the " *linkage of contemporaneous depositional systems (Brown and Fisher, 1977). Each is defined objectively by stratal geometries at bounding surfaces, position within the sequence, and internal parasequence stacking pattern*" (Posamentier et al., 1988). Each of these systems tracts is associated with a specific segment of the eustatic curve i.e. eustatic low - lowstand wedge; eustatic rise - transgressive; rapid eustatic fall - lowstand fan etc. and display distinct collection of lithofacies.

The basic building blocks of systems tracts are *parasequences*. A typical definition of parasequences "*are relative conformable successions of genetically related beds or bedsets bounded by marine flooding surfaces and their correlative surfaces*" (Van Wagoner, 1988). Van Wagoner (1988) introduced the term "parasequence" as a rock - based descriptive term and therefore it has no significance with respect to temporal and spatial relations. Facies successions and parasequences are synonymous, however the facies succession concept is broader and a parasequence is defined by a marine flooding surface. For instance, a fluvial - point bar consists of an upward - fining facies succession, defined by an erosion surface with channel lag deposits and therefore can not be regarded as a parasequence because it is not bounded by a marine flooding surface (Walker, 1992). Parasequences represent higher frequency flooding events that occur in all system tracts (Haq, 1991). Parasequences strictly describe a shoaling - upward succession bounded by marine flooding surfaces (Posamentier and James, 1993).

In order to interpret and analyse a parasequence one has to recognise the following:

- recognise the shoaling - upward nature of the section (i.e. identification of parasequences)

- followed by recognition and interpretation of key bounding surfaces (e.g. ravinement surfaces, maximum flooding surfaces, unconformities etc.), as well as condensed sections
- recognition of stratigraphic relationships and identification of systems tracts by means of parasequence stacking patterns and finally sequences.

Genetically related parasequences can be arranged into distinctive stacking patterns to define a parasequence set, bounded by a major marine - flooding surfaces and their correlative surfaces (Van Wagoner et al., 1988) (Figure 3.2). These parasequences and parasequence sets define the various systems tracts which can either consist of:-

- progradationally stacked parasequences or *a progradational parasequence set* (characterising the late highstand and early lowstand systems tracts)
- aggradationally stacked parasequences or an *aggradational parasequence set* (characterising the early highstand and late lowstand systems tracts)
- and retrogradationally stacked parasequences or a *retrogradational parasequence set* (characterising the transgressive systems tract).

Each systems tract can be interpreted to be deposited during specific increments of the eustatic curve (Posamentier et al., 1988):

- lowstand fan of lowstand systems tract - during a time of rapid eustatic fall
- slope fan of lowstand systems tract - during the late eustatic fall or early eustatic rise
- lowstand wedge of lowstand systems tract - during the eustatic fall or early rise
- transgressive systems tract - during a rapid eustatic rise highstand systems tract - during the late part of a eustatic rise, a eustatic stillstand and the early part of a eustatic fall.

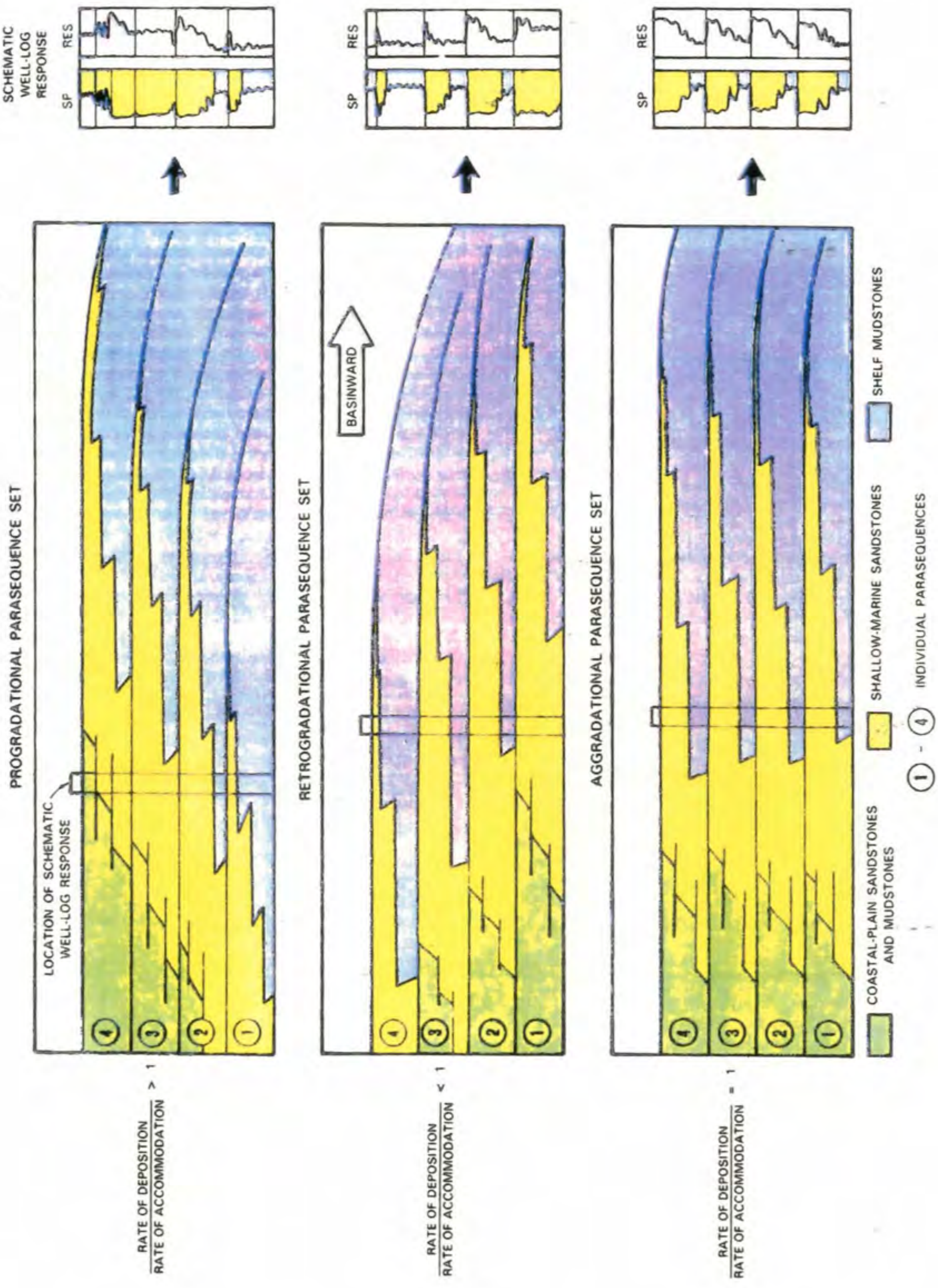


Figure 3.2 - Stacking patterns of parasequences in parasequence sets are progradational, retrogradational and aggradational, depending on the ratio of depositional rates to accommodation rates (From Van Wagoner et al., 1988).

It is important to recognise these stacking patterns in the Witwatersrand succession since they are predictable and provide information regarding the distribution and depofacies of the basin fill. Furthermore, they provide information as to the rate of deposition and rate of accommodation (Van Wagoner et al. 1988). The subdivision of sedimentary strata into sequences, parasequences and systems tracts provides a powerful tool for the analysis of time and rock relationships in sedimentary strata. Sequences are bounded by unconformities and their correlative conformities (sequence boundaries) provide a chronostratigraphic framework for correlation and predicting facies relationships within a sequence (Van Wagoner et al., 1988).

Before any *predictions* can be made regarding the succession of as many as four systems tracts, the systems tracts have to be refined by incorporating subsidence and sediment supply (Posamentier et al., 1988). In order to develop a generally applicable model, the concept of sea - level evolution i.e. *accommodation space* needs to be incorporated (Figure 3.3). Accommodation can be defined as the space made available for potential sediment accumulation and is expressed in two dimensions as available water depth and is a function of sea - level fluctuations and subsidence (Jervey, 1988).

Whether this accommodation space is filled or underfilled, becomes a function of environmental processes that allow sediment accumulation and accumulation rates. Critical assumptions are made by Posametier et al. (1988), regarding the following conditions and they might not all be applicable to Witwatersrand basin conditions:-

- at any single point on the basement a constant rate of subsidence exists. This does not hold true for all situations, but the general model can be modified to account for local conditions.
- differential subsidence occurs across most divergent basins, from slow in the centre to fast on the margins
- sediment supply remains constant
- eustatic changes tends to be curvilinear, approaching sinusoidal

eustatic changes are more frequent than that for tectonic activity (Posamentier et al., 1988).

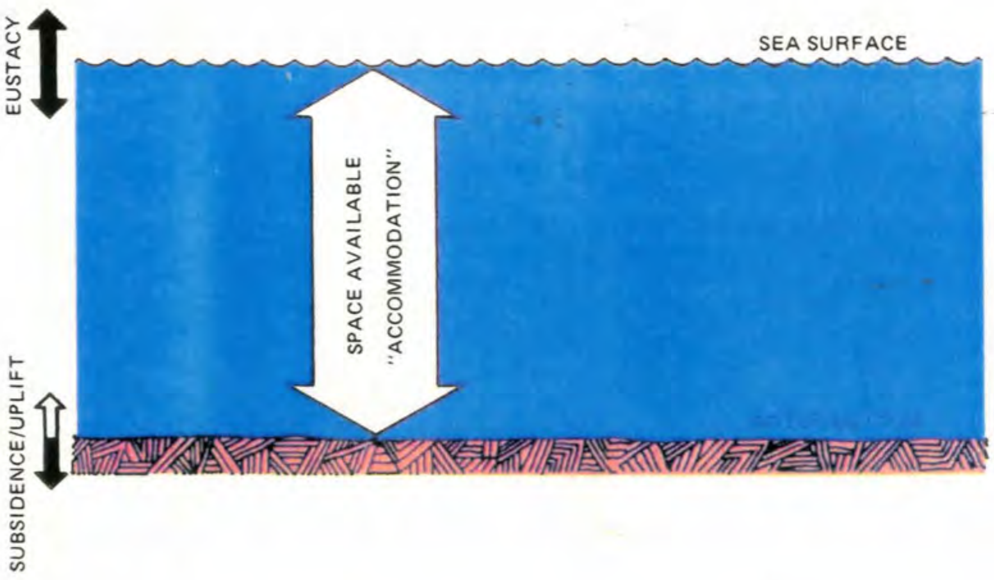


Figure 3.3 - Accommodation envelope as a function of eustasy and subsidence (From Posamentier et al., 1988).

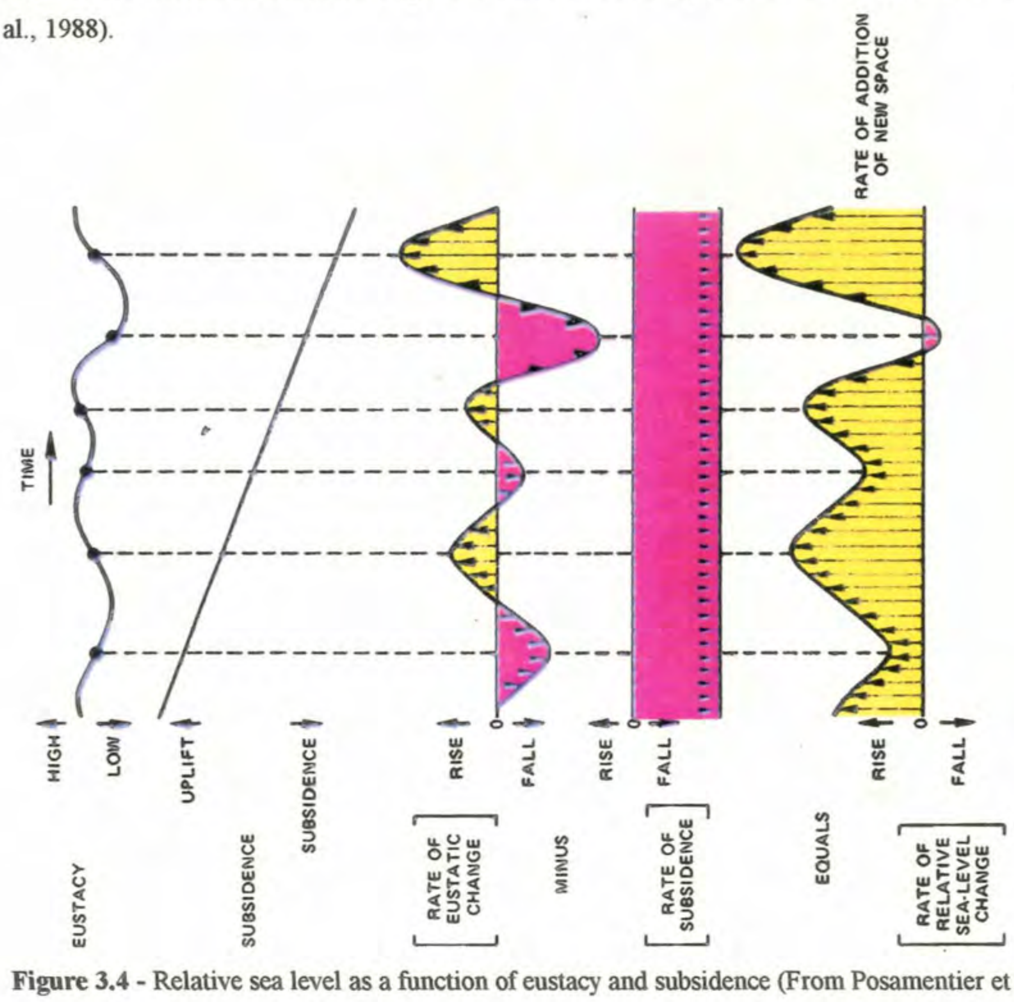


Figure 3.4 - Relative sea level as a function of eustacy and subsidence (From Posamentier et al. 1988)

A curve can be generated by adding the sinusoidal curve of eustasy to the subsidence, which is assumed to be constant, representing the total potential change in accommodation space (Figure 3.4). A relative sea - level curve is obtained by subtracting the constant subsidence from the aforementioned curve, which tends to amplify the sea - level signature in the residual curve for eustasy (Klein, 1991c). It is more convenient to refer to relative sea - level changes, rather than eustatic changes to account for factors controlling accommodation changes. Eustasy refers to the position of the sea surface by referring to a fixed datum i.e. the centre of the earth, whereas relative sea - level incorporates local subsidence and/or uplift by referring to the sea surface with respect to the position of a datum (e.g. basement) (Posamentier et al., 1988) (Figure 3.5). Therefore, any variation due to thermal cooling, loading by sediments/water or tectonics will be manifested by the relative sea - level changes along the profile. A relative sea level rise will add space, whereas a relative sea - level fall will take the space a way. Relative sea - level changes are independent of sediment accumulation, because even during a eustatic stillstand or slow eustatic fall, relative sea level may continue to rise and add new space due to local subsidence. The significance of relative sea - level is that it describes how sediment accommodation varies with time. For instance, if relative sea level continue to rise, adding new space to accommodate sediment, water depth may continue to decrease simultaneously if the sediment accumulates faster than relative sea level is rising.

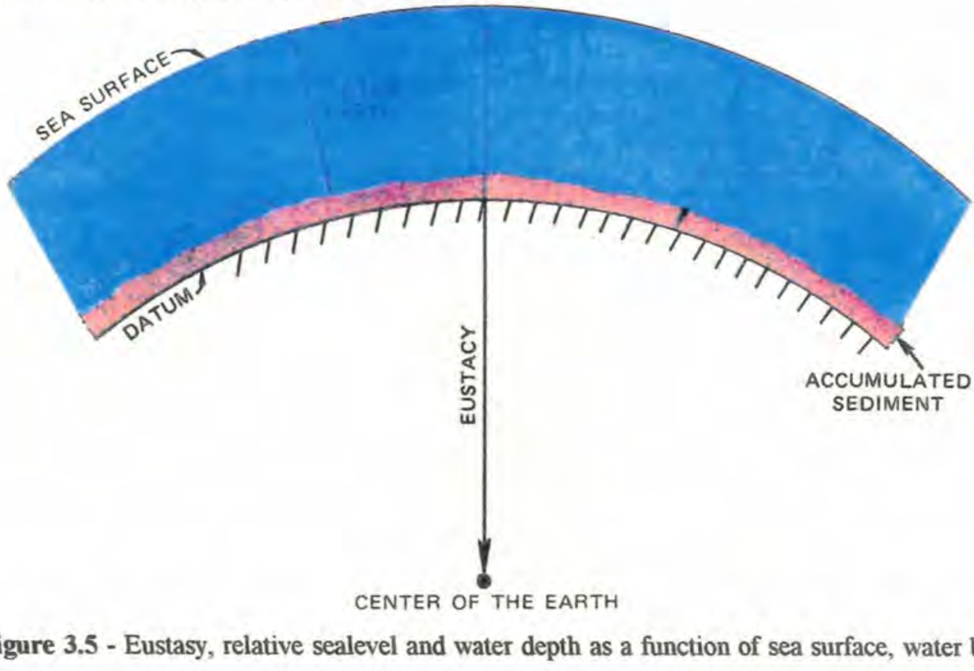


Figure 3.5 - Eustasy, relative sealevel and water depth as a function of sea surface, water bottom and datum position (From Posamentier, et al., 1988).

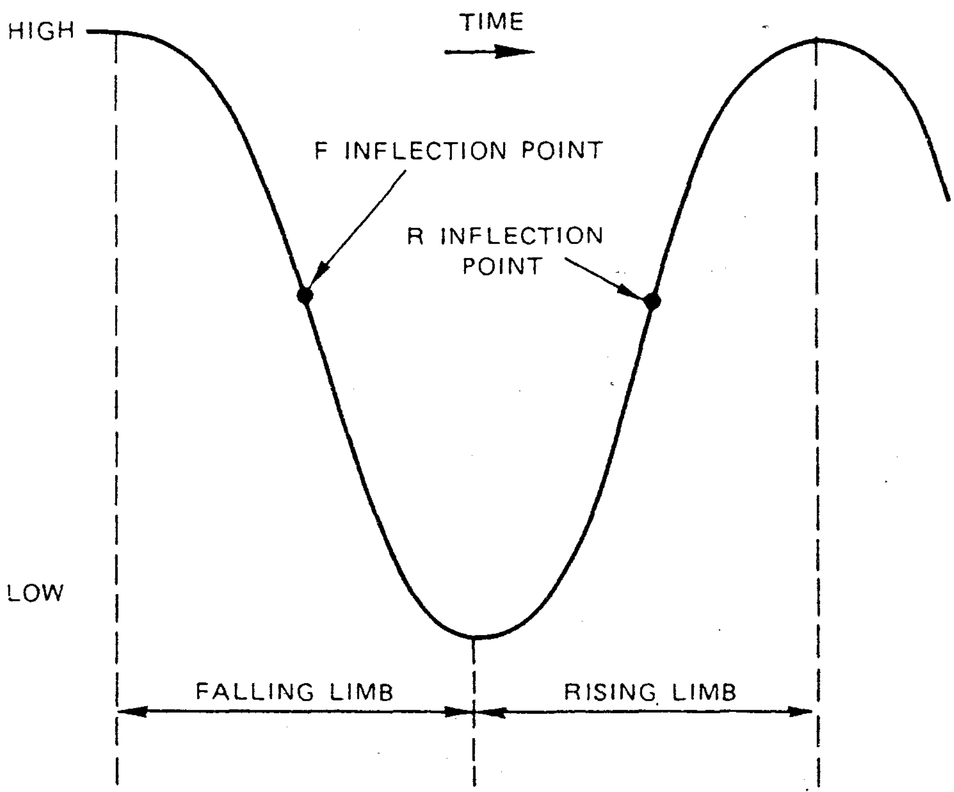


Figure 3.6 - Elements of eustatic change (From Posamentier et al. 1988)

Another term with similar meaning and usage, is *base level.* This is the interaction between sea - surface and basin floor movement. Base level can therefore be defined as the elevation of the point to which a fluvial system will be graded (Posamentier and James, 1993). It is therefore obvious that base level elevation has a profound influence on the position in vertical space of a graded or steady state or equilibrium profile and could have application to non - marine and continental settings. However, in this environment different factors will have an influence on the position and shape of the steady - state profile, other than relative sea - level changes. It is generally excepted that the effect of relative sea - level change on fluvial systems of the Witwatersrand sediments will become dampened in a upstream direction due to the increased relative importance of climatic changes as well as autocyclic changes (i.e. local tectonics, fluvial discharge in the upstream direction and sediment flux variations) (Posamentier and James, 1993).

Inflection points on the curve depict points where the rate of eustatic changes, is greatest (Figure 3.6). A hypothetical sea - level curve displays two inflection points, one on the falling limb, referred to as the F inflection point and one on the rising limb, referred to as the R inflection point (Posamentier et al., 1988). At the R inflection point, the greatest increase occurs when new space is available, causing an increase in the rate of aggradation, whereas the rate of progradation decreases. At the F inflection point, the opposite situation will occur. Thus, with a constant supply of sediment, progradation and aggradation are inversely related (Posamentier et al., 1988).

Each systems tract is associated with a specific segment of the sinusoidal eustatic sea - level curve. When each of the systems tracts are identified, it is critical to remember that each of them is correlated to a specific event on the sinusoidal eustatic sea - level curve. Figures 3.7 to 3.12 illustrates a succession of systems tracts models as proposed by Posamentier and Vail, (1988), interpreted to be deposited between two eustatic - fall inflection points, representing a complete sealevel cycle. Similar sea - level cycles can be identified within the Witwatersrand succession.

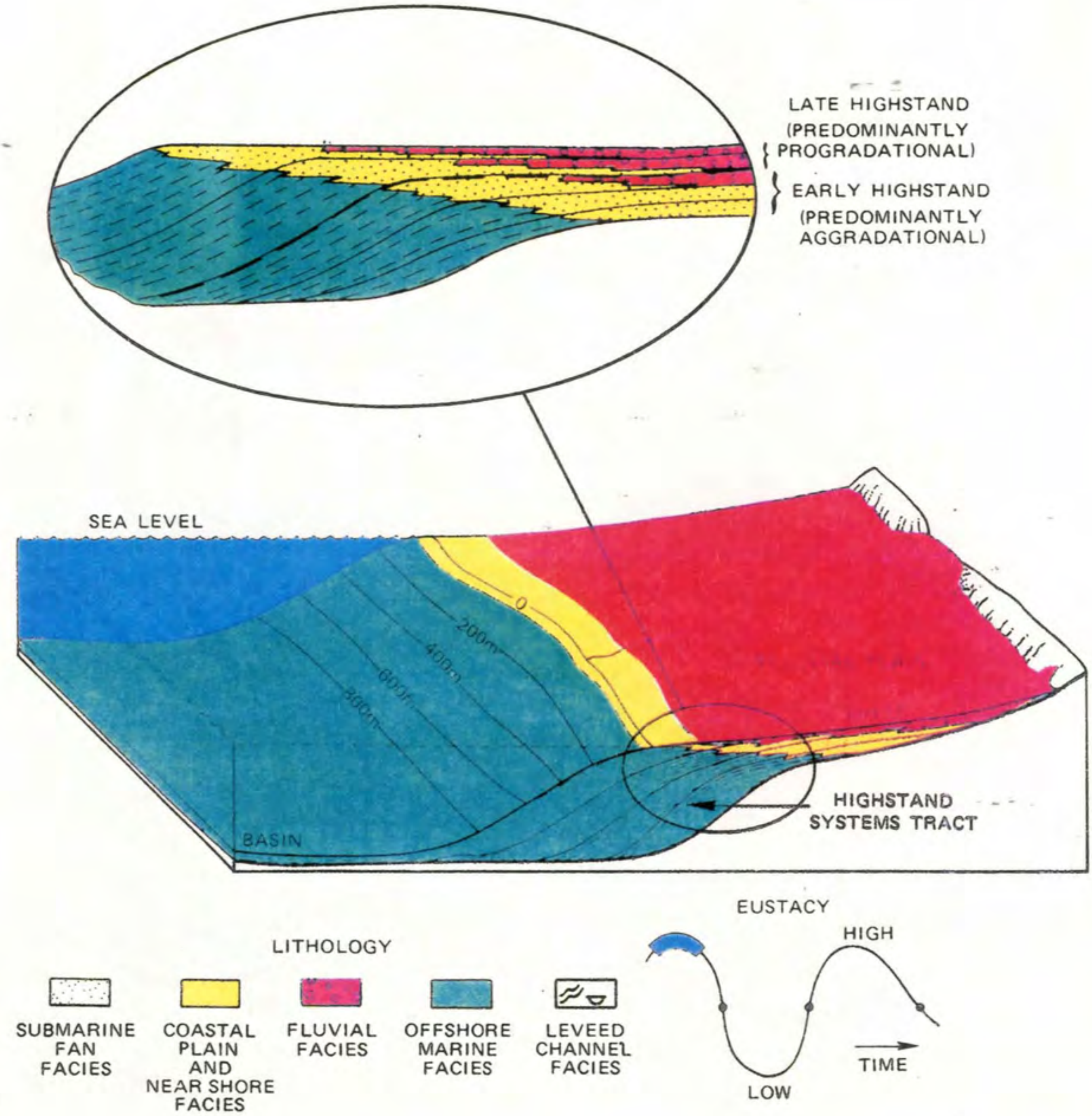


Figure 3.7 - Higstand systems tract , I (From Posamentier et al. 1988)

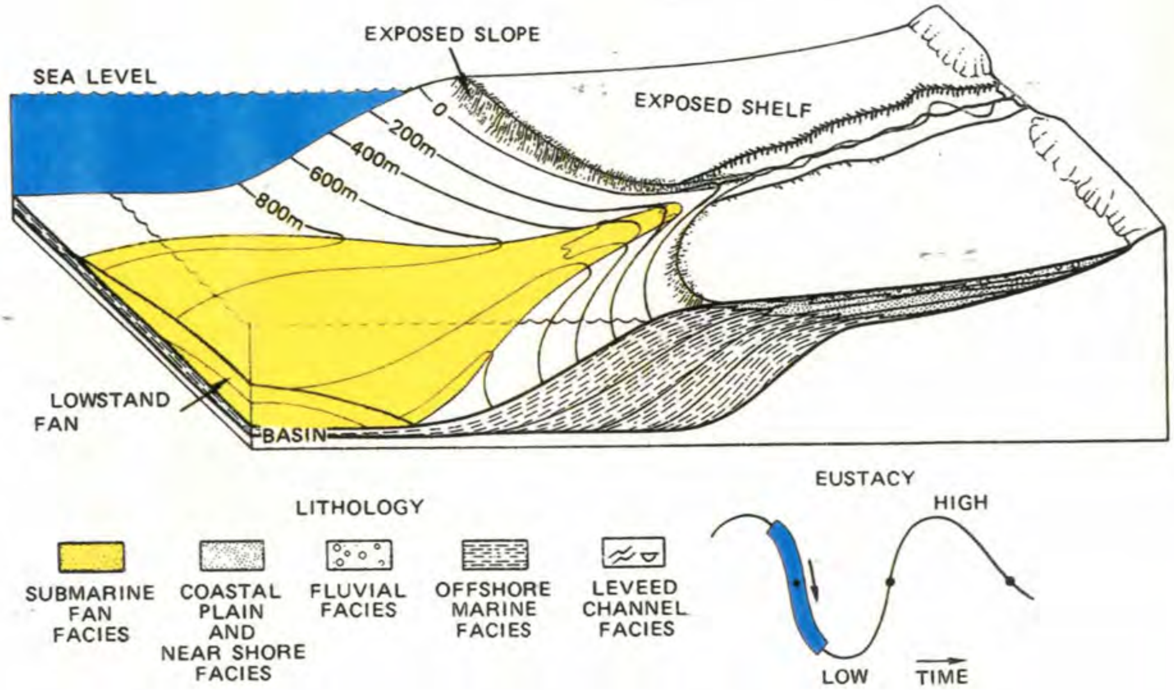


Figure 3.8 - Lowstand systems tract - lowstand fan (From Posamentier et al. 1988)

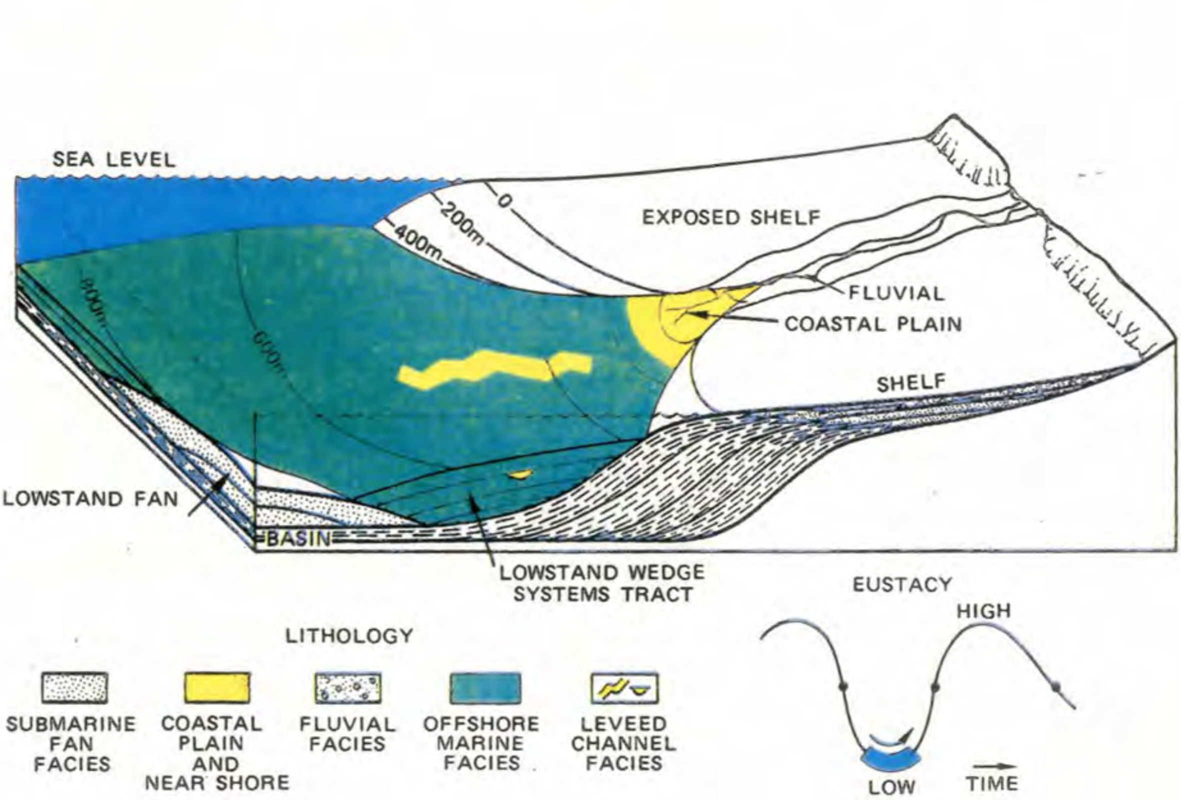


Figure 3.9 - Lowstand systems tract - lowstand wedge (From Posamentier et al., 1988)

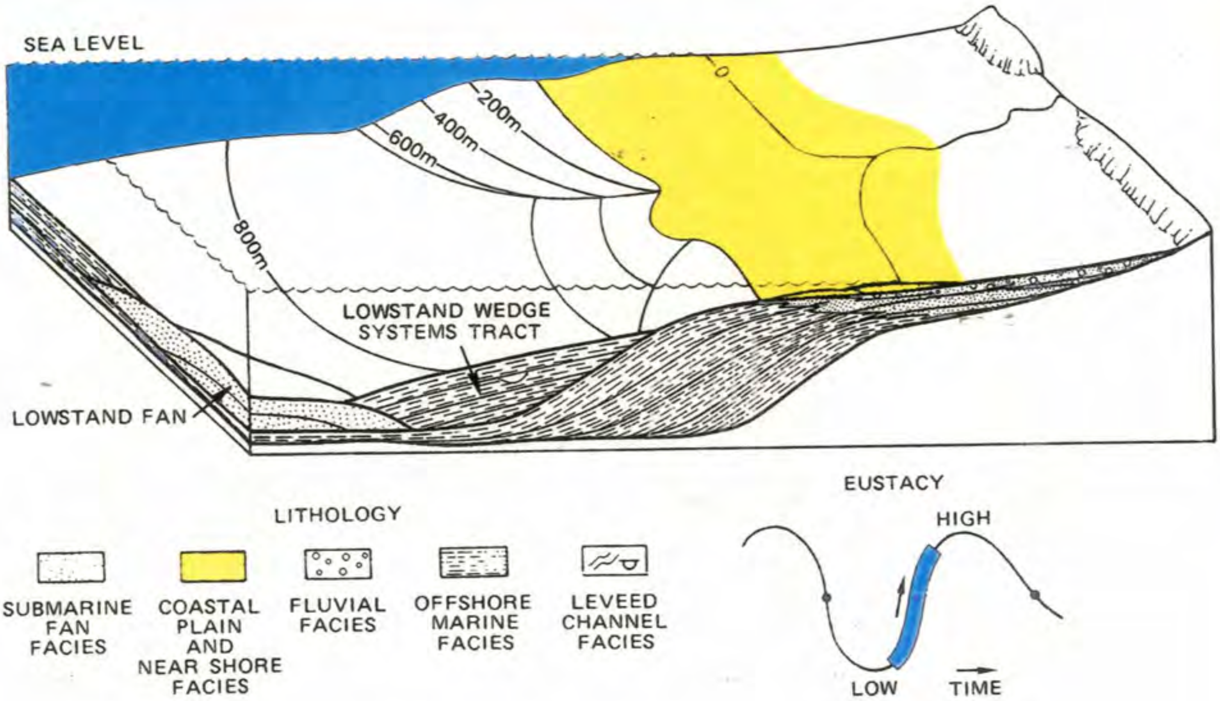


Figure 3.10 - Transgressive systems tract (From Posamentier et al. 1988)

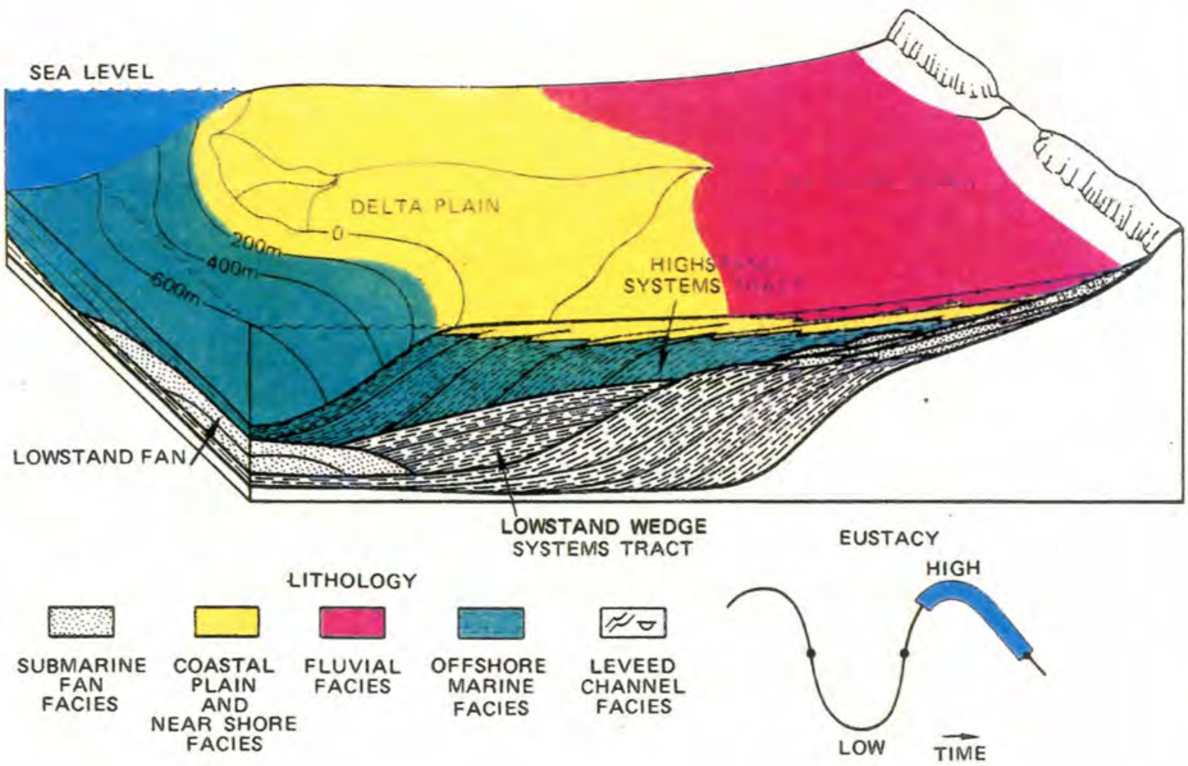


Figure 3.11 - Highstand systems tract, II (From Posamentier et al., 1988)

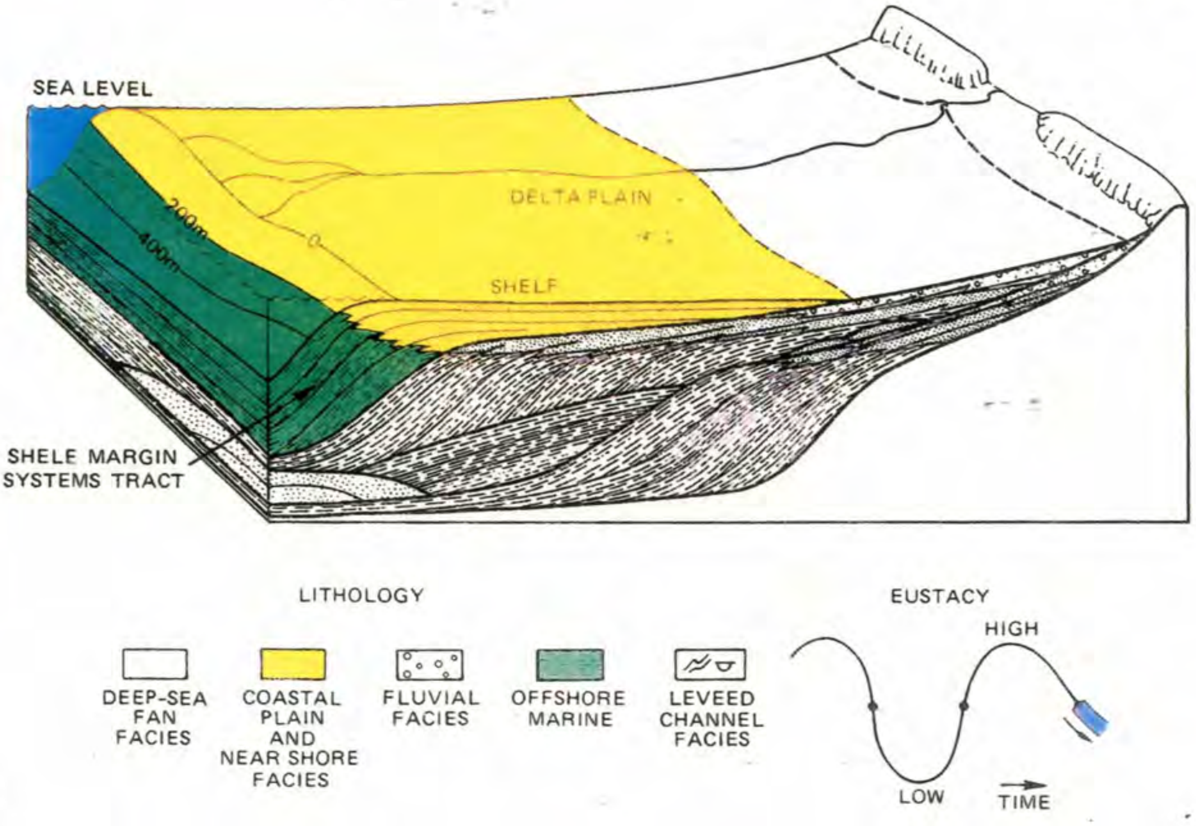


Figure 3.12 - Shelf - margin system tract (From Posamentier et al. 1988)

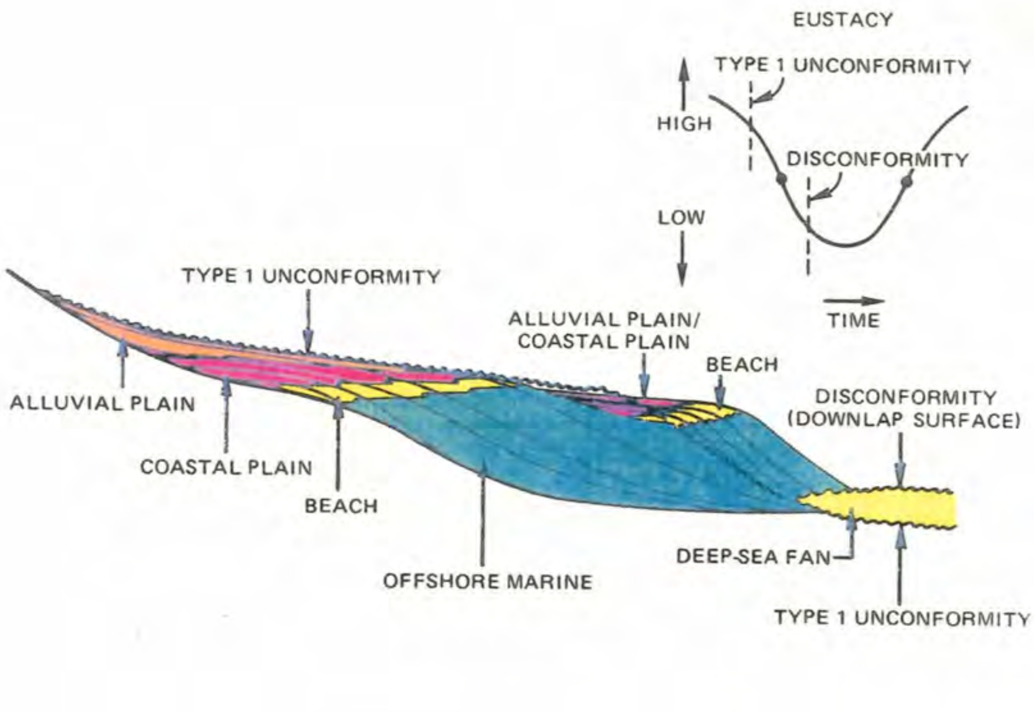


Figure 3.13 - Type I unconformity (From Posamentier et al., 1988)

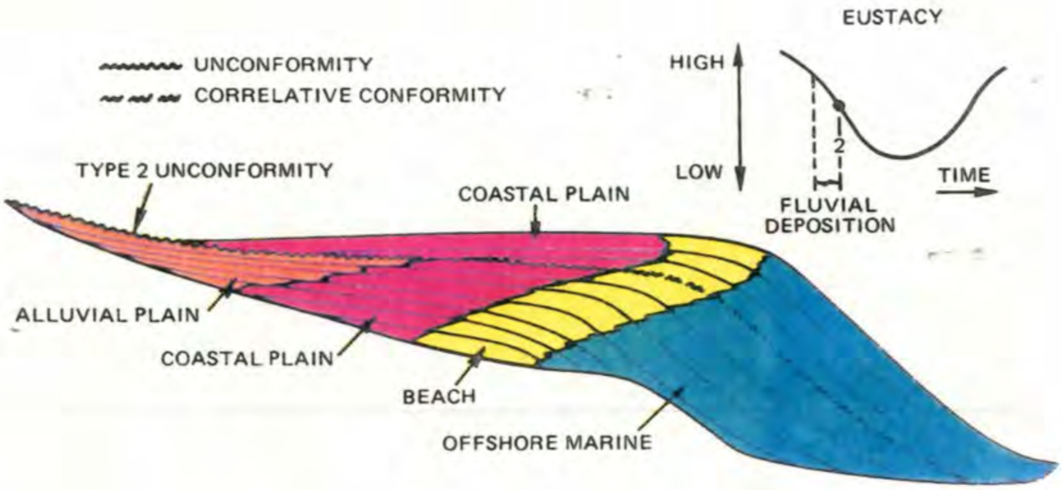


Figure 3.14 - Type 2 unconformity (From Posamentier et al. 1988)

Depending on the rate of the relative sea level fall one can recognise two types of sequences: Type 1 or Type 2. A Type 1 sequence consists of a lowstand, transgressive-, and a highstand systems tract, bounded by a Type 1 unconformity at its base and either a Type 1 or 2 unconformity at the top (Figure 3.13). A Type 2 sequence on the other hand is composed of a shelf margin-, transgressive -, and highstand systems tract, bounded by a Type 2 unconformity at the base and a Type 1 or 2 at the top of the sequence (Figure 3.14).

The *highstand systems tract* is initialised by a progradational stacking pattern, believed to have developed in response to a gradual slowing of relative sea - level rise after the R inflection point, with shoreline transgression giving way to regression (Figures 3.7 and 3.12). The actual timing of this event is dependent on the sediment supply, but usually occurs after the R inflection point but before the eustatic peak. The base of the systems tract is bounded by a condensed section and associated with downlap surfaces. Fluvial deposits will be deposited after the eustatic peak reached. The upper sequence boundary is characterized by either a Type 1 or Type 2 unconformity. These unconformities usually signify the cessation of fluvial deposition, when stream erosion becomes more gradual due to the gradual reduction of the stream equilibrium profile gradient (Posamentier and Vail, 1988).

Lowstands systems tracts are associated with a relative sea - level fall and subsequent slow rate of relative sea - level rise. If the systems tract occurs on a shelf with a distinct shelf edge it can be divided into a lowstand fan and a lowstand wedge which are not coeval (Figures 3.8 and 3.9). In this case the lowstand fan consists of submarine fan deposition, fed by sediments bypassing the shelf through actively incised valleys, followed by the lowstand wedge dominated by finer grained, wedge shaped slope - deposits.

If the lowstand occurs on a ramp margin without a discrete shelf edge, the systems tract will consist of a two part wedge. Initially the exposed shelf will be characterized by stream rejuvenation and sediment bypass with relatively course - grained basin - restricted wedge. This will be followed by a slow rise in relative sea - level, resulting in filling the incised valleys coupled with a slowed progradational shoreline replaced by an increased aggradation (Posamentier and Vail, 1988).

The *transgressive systems tract* is characterized by the development of a succession of backwardstepping or retrogradational parasequences during a rapid relative sea - level rise (Figure 3.10). The first major flooding event initiates the system tract after a period of maximum regression of the lowstand wedge systems tract. This system may be characterized by a succession of flooding events. Transgressive deposits associated with a Type 1 unconformity will be restricted to the incised valleys at first, whereas deposits associated with a Type 2 unconformity will be more widespread. The set of parasequences will change from being retrogradational to aggradational as soon as the rate of relative sea - level slows down. The surface at which this occurs is known as the *maximum flooding surface* (Posamentier and Vail, 1988).

The *shelf margin systems tract* is characterized by a decreasing progradational, followed by an aggradational parasequence stacking pattern, typical of a regressive stratigraphic unit (Figure 3.11). These sediments are generally characterized by vertical stacking of facies, gradually changing from a non - marine to marine environment. This systems tract occurs after the F inflection point during a progressive increase in the rate of relative sea - level fall. The lower boundary is marked by an erosional unconformity (or its correlative conformities) manifested by coastal plain or paralic/deltaic sediments overlying fluvial deposits. If the lower boundary is associated with a conformable contact, the only expression of this basal contact is the tendency of parasequence stacking patterns to change from rapid progradational to slowly progradational or aggradational. A transgressive surface marks the upper boundary separating the progradational - aggradational shelf margin systems tract from the overlying retrogradational transgressive systems tract (Posamentier and Vail, 1988).

The combination of information based on marine flooding and erosion surfaces allows the interpretation of the systems tracts and to construct a sealevel curve. This curve tends to reflect the concept that condensed sections (formation during starved shelf conditions and deposition of magnetic mudstones/iron formations) correspond with periods of maximum rate of sealevel rise and an erosional surface (sequence boundary Type 1 or 2) to periods of maximum rate of sealevel fall.

It must, however, be stressed that these models should not be applied to the Witwatersrand succession without taking local factors into consideration. Posamentier et al., (1988) specifically pointed out "that the models are generally applicable. The effects of local factors, such as tectonics, climate and variations in sediment supply must be incorporated into the models before these models can be applied".

3.2 Applications of Sequence Stratigraphy to Fluvial Strata

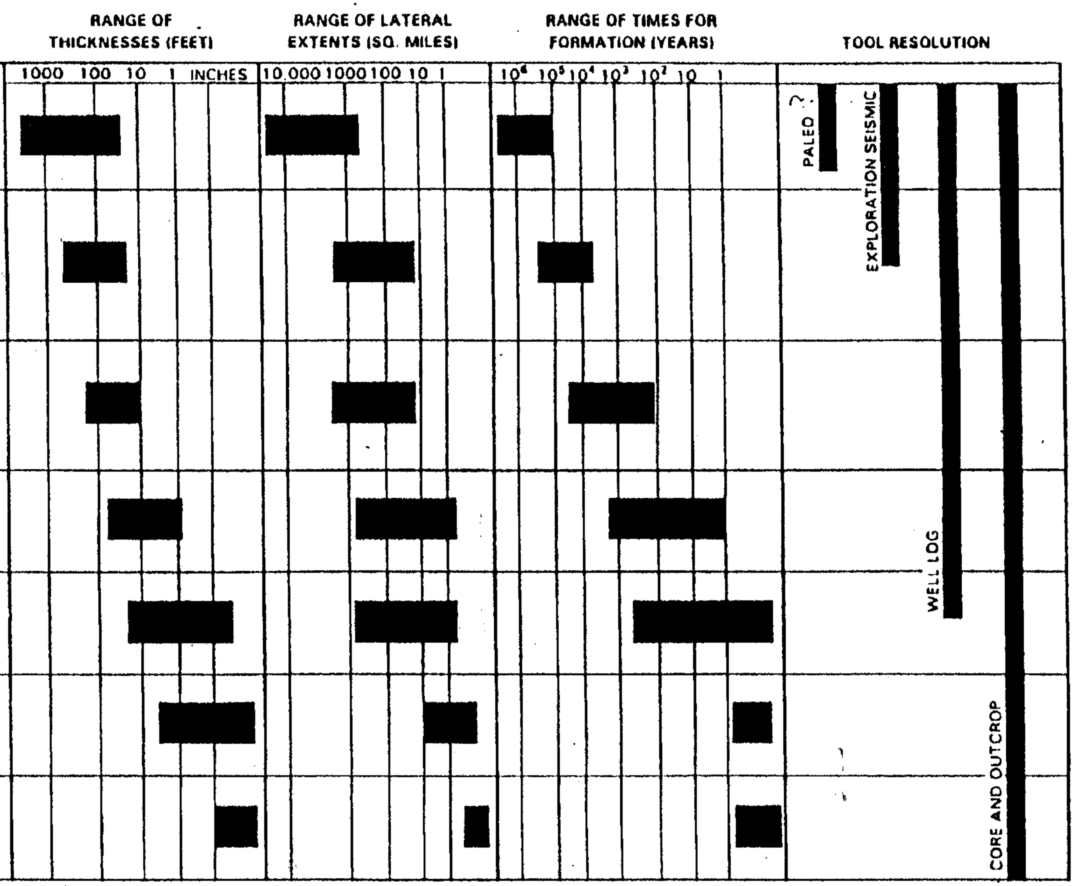
A prograding basin - margin sedimentary prism, such as the Witwatersrand basin, consists of sediments accumulating in a depositional system ranging from deep water slope and basin plain to paralic (deltaic, shore zone, and shelf) and terrestrial (fluvial and alluvial fan) (Galloway, 1989). Four bathymetric regimes can be recognised on such a prograding basin - margin i.e.:-

- slope
- shelf edge

Stratal Units in Hierarchy: Definitions and Characteristics

TABLE 1.3

STRATAL UNITS	DEFINITIONS
SEQUENCE	A RELATIVELY CONFORMABLE SUCCESSION OF GENETICALLY RELATED STRATA BOUNDED BY UNCONFORMITIES AND THEIR CORRELATIVE CONFORMITIES (MITCHUM AND OTHERS, 1977)
PARA-SEQUENCE SET	A SUCCESSION OF GENETICALLY RELATED PARASEQUENCES FORMING A DISTINCTIVE STACKING PATTERN AND COMMONLY BOUNDED BY MAJOR MARINE-FLOODING SURFACES AND THEIR CORRELATIVE SURFACES.
PARA-SEQUENCE	A RELATIVELY CONFORMABLE SUCCESSION OF GENETICALLY RELATED BEDS OR BEDSETS BOUNDED BY MARINE-FLOODING SURFACES AND THEIR CORRELATIVE SURFACES
BEDSET	SEE TABLE TWO
BED	SEE TABLE TWO
LAMINA-SET	SEE TABLE TWO
LAMINA	SEE TABLE TWO



- shelf and
- coastal plain.

Parasequences for deltaic, coastal plain and shelf are easily recognisable, whereas fluvial deposits are more difficult to recognise.

Identifying sequences is vital in diamond drill core and surface exposures are important for stratigraphic analysis of the Witwatersrand succession. The application of sequence stratigraphy depends entirely on the recognition of a hierarchy of stratal units including beds, bedsets, parasequences, parasequences sets and the surfaces bounding sequences. The latter have chronostratigraphic significance, (Van Wagoner et al., 1990). The hierarchy of stratal units is readily identifiable in geophysical borehole logs, diamond drill core and surface outcrops (Table 1.3). Bounding surfaces are correlatable because of their chronostratigraphic significance. Therefore they form the basis for the construction of a chronostratigraphic framework for facies analysis regardless of their relationship to change of eustasy (Van Wagoner et al., 1990).

Vertical facies analysis is concerned with prediction *within* conformable stratal packages by interpreting coeval, lateral facies relationships along a single depositional surface (Van Wagoner et al., 1990; Walker, 1992). Above and below these depositional surfaces facies are decoupled and therefore vertical facies analysis should be done *within* the context of parasequences, parasequence sets and sequences when lateral facies relationships are interpreted from one borehole to the other i.e. for 3D interpretation of the geometry of each stratal unit in a goldfield (Van Wagoner et al., 1990).

Geophysical borehole logs, diamond drill core and outcrops can be utilised to subdivide sequences into stratal units known as systems tracts based on their facies associations, positions within the sequence and stacking pattern of the parasequence sets. The parasequence set stacking patterns within the systems tracts, depending on whether they are progradational, aggradational or retrogradational, provides a high degree of facies predictability within the chronostratigraphic framework of sequence boundaries (Van Wagoner et al, 1990). It allows prediction from one depositional system to another e.g. a prograding storm - dominated shoreface suggests the possibly of an aggrading of storm - dominated shelf deposits i.e. Orange Grove Quartzite to Parktown shale. Facies models allows for prediction within the specified depositional system, whereas systems tracts allow prediction from one contemporaneous depositional system to another (Walker, 1992).

Boreholes logging should initially concentrate on recording the parasequences within the stratigraphy since they, provide the basic building blocks for the interpretation of systems tracts. Contemporaneity of systems tracts are usually established by identifying and correlating their bounding discontinuities, which are essentially time surfaces formed as a result of relative sea - level fluctuations. This allows for the recognition of the three systems tracts, known as highstand, lowstand and transgressive systems tracts (Walker, 1992). Parasequences can be composed of beds, bedsets, laminae and laminae sets (Van Wagoner et al., 1990) (Table 1.3). Parasequences are characterised by shoaling - upward successions and this forms the basis of their identification in geophysical borehole logs. A shoaling - upward sequence is characterised by an association of facies in which the younger bedsets are deposited in progressively shallower water. In borehole logs these shoaling upward sequences are readily identifiable as typical upward coarsening parasequence characterised by, bedsets thickening, the quartzite grains size coarsens and the quartzite/argillite ratio increase upward (Figures 3.15A, B & C). In upward fining sequences bedsets become thinner, grain size decreases and the quartzite/argillite ratio decreases (Figure 3.15D). Both these upward - coarsening and upward - fining parasequences indicate shoaling upward conditions (Van Wagoner et al., 1990).

In basin analysis the interpretative - predictive rationale involves a process - response approach. This means that sedimentary rocks owe their origin to physical and chemical processes that generate specific responses. Specific depositional processes will have diagnostic responses consisting of sedimentary structures, particle size variation and geometry of sedimentary bodies that are unique to a process and the environment in which it forms. The processes can include rivers, tides, and turbidity currents varying in response to sea level fluctuations. Geophysical log shapes accompanied by seismic reflection surfaces can contribute much needed data when interpreting diagnostic

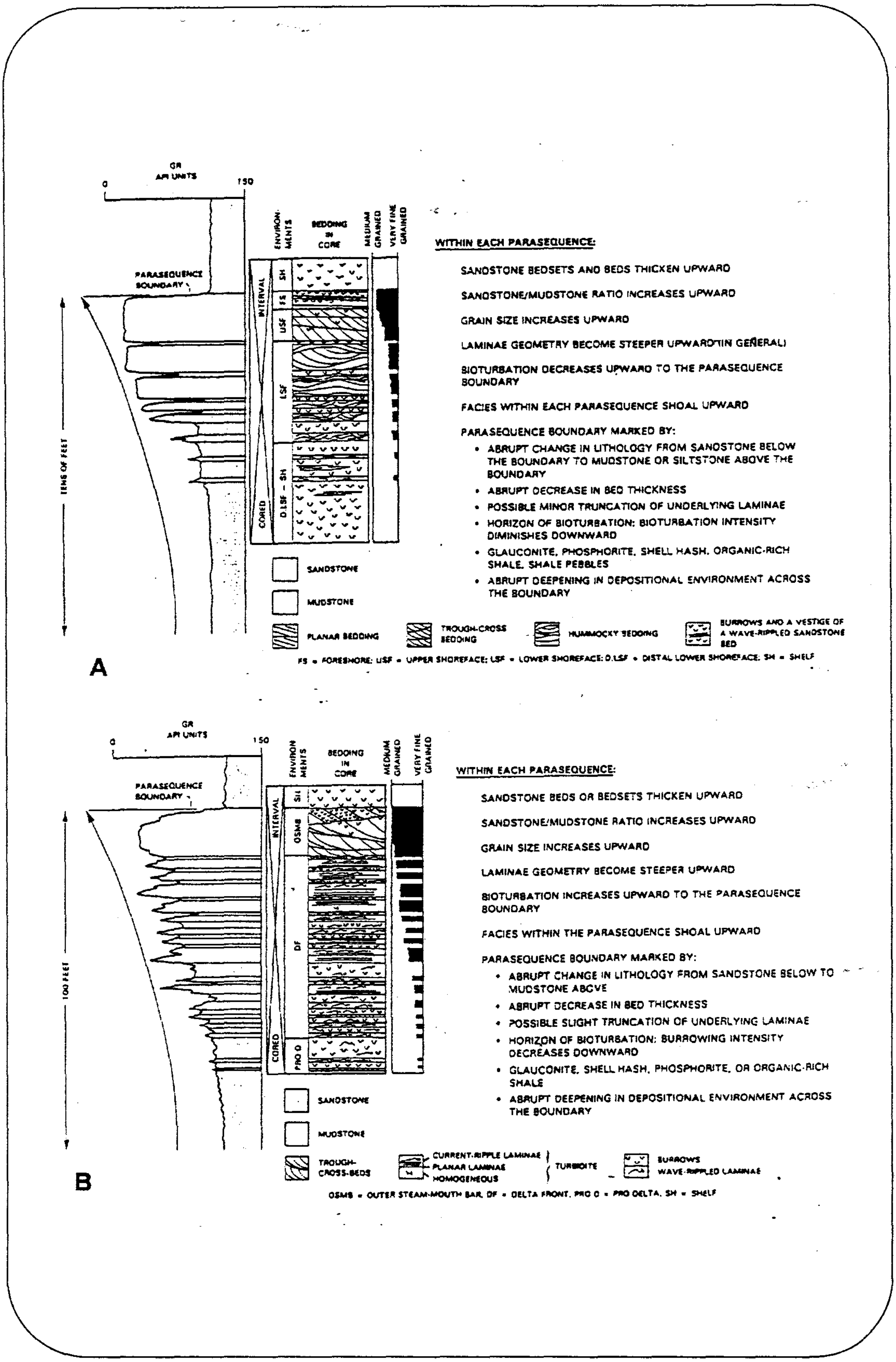
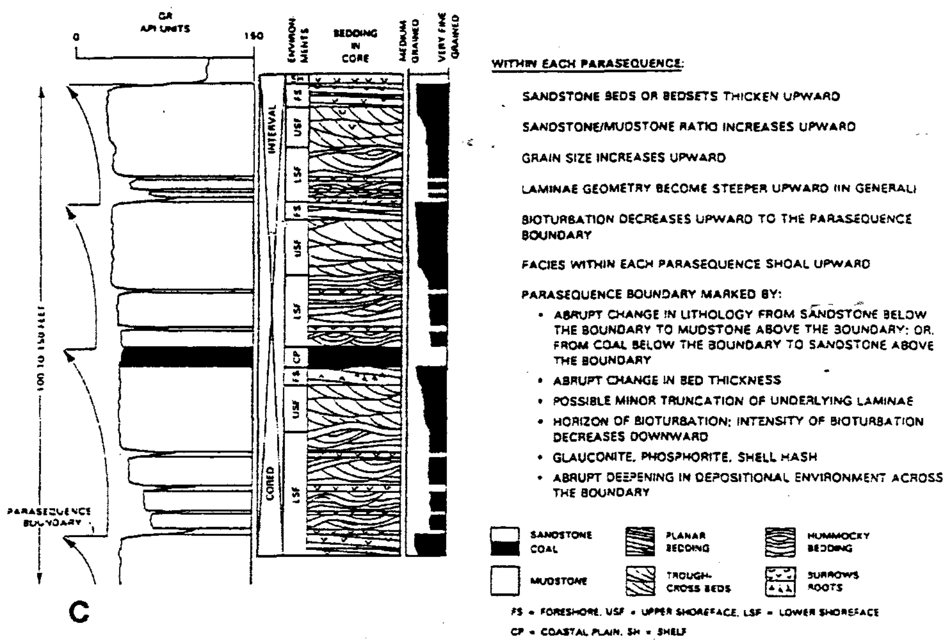


Figure 3.15 - Vertical sequences. (A) Stratal characteristics of an upward - coarsening parasequence. This type of parasequence is interpreted to form in a beach environment on a sandy, wave - or fluvial - dominated shoreline (B) Stratal characteristics of an upward - coarsening parasequence. This type of parasequence is interpreted to form in a deltaic environment on a sandy, fluvial - or wave - dominated shoreline (From Van Wagoner et al., 1990).



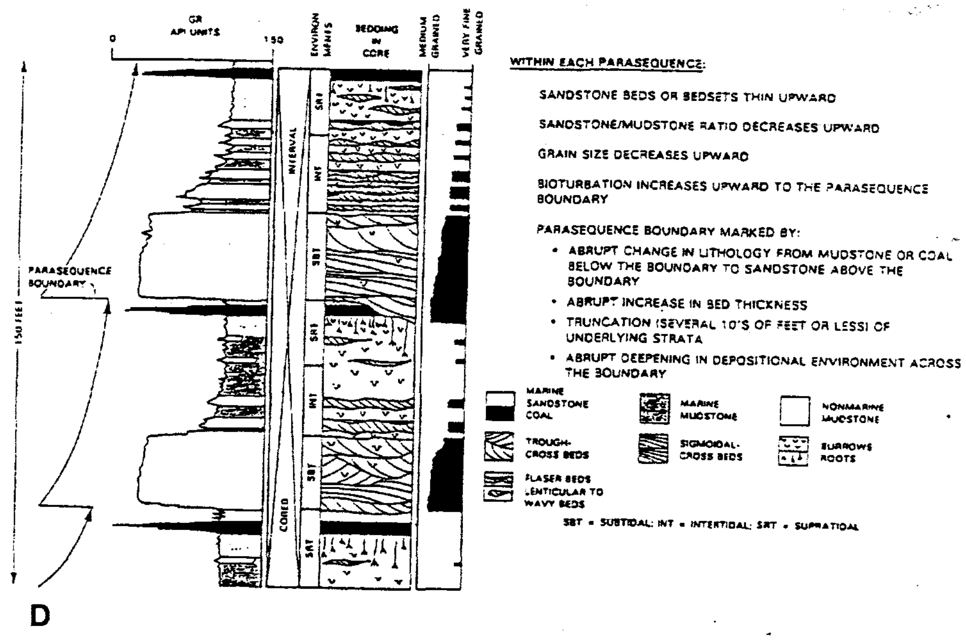


Figure 3.16 - Vertical sequences (C) Stratal characteristics of a stacked upward - coarsening parasequence. This type of parasequence is interpreted to form in a beach environment on a sandy, wave - or fluvial - dominated shoreline where the rate of deposition equals the rate of accomodation. (D) Stratal characteristics of two upward - fining parasequence. These type of parasequence are interpreted to form in a tidal flat to subtidal environment on a muddy, tide - dominated shoreline (From Van Wagoner et al., 1990).

responses to serve as predictors of large scale features (Figures 3.16 & 3.17). These large scale factors can include features such as depositional environment, basinal trends, diagenetic reactions and the occurrence of economic mineral deposits such as placer gold (Klein, 1991a).

Although sequence stratigraphic models have been rigorously applied to marine sequences, they have not been widely used or applied to non - marine deposits such as the late Archaean Witwatersrand basin (Wright and Marriott, 1993). The failure of present sequence stratigraphic models to accurately predict sedimentary facies and surfaces in fluvial sequences is because of the inadequacies of incorporated modern principles of fluvial geomorphology (Wescott, 1993). These were based upon the assumption that geomorphic systems respond predictably to sea level changes and result in recognisable stratal geometries. Base level changes of fluvial systems due to a relative sea - level fluctuation has become a key element in sequence stratigraphic models (Wescott, 1993).

One should be cautious when applying sequence stratigraphic concepts to continental strata (fluvial deposits) considering the complex response of rivers to base level changes and sediment supply. Relative sea - level can be considered as a stratigraphic as well as geomorphic base level for shallow marine and coastal non - marine settings. Stratigraphic base level further inland is more complex and is variable such as graded stream profile for fluvial strata, groundwater table for aeolian strata and lake levels of intermontane sediments (Shanley and McCabe, 1994).

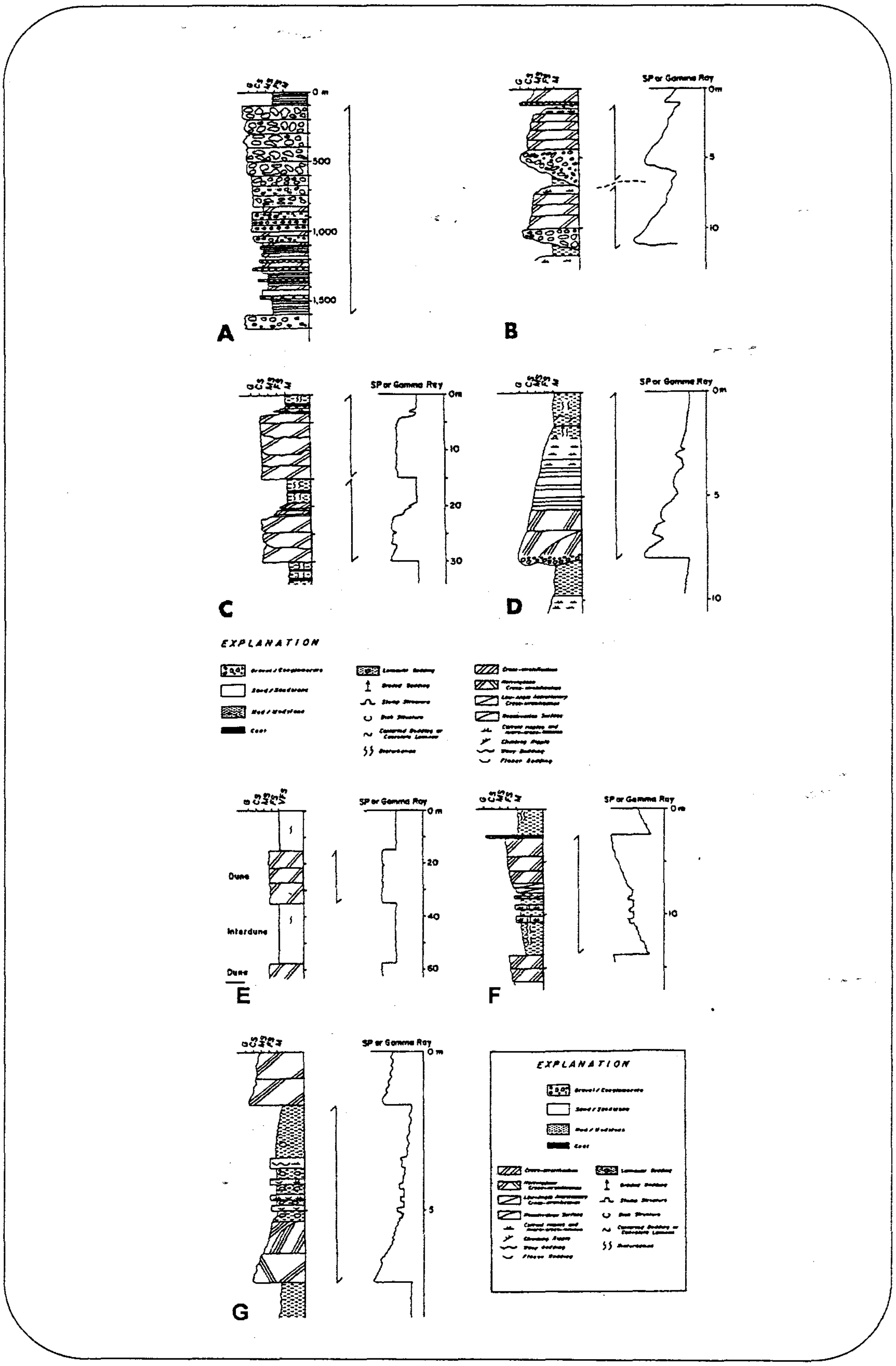


Figure 3.16 - Vertical sequences and geophysical log shapes. (A) Alluvial fan, (B) Braided river, (C) Anastomosing river, (D) Meandering river, (E) Eolian dune, (F) Barrier island and (G) Tidal flat deposits (From Klein, 1991c).

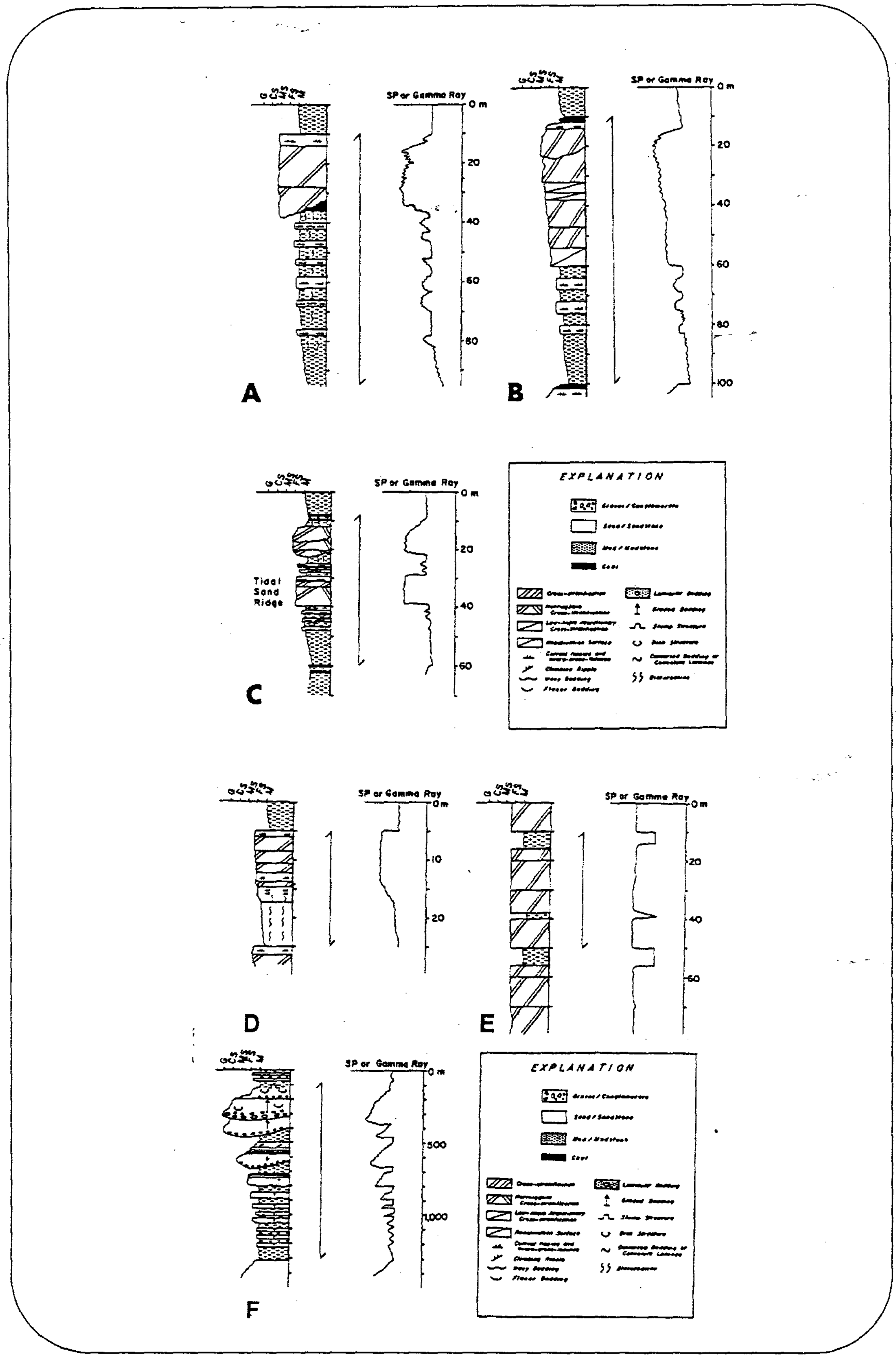


Figure 3.17 - Vertical sequences and geophysical log shapes. (A) River - dominated deltaic, (B) Wave - dominated deltaic. (C) Tide - dominated deltaic sediments, (D) Wave - dominated shelf, (E) Tide - dominated shelf and (F) Ancient submarine fan deposits (From Klein, 1991c).

4 TECTONIC SUBSIDENCE AND GEOHISTORY ANALYSIS

Geohistory analysis provides a basis to unravel different mechanisms of basin subsidence from quantitative analysis obtained from the stratigraphic record. The combination of geohistory analysis and sequence stratigraphic analysis enables one to distinguish between tectonic and eustatic depositional responses in the basin.

Geohistory analysis aims to produce a graphical representation of basin subsidence by using the vertical movement of a stratigraphic horizon in the basin as a reference point to study the subsidence and uplift history of the whole basin after the horizon was deposited (Angevine and Heller, 1987). Geohistory analysis requires:

- age interval,
- a stratigraphic column showing dominant, exclusive or average lithology for that age interval, and
- stratigraphic thickness for that age interval.

Most of the data can be obtained from borehole intersections or stratigraphic sections across an outcrop combined with geochronology of the rock unit. A few corrections are needed before a true tectonic subsidence curve can be generated. Additional assumptions and uncertainties need to be built into the analysis. However, most of the problems will be overcome if thick stratigraphic sections of relatively shallow water - deposits are utilised and only long term, large scale changes are studied (Klein, 1991b; Angevine and Heller, 1987).

The *time scale* determines the accuracy of the results which in turn is dependant on the accuracy of the time scale that one chooses to work with. The *paleobathymetric scale* is even less understood than the time scale. Calibrating the depth scale can be done with a certain amount of accuracy using fossils in stratigraphic sequences consisting of shelf - deposits. However, this is not applicable to late Archaean age basins, such as the Witwatersrand basin due to the lack of fossils. Most of the compaction correction methods can correct for sediment compaction but they are all based on empirical porosity/depth relationships derived from a variety of sediments. The range of data collected must be scattered sufficiently enough so that one can determine a representative range of values for depth and porosity. Furthermore, the effect of overpressured horizons, cementation and late - stage diagenesis, may lead to uncertainties (Angevine and Heller, 1987). *Sea - level* effects can lead to errors in calculating the basin subsidence history. The safest way to approach sea level corrections is to realise that short - term, small scale changes in subsidence may represent sea - level fluctuations and rather concentrate on the large scale studies by working on relatively thick sections (tens of metres) where small scale sea - level changes have a minor effect.

Angevine and Heller (1987) illustrate a simple method provided by Van Hinte (1978) to generate a tectonic subsidence curve. Initially one has to generate a stratigraphic accumulation by simply using a stratigraphic column with lithological data (Figures 4.1 and 4.2). The column needs to be subdivided into somewhat arbitrary lithological units, either based on a number of ages, number of unconformities, significant changes in paleo - water depth or significant lithological breaks, with their respective thickness (including major unconformities). In addition a chronostratigraphic column with the relevant ages of each unit is necessary to provide a geochronolgic framework. Furthermore, a column showing the distribution of the different paleobathymetric environments of each unit is necessary (Figure 4.1).

The next step is to correct this lithostratigraphic column for compaction by applying the method provided by Van Hinte (1978). It is based on the principle that the thickness of a unit at the time of deposition, and any time thereafter, is related to change in porosity of the sediment during burial (Figure 4.3). When decompacting, one actually increases the pore space to its original state by putting water back into the sediment. Therefore one needs to expand the porosity parameter by a unknown value (ϕ_N). Van Hinte (1978) points out that the grain volume does not change (assuming no significant diagenesis), only the volume of the pore space decreases during burial. Most studies suggest that porosity will decrease with depth exponentially, regardless of the lithology. Slater and Christie, (1980) have chosen simple exponential relationships for sandstone, shale and limestone (Figure 4.5) (Angevine and Heller, 1987; Klein, 1991b).

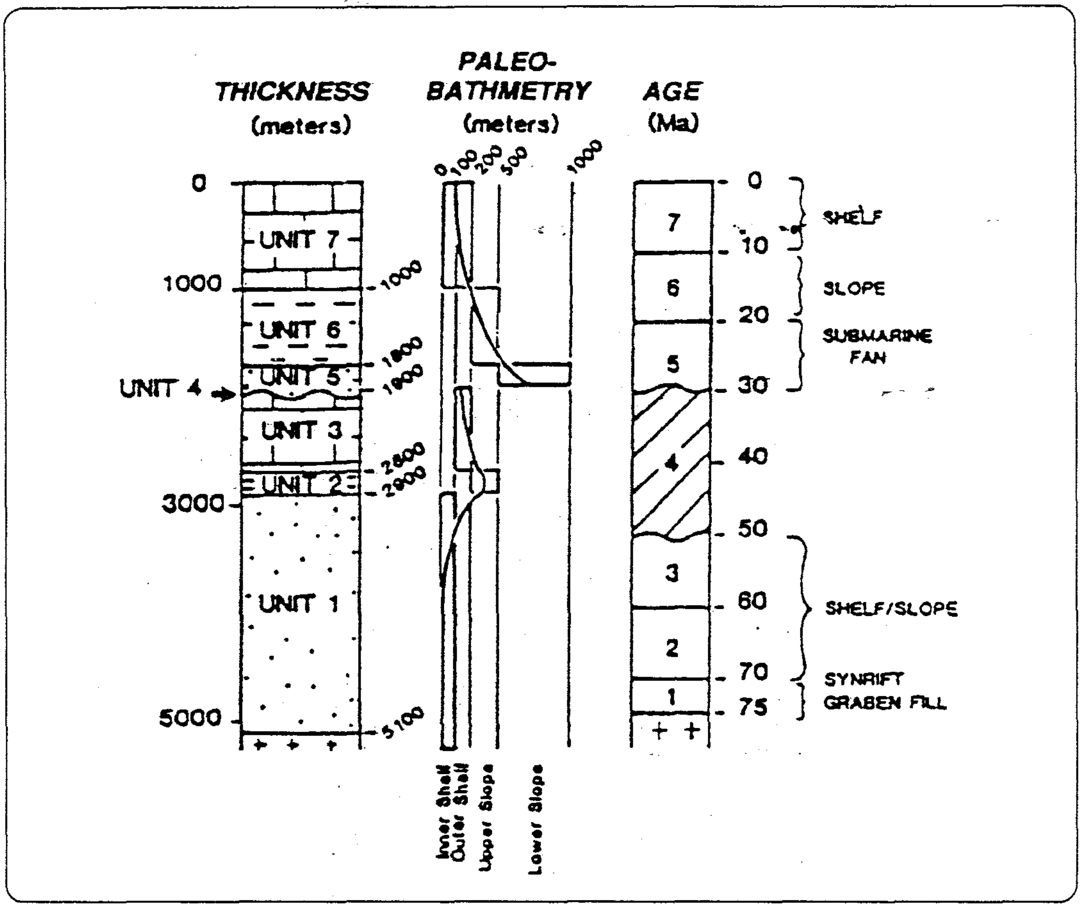


Figure 4.1 - Example of stratigraphic section used in tectonic subsidence/ geohistory analysis as discussed in text. Modified after Angevine and Heller (1987) (From Klein, 1991b).

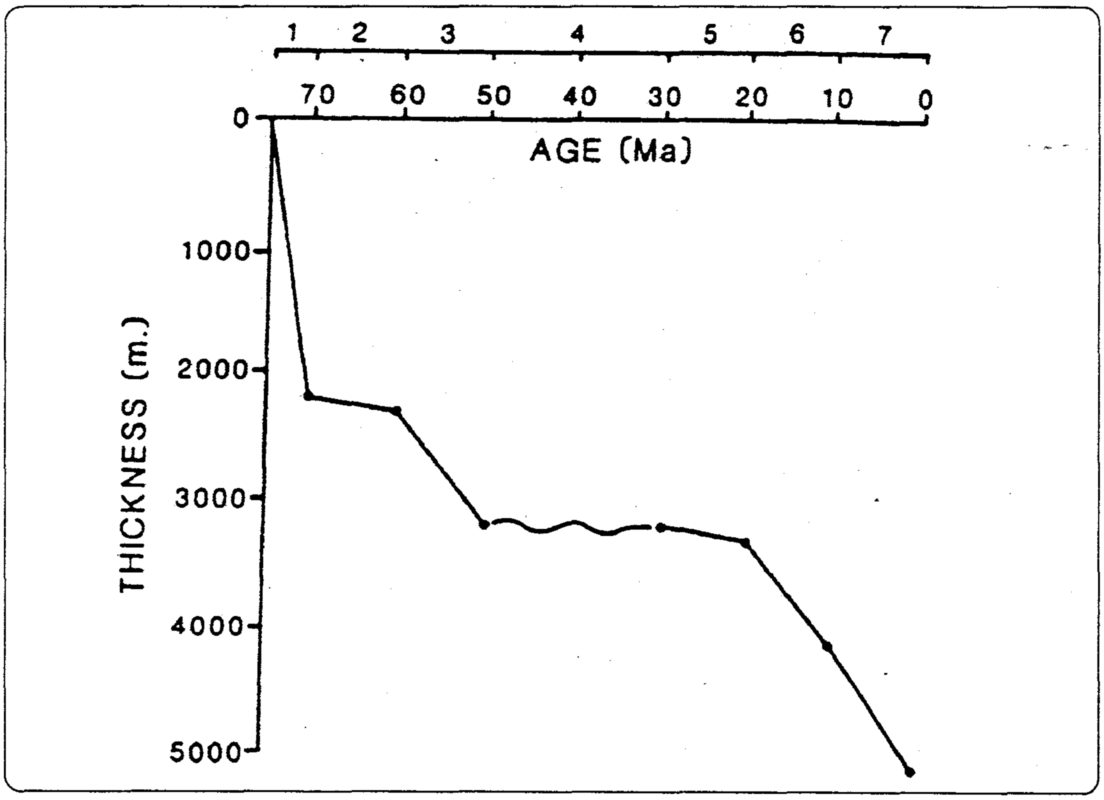


Figure 4.2 - Uncorrected sediment accumulation curve for stratigraphic section showing in Figure 4.1. Modified after Angevine and Heller (1987). (From Klein, 1991b).

Figure 4.3 provide the necessary equations to restore the thickness of units at successive stages of burial. A worksheet is set up to assist with the calculations, showing the thickness and porosity of each stratigraphic unit during burial (Figure 4.4). An example is used to illustrate the philosophy of generating a subsidence curve for a synthetic stratigraphic section that may represent a real stratigraphic section along a passive margin setting. The original data set is illustrated in figure 4.1. Porosities are determined for the midpoint of each unit by using the calculations in figure 4.3 and placing the calculated porosity in the first column. In the second column unit 7 has been removed, therefore unit 6 is 800 m and its midpoint 400 m. The porosity for a shale at 400 m can be determined from the graph in figure 4.5. This value is ϕ_N and can be used to calculate the corrected thickness (T_O) by using the equation in figure 4.5. To calculate unit 5, take half of the thickness of unit 5 (50 m) and add it to the calculated thickness of unit 6 (T_O), to obtain the midpoint elevation. Use figure 4.5 to determine the porosity for that elevation and calculate the thickness again. The entire process can be repeated for each column across the diagram, which will complete the decompaction procedure for the entire section (Angevine and Heller, 1987; Klein, 1991b).

By plotting the $\sum T^*$ values from the worksheet, a compaction corrected curve is generated which has removed the effects of compaction (Figure 4.6). The compaction corrections made are approximations of the true accumulation curve and therefore reflects the change of the magnitude of the curve, but not necessarily the shape of the curve.

The wiggled line represents a hiatus with no record of its subsidence history. The only information available, is the amount of basin subsidence prior to the hiatus and the change in water depth recorded in the sediments that directly overly the unconformity (Figure 4.6). In this case the sea floor continued subsiding during the hiatus, since the depth prior to the hiatus was 100 m deep and by the beginning of unit 5 it was in the excess of 500 m. There could be several other reasons for the difference in depth during the period when the hiatus was formed ie. faulting during time of unconformity, subsidence due to water loading, or sea - level fluctuations may have been a factor.

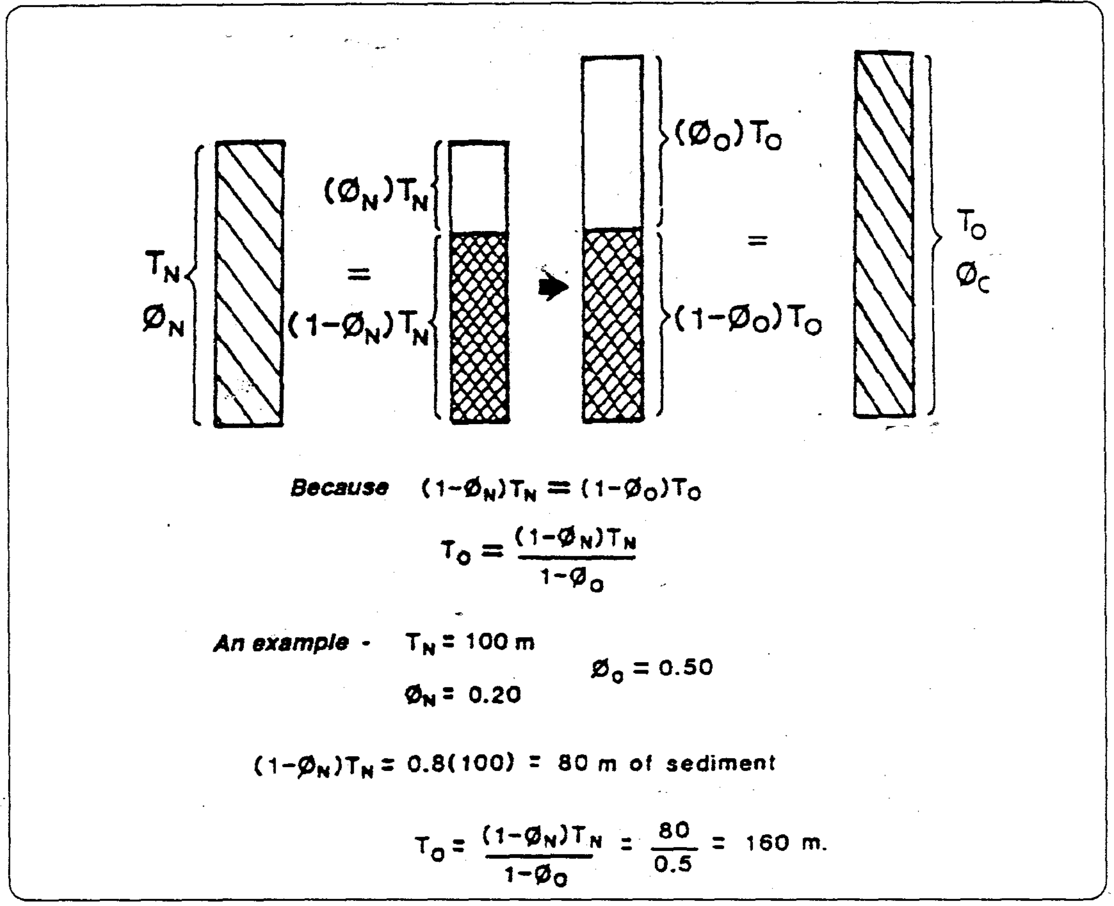


Figure 4.3 - Theory of compaction corrections developed by Van Hinte (1978) with example of calculation. Modified after Angevine and Heller (1987). (From Klein, 1991b).

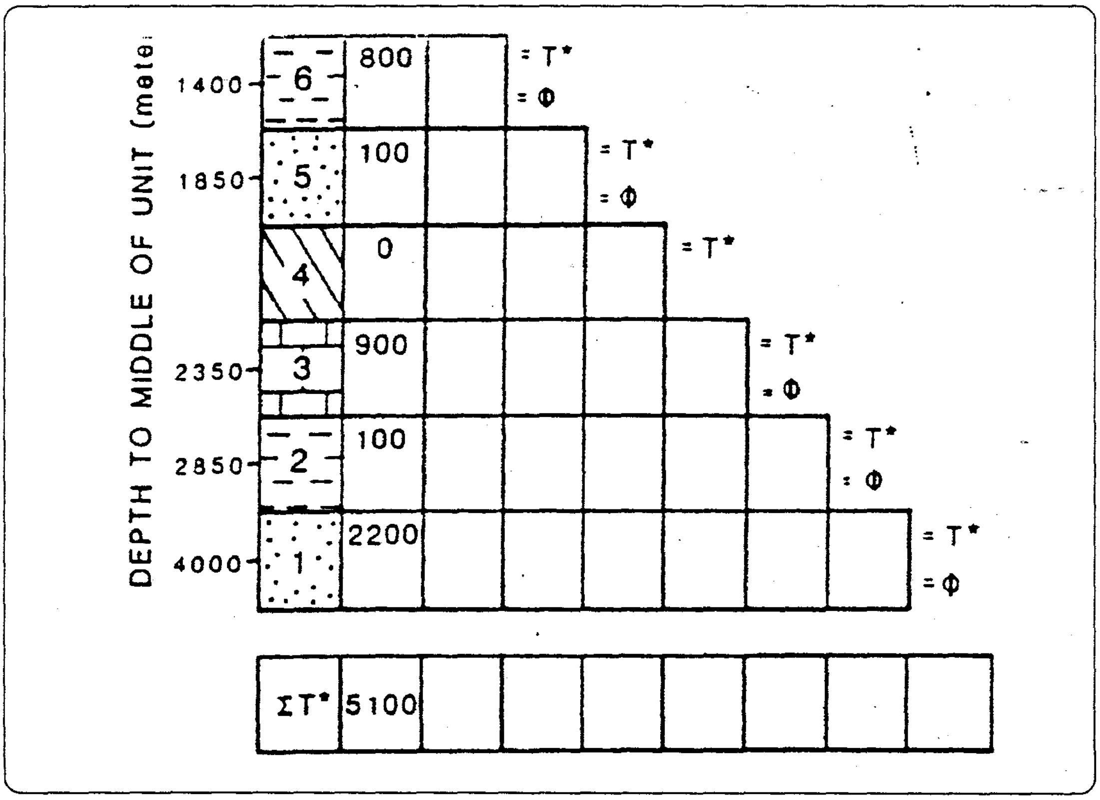


Figure 4.4 - Worksheet used for completing compaction corrections for stratigraphic section shown in Figure 4.1. Modified after Angevine and Heller (1987). (From Klein, 1991b).

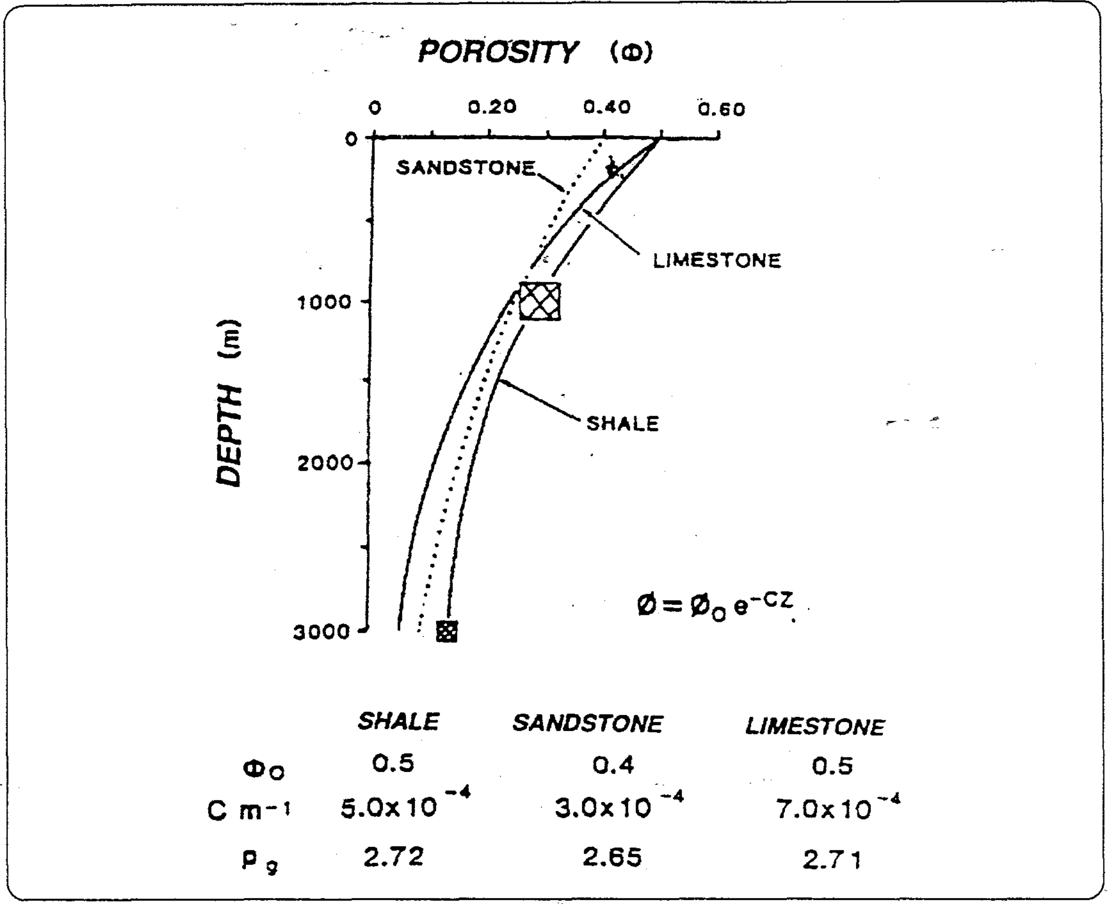


Figure 4.5 - Idealised curves for different lithologies showing porosity vs depth. Modified after Angevine and Heller (1987). (From Klein, 1991b).

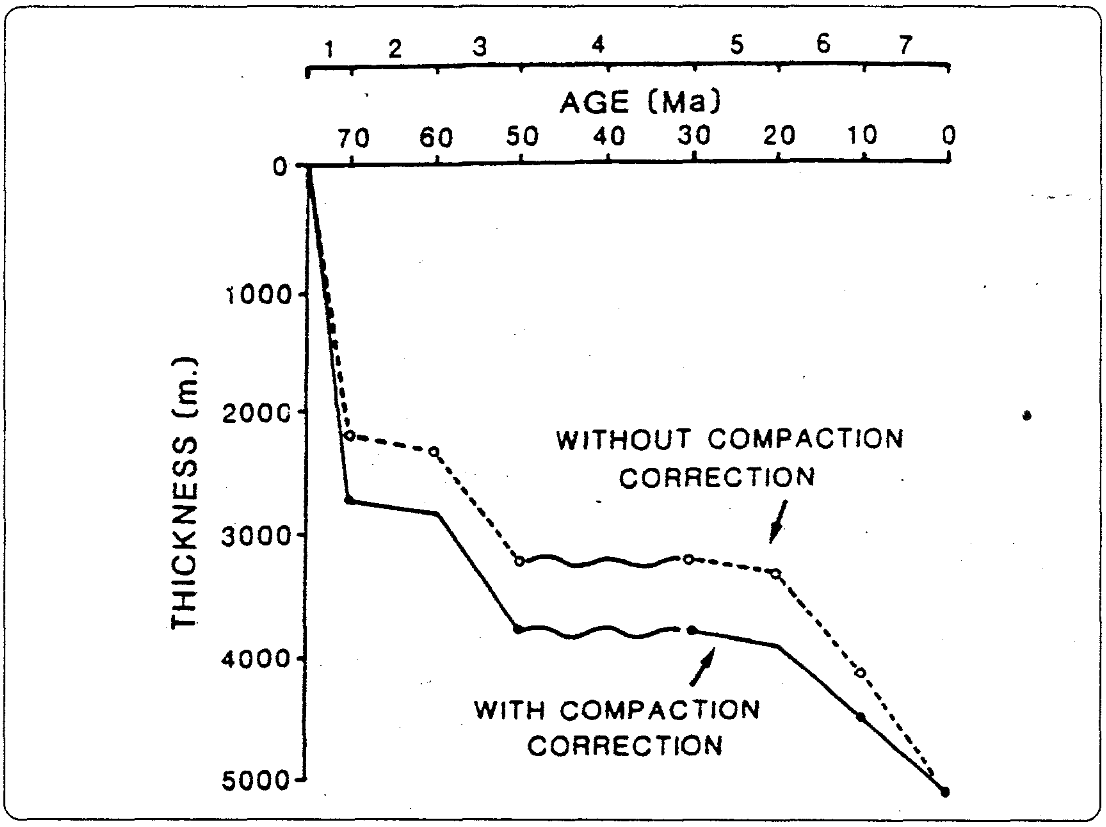


Figure 4.6 - Subsidence curve corrected for compaction based on worksheet results. Modified after Angevine and Heller (1987). (From Klein, 1991b).

A total subsidence curve has been generated so far, illustrating the effect of all the factors influencing the subsidence of the basin, sediments loads, tectonic loads and sea - level changes (Figure 4.6). In order to understand the tectonic history or the record of sea - level changes, all the effects of subsidence caused by loading during sediment deposition needs to be eliminated. This is done by a method called *backstripping*, which permits the calculation of the correction for sediment density and sediment load, to produce a tectonic subsidence curve (Figure 4.7). This method is based on the assumption of an isostatic model where sedimentary units can be removed and the basin is allowed to isostatically rebound (isostatic balancing) (Figure 4.7).

The density and isostasy needs to be calculated by utilising the following formulas (Figure 4.7):-

$$\mathbf{Z_i = S^*(\rho_a - \rho_s / \rho_a - \rho_w) + Wd_i}$$

Where :

$\mathbf{Z_i}$	= depth to tracked horizon relative to sea - level (to be determined).
$\mathbf{S^*}$	= uncompacted sediment thickness ($\Sigma^i_1 T_i^*$ = total thickness of sediment column under the top of unit i)
$\mathbf{Wd_i}$	= water depth for unit i (known)
ρ_a	= mantle density (3.3 g/cc)
ρ_w	= water density (1.0 g/cc)
ρ_s	= the density of the sediment column at the time of deposition of unit i can be calculated by using the following equation :

$$\rho_s = \frac{\Sigma[\, \phi\rho_w + (1-\phi)\rho_g \,]\, T^*}{S^*}$$

To determine the changing load through time, one needs to repeat the calculation incrementally by successively removing individual layers. After completing these corrections one can plot the Z values of the tectonic subsidence on the same graph as that of the total subsidence curve and display the subsidence history of the basin that is due to other effects other than sediment loading (Figure 4.8).

Balance columns:

$$\rho_w W_d + \rho_s S + \rho_L 1 + \rho_a X$$
$$= \rho_w Z + \rho_L 1 + \rho_a (W_d + S + 1 + X - Z - 1 \qquad [15]$$
$$= \rho_w Z + \rho_L 1 + \rho_a W_d + \rho_a S + \rho_a 1 + \rho_a X - \rho_a Z - \rho_a 1 \qquad [16]$$
$$= \rho_w Z + \rho_a W_d + \rho_a S + \rho_a X - \rho_a Z \qquad [17]$$

Lump Z terms:

$$\rho_a Z - \rho_w Z = \rho_a W_d - \rho_w W_d + \rho_a S - \rho_s S \qquad [18]$$

Factor:

$$(\rho_a - \rho_w) Z = (\rho_a - \rho_w) W_d + (\rho_a - \rho_s) S \qquad [19]$$

Divide through by $\rho_a - \rho_w$:

$$Z = \frac{\rho_a - \rho_s}{\rho_a - \rho_w} S + W_d \qquad [20]$$

If change of sea level (ΔSL) is known:

$$Z = \frac{\rho_a - \rho_s}{\rho_a - \rho_w} S + W_d - \Delta SL \frac{\rho_w}{\rho_a - \rho_w} \qquad [21]$$

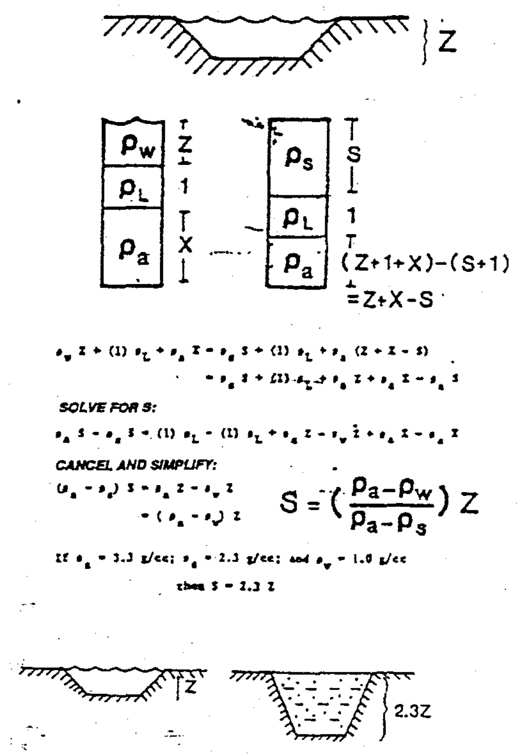


Figure 4.7 - An illustration of the backstripping technique which allows to eliminate all the effects of subsidence caused by loading during sediment deposition. The formula allows for the calculation of the Z value (From Angevine and Heller, 1987).

The magnitude of stretching can be determined by comparing the tectonic subsidence curve with the McKenzie b curves for stretching factors of 35 km thick crust (Figure 4.9). For this particular example there seems to be good agreement between the thermal subsidence with a β factor of 1.75 for the first 50 - 60 Ma of subsidence. This is followed by a much smaller stretching factor of 1.5 which appears to be caused by inversion tectonics from extension to compression (Cloetingh, 1988).

The ultimate goal of these exercises are to generate the tectonic subsidence curve in order to distinguish tectonic from eustatic signals in a stratigraphic column for the Witwatersrand basin.

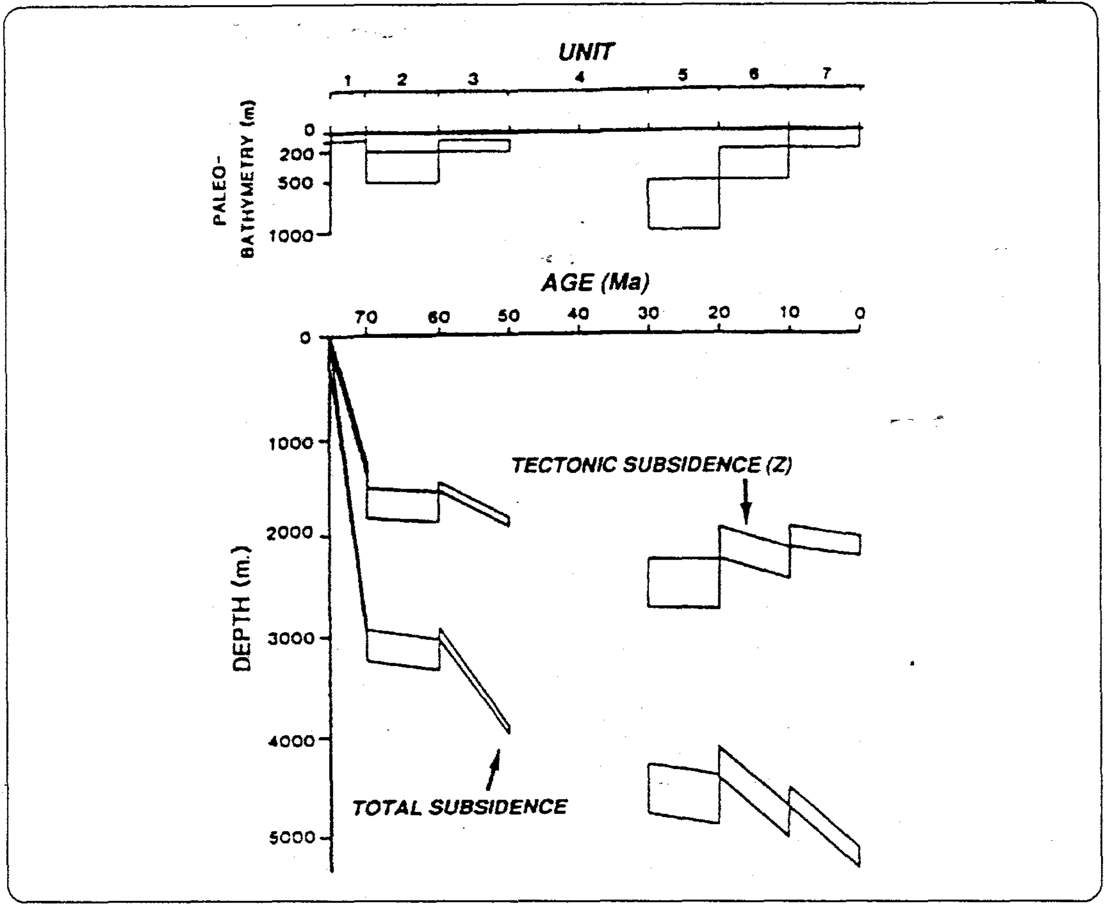


Figure 4.8 - Tectonic subsidence curve generated by eliminating all the effects of subsidence caused by sediment loading . Modified after Angevine and Heller (1987) (From Klein, 1991b).

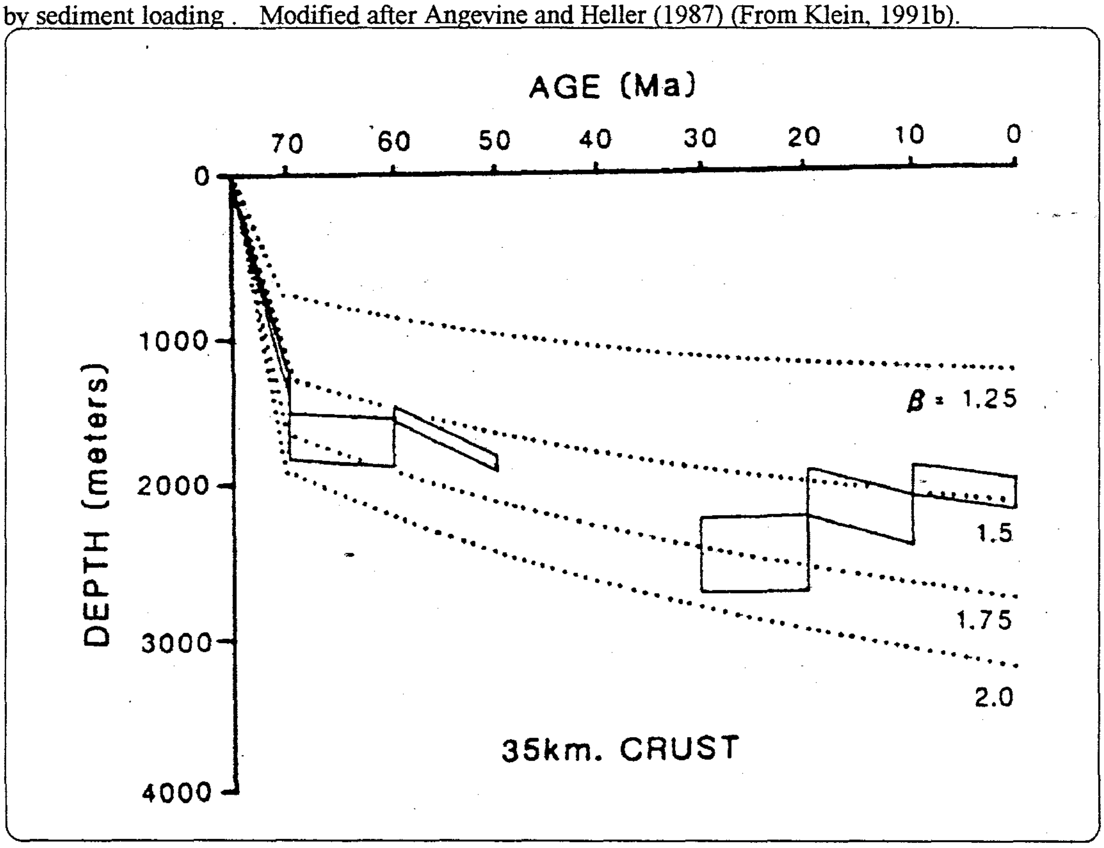


Figure 4.9 - Thermal subsidence history of 35 km crust compared with tectonic subsidence curve generated from stratigraphic section in Figure 4.1. Modified after Angevine and Heller (1987). (From Klein, 1991b).

5 WITWATERSRAND BASIN AS AN EXAMPLE OF BASIN ANALYSIS APPLIED TO PLACER FORMATION IN A FLUVIAL/MARINE SETTING

5.1 Introduction

After presenting a comprehensive overview of the principles of basin analysis and sequence stratigraphy, the techniques and methodology can now be applied to fluvial and marine sequences within the Witwatersrand Supergroup. Reference will be made to specific placers (eg. Carbon Leader placer, Steyn placer, Vaal placer, Kimberley placer).

Minter et al. (1986, 1988), Kingsley (1987), Els (1991) and numerous other authors contributed a large amount of detailed work on the sedimentology of the auriferous placers of the Witwatersrand Supergroup. However, very little, if any, work has been undertaken to understand the overall genetic stratigraphic setting of the placers and even less is understood about the tectonic subsidence history of the Witwatersrand basin.

The majority of investigators concentrated on the "reef envelope" and interpreted their findings in complete isolation from depositional events that occurred in the deep footwall and hangingwall strata. They therefore overlooked important data regarding the temporal and spatial distribution of depositional sequences and their bounding surfaces.

Sequence stratigraphy/basin analysis has not really been applied to the Witwatersrand basin. However, it is obvious from the previous reviews in chapters 2, 3, and 4 that such techniques have great applicability and considerable economic implications with respect to exploration for extensions of existing mines and new ore bodies within the Witwatersrand Supergroup.

Of interest to the geologist exploring the Witwatersrand basin is, the spatial and temporal relationship of these ore bodies within a depositional sequence and their recognition in non - marine/marine - marginal settings. So far much criticism has been directed to the use of sequence stratigraphic concepts, in particular as to what extent does the sealevel curve portray truly global (eustatic) or more local (relative) changes of sea - level (Miall, 1991). Miall (1991) continued his criticism by arguing that the concept of rivers grading to a bayline (Jervey, 1988; Posamentier et al., 1988; Posamentier and Vail, 1988) is far too simplistic to explain fluvial response to eustasy and subsidence (Wescott, 1993). Shanley and McCabe (1994) went on by stating that many of the original conclusions regarding changes in both stratigraphic and geomorphic base level, slope of the fluvial system and fluvial responses to base - level changes, were oversimplified. Thus, considering stratigraphic base level in terms of a graded stream profile as defined by Leopold and Bull (1979), i. e. "one *in which, over a period of years, slope, velocity, depth, width, roughness, pattern, and channel morphology delicately and mutually adjust to provide the power and efficiency necessary to transport the load supplied from the drainage basin without aggradation or degradation of the channels*", allows for better understanding of changes of an equilibrium surface within a fluvial setting (Shanley and McCabe, 1994). Therefore, a major factor controlling stratigraphic base level of a fluvial system will be a change in relative sea - level in the lower coastal plain (eustasy and subsidence). Consequently, much work has been done to gain new insights into stratigraphic base - level changes and the associated fluvial response (Shanley and McCabe, 1994).

These reservations regarding the basic concepts of equilibrium profile changes are valid and have great implications on the fluvial sequences within the Witwatersrand Supergroup. Simple base - level controlled models will not necessarily apply to most river systems that are influenced by "complex responses" (Schumm, 1993) and climatic factors, thus creating highly variable depositional sequences (Wright and Marriott, 1993). Despite all the criticisms, justified or not, subdivisioning of the stratigraphic record into more meaningful sequences and parasequences of sequence stratigraphy based on the disconformities and the relationship of these surfaces to relative sea - level fluctuations, is here to stay (Plint et al., 1992).

5.2 Lithostratigraphic Description of the Witwatersrand Supergroup

Before discussing the Sequence stratigraphic subdivision of the Witwatersrand Supergroup, an overview of the lithostratigraphy is given. The reference profile used to illustrate the different lithologies of the Witwatersrand Supergroup is representative of the Carletonville/Krugersdorp area (Figure 5.1).

The Witwatersrand is subdivided into a lower West Rand Group and an upper Central Rand Group. The West Rand Group however, is further subdivided into the Hospital Hill, Government and Jeppestown Subgroups. Three broad depofacies can be recognised in the West Rand Group i.e. marine shelf, fluvial braidplain and debris flow. Debris flows can be considered to be part of either the shelf or braidplain sequences (Beukes and Buxton, 1991).

The Hospital Hill Subgroup is mainly dominated by marine inner and outer shelf deposits, initiated by widespread transgressive, wave - dominated (mesotidal) conditions, with the first appearance of shale horizons marking the change to low energy tide dominated (macrotidal) depositional conditions (Figure 5.1) (Tainton, 1994).

Sandy inner shelf orthoquartzites consist of two varieties i.e. current dominated and storm wave dominated deposits. The current dominated variety consists of small scale trough cross - bedded, well sorted medium to coarse - grained or gritty orthoquartzites with subordinate pebble layers. Whereas the storm wave - dominated inner shelf consists of medium to coarse - grained orthoquartzites with hummocky cross - stratification and large symmetrical wave ripple marks capped by wave - ripple, crossed - laminated, very fine grained quartzites and siltstones.

The outer shelf deposits are represented by massive to finely laminated mudstones. The starved shelf deposits are represented by magnetic mudstones, finely laminated iron - formation and banded - iron formations (i.e. Contorted bed).

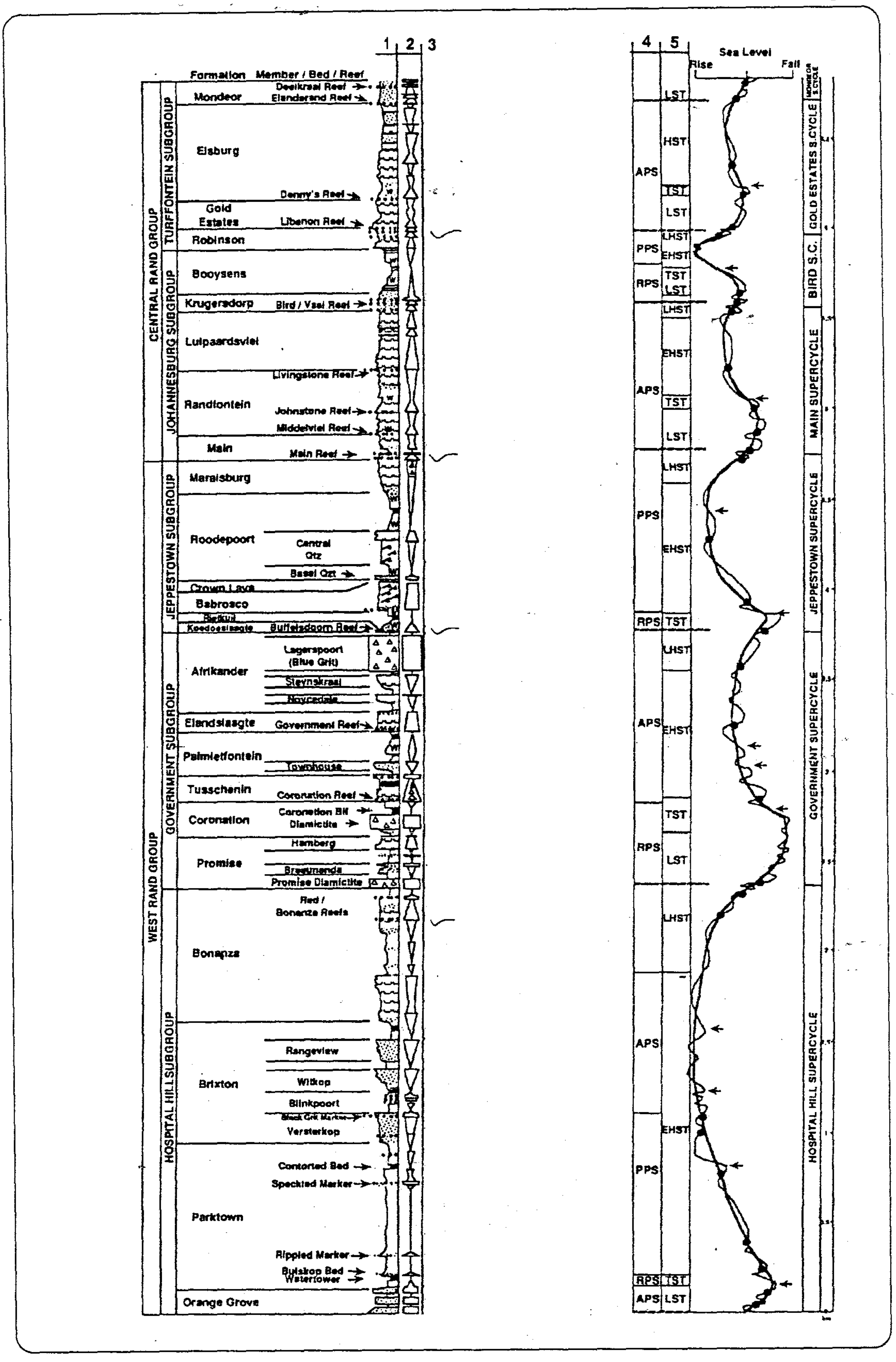


Figure 5.1 - Composite genetic and sequence stratigraphic reference profile for the Witwatersrand Supergroup in the Carletonville - Krugersdorp area. Column 1: Lithology - (w = wackstone, clear = argillite, black dots = conglomerate, troughs = argillaceous quartzite, triangles = diamictites). Column 2: Genetic succession. Column 3: Incised valleys. Column 4: Parasequence sets - (A = aggradational, P = progradational, R = retrogradational). Column 5: Systems tracts (ST) - (L = lowstand, T = transgressive, EH = early highstand, LH = late highstand) (Modified from Beukes and Nelson, 1995).

Fluvial braidplain deposits are common in the Government Subgroup, with minor incursions of transgressive shelf orthoquartzites (Figure 5.1). The fluvial deposits are sand - dominated braided fluvial sequences, consisting of slightly argillaceous to argillaceous quartzites, with trough and planar cross - bedding. The grain sorting is moderate to poor, unimodal palaeocurrent patterns are characteristic together with the upward fining facies successions. Parts of the Jeppestown Subgroup are also considered to be fluvial braidplain deposits in particular the upper parts of the Roodepoort and Maraisburg Formations (Beukes and Buxton, 1991).

In a broad sense the overlying Central Rand Group can be subdivided into a lower Johannesburg and an upper Turffontein Subgroup. The Johannesburg Subgroup is composed of mainly fluvial braidplain deposits consisting of predominantly interbedded quartzites and conglomerates with minor occurrences of shales and wackstones. The two main economic horizons are located near the base and top of the subgroup in the Main Conglomerate Formation and Krugersdorp Formation (Figure 5.1).

The Booysens Shale Formation separates the two subgroups and is used as a regional marker in the Central Rand Group. It is represented mainly by marine shelf sands and shelf muds during a period when the Central Rand basin was flooded by a major transgression. A condensed section marks the end of the transgressive systems tract and demarcates the maximum flooding surface (Figure 5.1).

The succeeding Turffontein Subgroup can be subdivided into three distinct lithological units. The lowermost unit is represented by the Robinson and Gold Estates Formation, composed entirely of cobblestone and conglomerate interbedded with granulestones, gritstones and coarse - grained sand. The central unit (Elsburg Quartzite Formation) is mainly represented by proximal marine shelf deposits, comprising predominantly marine orthoquartzites with minor interbeds of gritstone and wackestone. The Mondeor Conglomerate Formation defines the upper unit which is composed of conglomerate and medium to coarse - grained quartzite (Figure 5.1).

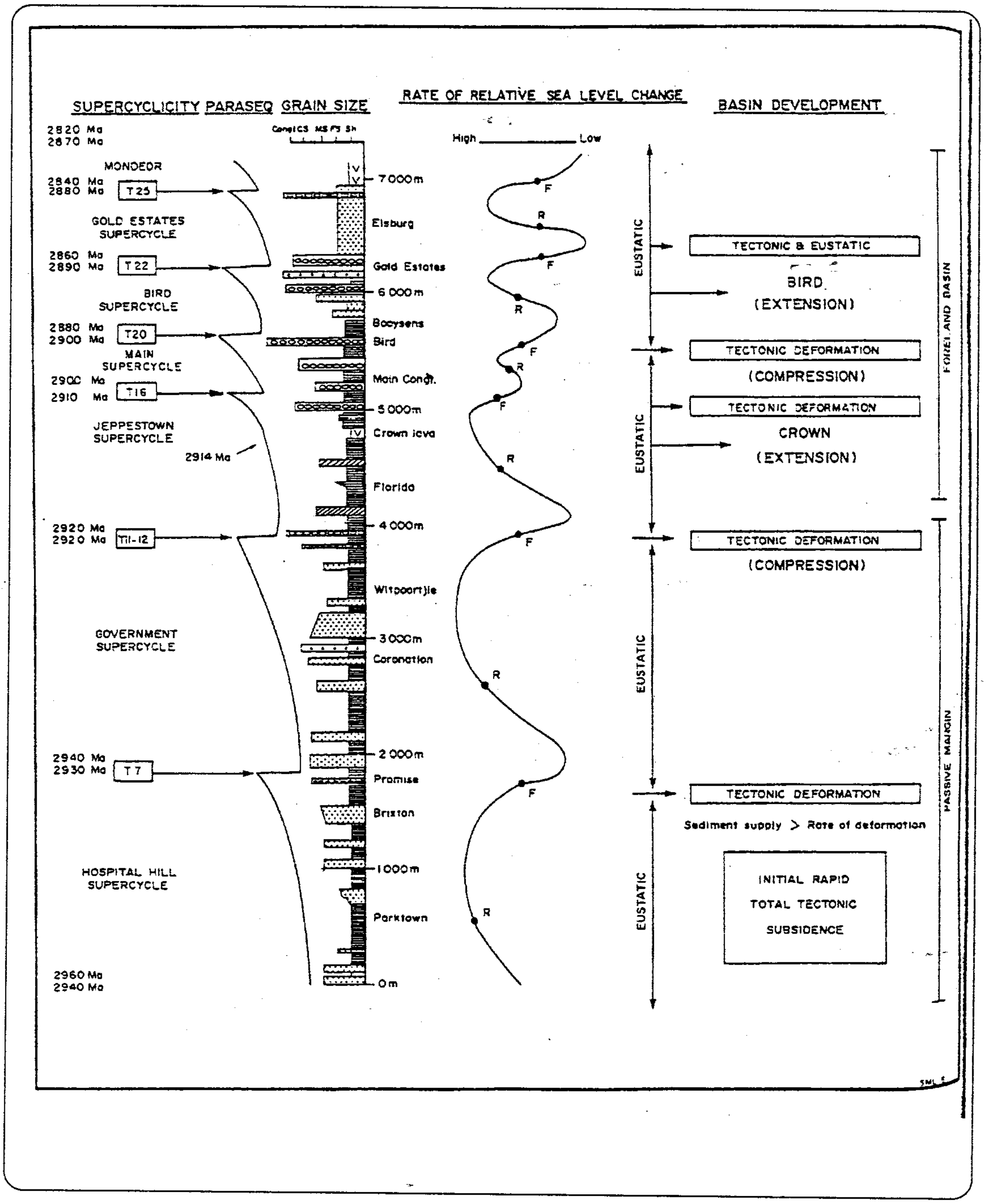


Figure 5.2 - A simplified sequence stratigraphic profile for the Witwatersrand Supergroup of the Carletonville - Krugersdorp area showing rate of relative sea - level change and basin development. The lefthand column depicts the correspnding supercyclicity parasequences (From Nelson et al., 1995).

The economic horizons within the Turffontein Subgroup are associated with conglomeratic zones in the fluvial braidplain deposits (Figure 5.1). Two prominent erosive cycles are identified on a basin wide scale near the lithostratigraphic bases of the Johannesburg and Turffontein Subgroups (Figure 5.1). They are characterised by valley incision filled with basal lag deposits, diamictites, interbedded wackestones, with prominent marine shelf mudstones/siltstones.

5.3 Placers in the Central Rand Group

Concentrations of relatively high density, stable and durable minerals, which have been deposited and altered mechanically in a sedimentary environment are known as placers. Within the Witwatersrand these are referred to as "reefs" (a term that has no association with coral reefs).

Placer deposits require five factors for formation and preservation :-

- a large volume precursor
- followed by deep weathering to unlock the heavy mineral/metal from the source rocks in the hinterland and/or underlying alluvial sequences within the depository
- suitable transport medium
- concentration of the unlocked material
- and preservation of the placer from subsequent destruction (Karpeta, 1994).

Within the Witwatersrand most of the economically important placers are found within the Central Rand Group. Karpeta (1994) and Beukes and Nelson (1995), studied the effects of sea - level fluctuations during the formation of auriferous palaeoplacers in the basin (Figure 5.1).

Four major gold - bearing placer zones have been identified in the Central Rand Group which are assigned to the following supercycles as defined by Beukes and Nelson (1995) (ie. Main, Bird, Gold Estate and Mondeor Supercycle) (Figures 5.1 and 5.2).

Due to their cyclical nature, these supercycles are predictive. Beukes (1990) generated an idealized sequence stratigraphic profile of important kerogen - bearing auriferous placers in the Witwatersrand Sequence (Figure 5.3).

It consists of the following:-

- Most of the placer zones are in close stratigraphic proximately to major transgressive units, such as the Jeppestown and Booysens shales.
- They are usually capped by upward coarsening aggradational and progradational sediments, composed of offshore muds, interbedded with greywackes (turbidites). In turn they grade upward through offshore marine orthoquartzites into distal fine grained alluvial plain argillaceous quartzites. These are overlain by more proximal gritty to pebbly braided alluvial plain quartzites.
- This upward shoaling sequence is capped by planes of equilibrium sedimentation or erosional disconformity (Type 1 or Type 2). Placer formation takes place in this environment.
- Progradation is succeeded by a transgressive sheet conglomerate and well sorted orthoquartzites representing a marine transgressive ravinement lag deposit. A thin shale unit sometimes caps this unit probably, deposited in deeper water conditions.

According to Karpeta (1994) two major types of conglomeratic packages can be distinguished within these supercycles that are associated with gold - bearing placers. These are planar sheet - like conglomerates and channelised conglomerates.

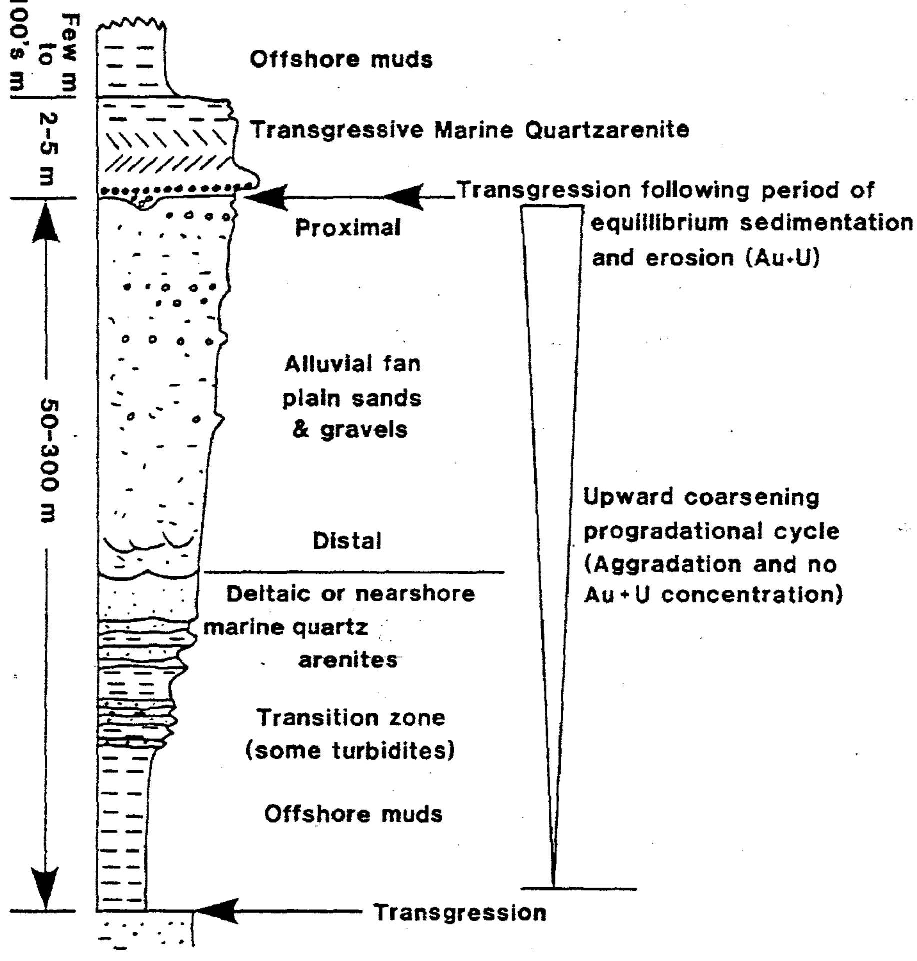


Figure 5.3 - Ideal sequence stratigraphic setting of important kerogen - bearing auriferous placers in the Witwatersrand sequence. The specific illustration is based on the Carbon Leader placer setting but, with minor modifications may also fit others such as the Vaal/Basal/Steyn combination or UK9a placer (From Beukes, 1990).

Planar sheet - like conglomerates are equated with placers such as : e.g.

- Mondeor supercycle - Elandsrand Reef, Deelkraal Reef
- Gold Estates supercycle - Kimberley Reef/Uk9a Reef/ May Reef/ Libanon reef
- Bird supercycle - Bird Reef, Vaal Reef, Basal Reef
- Main supercycle - South Reef, Nigel Reef, Main Reef Leader, Main Reef, Middelvlei Reef, Carbon Leader Reef (Figures 5.4 and 1.3)

Channelised placers can be equated with : e.g.

- Gold Estates supercycle - Kimberley channels
- Bird supercycle - Erosion channels
- Main supercycle - Brakpan channels, Erosion channels (Figure 5.4)

All the above mentioned placers have similar stratigraphic settings in each of the supercycles (Figures 5.2 and 5.3). They formed in the same basinal and depositional environments, similar placers (reef zones) can be also be identified in the West Rand Group although economically not as important as the Central Rand placers (Figure 5.1) (Beukes and Nelson, 1995).

The four major auriferous conglomerate zones are marked by four second order sequence boundaries (unconformity surfaces) (Figure 5.1). These sequence boundaries can either be Type 1 unconformity, rapid relative sea - level fall (channelised placers) or Type 2 unconformity, slow rate in relative sea - level fall (sheet - like placer), both resulting in low accommodation space potential on the shelf. It is sometimes extremely difficult to determine which disconformity actually marks the sequence boundary within these conglomeratic zones due to the obliteration of the eustatic maximum and minimum sea level changes within an alluvial sequence (Wright and Marriott, 1993; Beukes and Nelson, 1995).

The sheet - like character of the placers can be distinguished from other non - economic conglomerates in the sequence on the basis of their intimate association with

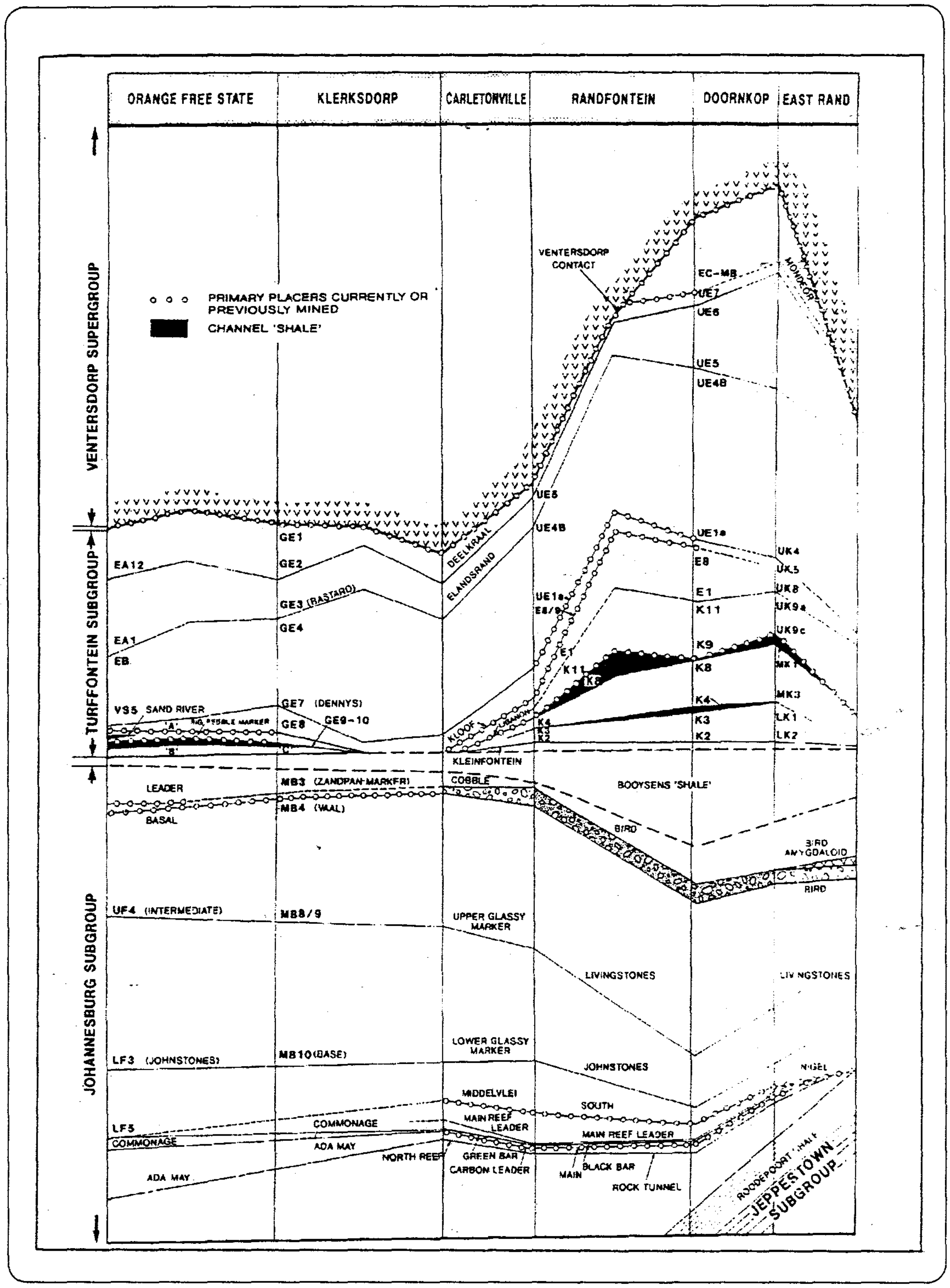


Figure 5.4 - Regional correlation of the major economic placers and lithological breaks of the Central Rand Group around the northwestern periphery of the Witwatersrand Basin adjacent to the uplifted source region (From Tainton, 1994).

a major low angle disconformity and clean, mature, glassy orthoquartzite with marine transgression characteristics. They formed during periods of gentle open folding of the strata in the basin.

Externally, placers such as the Basal/Steyn, Vaal, Carbon Leader, and Kimberley/UK9a (Evander/East Rand goldfield) are characterised by a general tabular and sheet - like geometry with a slightly undulating channelled base and a flat top (Figures 5.1 and 1.3). Basal channels of the Basal/Steyn, Vaal and Carbon Leader placers comprise an interconnecting system of broad, low - sinuosity, ribbon - like bodies of approximately 500 m wide and 5 km long. Sometimes the placers are confined to the channels and form shoestring orebodies (Minter, 1991).

Internally these placers occur as gravel facies consisting either of a single pebble lag on an erosional surface or discrete sheet - like accumulations of pebbles. The average thickness of these bodies is approximately 60 cm, but they can reach thicknesses that upto 1 - 6 m. The gravel facies contains planar and trough cross - beds forms with a sandy matrix. The associated quartz arenites are either horizontally or trough cross - bedded and rarely planar cross - bedded. The gravel facies consist of clast sizes ranging from small - pebbles to cobbles, well rounded to subrounded shapes and a small fraction displaying ventifacts suggesting conditions of abrasion by dominantly fluvial processes with and element of aeolian modification.

The orthoquartzites immediately overlying the placers have greater degree of textural maturity than their bounding lithologies. This can either be attributed to the nature of the hydraulic flow regime or to marine transgressions slightly modifying the previous alluvial - fan delta deposits. Evidence to substantiate the marine influence is the lateral continuity of these orthoquartzites and the sudden deviations in palaeocurrent patterns of the overlying marine deposits i.e. Zandpan member above the Vaal placer (Tainton, 1994) (Figure 5.4). It is suggested that perennial flows produced the more mature sediments and the ephemeral flows account for the conditions depositing the subwackes (Minter, 1991).

The pebble assemblages of sheet - like placers such as the Basal/Steyn, Vaal, and Carbon Leader consist of an oligomictic assemblage of vein quartz, chert and minor amounts of quartz arenite, silicified shale and quartz porphyry clasts.

Gold and other heavy minerals are usually concentrated either on:-

- scour surfaces (pebble lags)
- within clast supported conglomerates
- and on foreset and planar bedding surfaces in the quartzites.

Within the clast - supported conglomerates, the top and the bottom surfaces tend to be better mineralised, representing the lowermost surface of degradation and the topmost winnowed surface (Minter, 1991).

The channelised conglomerates (Kimberley erosion channels, Carbon Leader Erosion Channel etc.) on the other hand, are characterised by disconformities associated with valley incision, filled with a complex succession of slumped sheets, load cast pod - like remnants, reworked basal conglomerates by erosion and recycling of the underlying sequence, pyritic quartzites, diamictites and chloritoid shale (Karpeta, 1994; Beukes and Nelson, 1995) (Figure 5.1). The deeply incised channels can be up to 100 m deep and 1 km or more in width. These channel courses occur basin wide. Those east of Johannesburg (East Rand goldfield) are more sinuous than the ones in the Carletonville, Klerksdorp and Welkom goldfields where they appear to be more linear.

A common feature of these gold - bearing placers is their relationship with palaeosurfaces. Many of these placers, if not of them, owe their lateral continuity to the fact that they rest on major erosion surfaces (Figure 5.4). These have chronostratigraphic significance. These erosional surfaces can be equated with sequence boundaries that separate sequences within the Witwatersrand stratigraphy in a sequence stratigraphic context. However, according to Minter (1991) many placers, such as the Vaal, Carbon Leader, Basal/Steyn and Middelvlei Placer are localised and owe their lateral extent to two or more coalescing alluvial - fan deltas.

Karpeta et al. (1994) and Beukes and Nelson, (1995) recognised the potential of sequence stratigraphy to late Archaean Witwatersrand Supergroup auriferous palaeoplacers. Unfortunately these palaeoplacers are essentially deposited in a fluvial dominated environment, on alluvial - fan deltas and are therefore non - marine. Although sequence stratigraphy has had its main application to littoral and shallow marine sequences, recent studies have indicated that juxtaposed non - marine and marine sequences can be dealt with equally successfully by sequence stratigraphic models (Nummendal et al., 1993; Posamentier and James, 1993; Wright and Marriott, 1993; Wood et al., 1993; Shanley and McCabe, 1994).

5.4 Important Factors Controlling Placer Formation in the Witwatersrand basin

The application of sequence stratigraphy provides a method to collect the necessary information to construct a relative sea - level curve for the Witwatersrand Supergroup, based on the interpretation of sequence boundaries and their correlative conformities as well as maximum flooding surfaces (Van Wagoner et al., 1990). This information combined with parasequence stacking patterns, provides the basic building blocks for the interpretation of the systems tracts, which allows one to generate a curve depicting sea - level fluctuations (Figure 5.1). In addition, the tectonic subsidence curve provides the necessary information regarding the evolution of the Witwatersrand basin (Figure 5.9). This combined information has major predictive utility as to the distribution of lithofacies and depofacies in the basin fill, the rate of deposition, the rate of accommodation created in the basin, accelerated and decelerated tectonic subsidence rates reflected by the sediment loading of the basin floor, etc.

Differences between shallow marine shelves and alluvial plains are an important variable controlling the way a fluvial system responds to a relative sea - level fall. It is the change in slope between the continental valley and the shallow marine shelf that is of interest during the late highstand preceding the relative sea - level fall (Figure 5.5).

These factors can become important when considering conditions for placer formation.

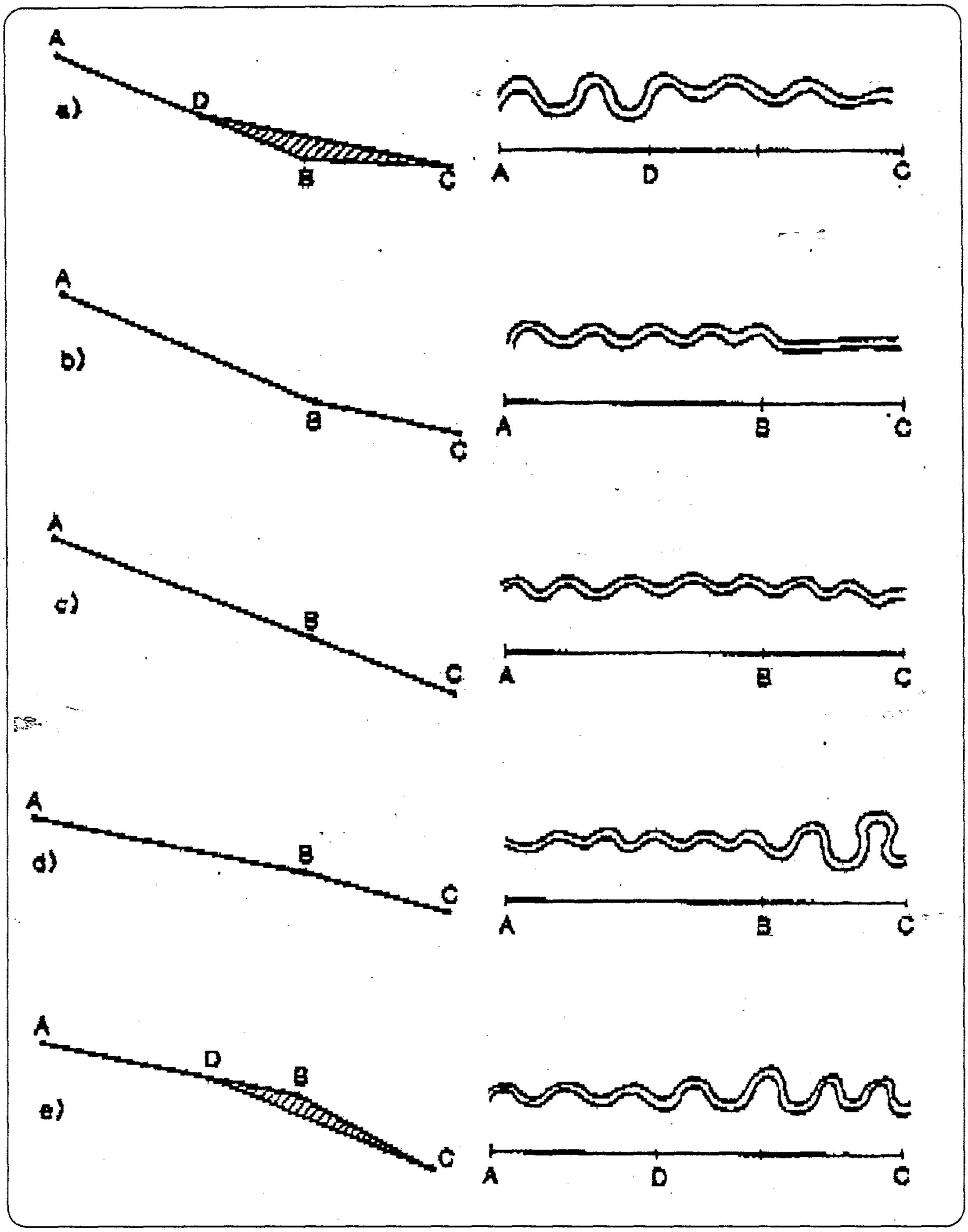


Figure 5.5 - Effects of baselevel changes across continental shelves of different inclinations from B to C. In examples (a) and (b) the slope of the continental shelf is gentler than the stream channels. In example (c) it is identical and in examples (d) and (e) it is steeper. However in example (a) the the decrease in slope is larger and in example (e) the slope of the shelf is much steeper. See text for explanations (From Schumm, 1993).

Three possible situations can occur across the continental shelf:-

- the slope of the shallow marine continental shelf is gentler than the alluvial plain (stream channels) or
- it can be identical or
- steeper (Schumm, 1993; Wescott, 1993; Shanley and McCabe, 1994).

Posamentier and Allen (1991), Shanley and McCabe (1994), Schumm (1993) and Wescott (1993) attempted to provide a framework to understand and predict the response of fluvial deposits to a base fall in all three of these situations, assuming water and sediment discharge to be constant (Figure 5.5).

- case 1 - they predicted significant fluvial deposition and no incision, which can be accomplished by an increase in channel gradient, decreasing the sinuosity resulting in a wider, shallower and straighter channel (Figure 5.5(b)).
- case 2 - minor incision will occur with the channel extending itself across the shelf without aggradation or degradation (state of equilibrium) (Figure 5.5(c)).
- case 3 - an increase in sinuosity will adjust the channel flowing across the previously submerged shelf by becoming deeper and narrower in order to maintain equilibrium, with minor incision (Figure 5.5(d)).

In most cases it is not necessary for the fluvial system to respond to a relative sea - level drop by major valley incision, even when the rate of relative sea - level fall is greater than the rate of subsidence (Figures 5.5 (b - d). Only minor entrenchments are necessary to adjust accompanied with merely changing the channel pattern to accommodate the base level change across differences in slope gradients. However, in cases where the slope of the shelf is much steeper, the channel will not be able to adjust by just changing its pattern but channel incision will also take place (Schumm, 1993;

Wescott, 1993, Shanley and McCabe, 1994). Slope conditions between the prograding alluvial - fan deltas and marginal marine shelf within the Witwatersrand basin must have been a controlling factor in the response of river systems during the formation of the placers (e.g. Basal and Steyn placers).

Posamentier and Vail (1988) developed the following sequence stratigraphic models for non - marine (fluvial) sequences, which have been widely accepted and applied. According to Posamentier and Vail (1988) there are two sequences of fluvial clastic deposits, dependent on the timing of eustatic changes. *Type 1 sequences* consist of fluvial deposits occurring as linear, incised valley fills, during lowstand wedge and transgressive deposits. Unconfined fluvial deposition of widespread floodplain deposits only occurs after the incised valleys are filled. This occurs within the late highstand systems tract. *Type 2 fluvial deposits* are limited to widespread floodplain deposition during the *late* highstand systems tract (Wright and Marriott, 1993).

Placer formation is preceded by late highstand fluvial deposits, responding to basinward shifts of the equilibrium point due to an increase in subaerial accommodation space (Figure 5.8A). This gives rise to widespread alluviation, characterised by prograding fluvial parasequences sets, sometime after the eustatic peak, during a relative sea - level fall. These sets generally grade from distal finer grained quartzites to more proximal gritty and pebbly quartzites, typical of braidplain deltas during the late highstand phase. The fluvial deposits seldom, if ever, contain any significant gold concentrations (Beukes, 1991).

A fall in relative sea - level produces two types of unconformities, denoted Type 1 or Type 2 (Figure 5.8B). The nature of these two unconformities depends on rate of relative sea - level fall at the physiographic shoreline break and the difference in slope between the alluvial plains and the shallow marine shelf (Shanley and McCabe, 1994; Schumm, 1993; Wescott, 1993; Wood et al., 1993; Wright and Marriott, 1993).

Channelised placers such as the Kimberley channels or Carbon Leader Erosion Channels are developed on narrow continental shelf or perched shelf settings with prominent physiographic shoreline - breaks. These shelf settings can result in Type 1

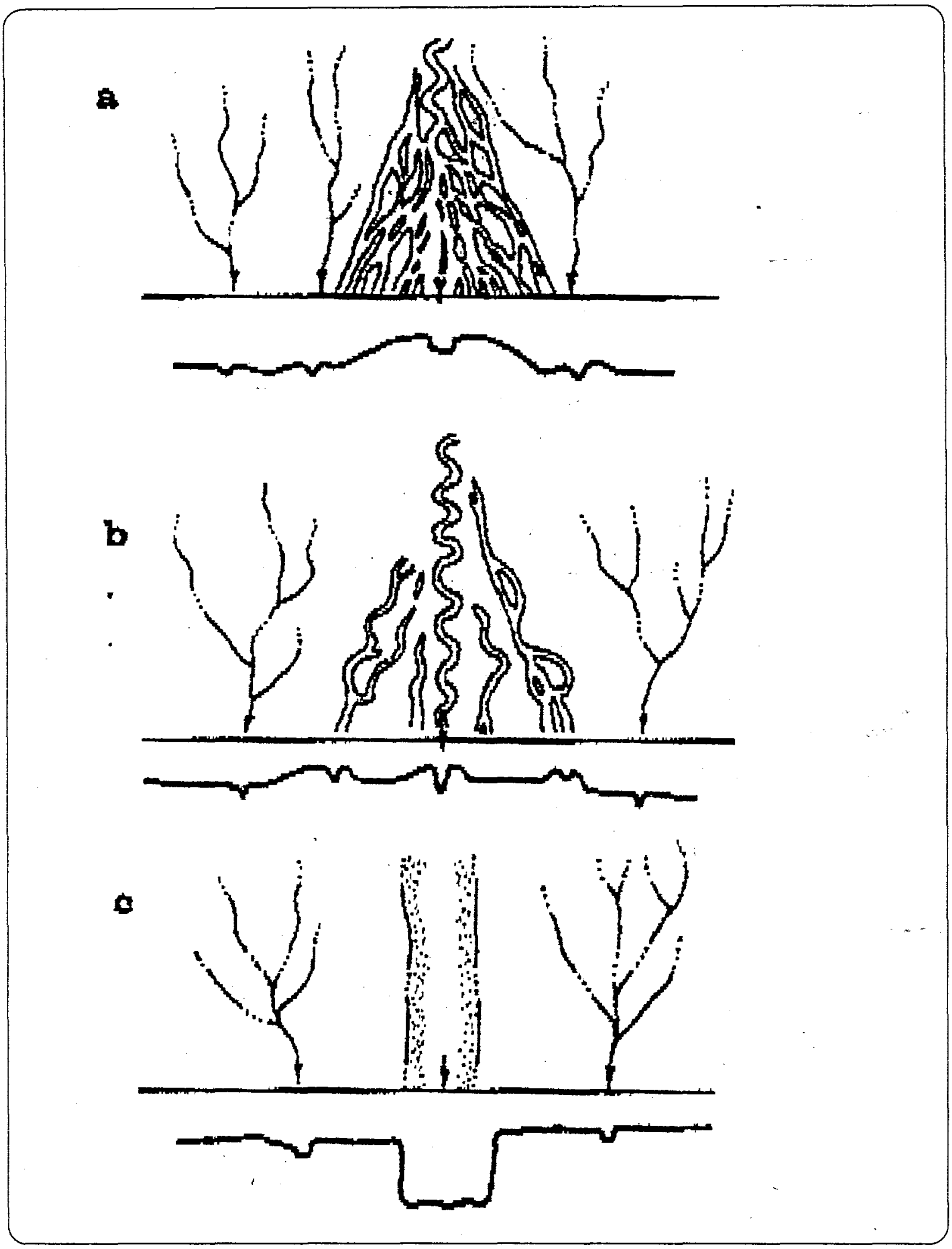


Figure 5.6 - Plan views and cross sections of three different shelf situations: (a) slope of shelf is gentle (figure 5.5a) and deposition produces an alluvial - fan delta which in the cross section rises above the shelf surface and sets the stage for fan head entrenchments. (b) slope of shelf is similar to valley slope (figure 5.5c) and the channel extends across the slope building natural levees and being subjected to avulsions. (c) slope of shelf is steep (figure 5.5e) and incision occurs leaving smaller shelf channels perched above the incision (From Schumm, 1993).

sequence boundaries during a rapid fall in relative sea level, characterised by deeply incised channels extending far out across the marine shelf, delivering sediment to the shelf edge. This is followed by a rapid rise in relative sea - level causing major flooding of the incised valley. This is typical of the *channelised placers* encountered in all four of the major auriferous placer zones in the Witwatersrand basin (Karpeta, 1994; Beukes and Nelson, 1995) (Figure 5.1).

Different slope conditions can also create similar conditions of incision into underlying sequences due to a change in stratigraphic base level (Wood et al, 1993; Wescott, 1993; Schumm, 1993; Shanley and McCabe, 1994). Often shallow marine shelves have steeper gradients than the adjacent alluvial plains (Figure 5.5(c)). A change in relative sea - level in such a case, would not only lead to adjustment by changing the channel pattern but also by channel incision (Figure 5.6c). The channel will adjust its pattern by widening and braiding. This effect propagates to a point upstream until it can accommodate the new slope conditions by changing the pattern as well as the shape and roughness of the channel (Figure 5.5 (d & e)).

In most cases incised channels are known to become out of phase because upstream degradation causes downstream aggradation (Figure 5.7). During the headward erosion of the tributaries in a drainage basin, erosion gives rise to excess sediment which is conveyed to the main channel, and is temporarily deposited in the channel. This raises the local base level of the tributaries causing a readjustment of the slope and sediment load decreases. The main channel can then transport the previously deposited excess load. Slight degradation takes place, with some reworking of the channel fill, resulting in further concentration of heavy minerals (Schumm, 1993; Wescott, 1993). The initial concentration of heavy minerals tends to take place along the basal lag of the channel during the initial phase of downcutting. Both these stages are favourable conditions for placer formation (Karpeta, 1994; Beukes and Nelson, 1995). This process is episodic because the wave of erosion moving up the channel reaches the drainage basin causing renewed erosion of the tributaries and the cycle repeats itself. The net effect of this intrinsic reponse is the transportation of sediment via the main channel to the depositional basin in a series of pulses. This is recorded in the stratigraphy as a stacked series of upward - fining sequences which become progressively finer and thinner. However, a rapid sea - level rise would flood the main channel, preventing the formation of the stacked upward - fining sequences and instead deposit marine shelf mudstones and siltstones (Schumm, 1993; Wescott, 1993).

Sheet - like placers (e.g. Vaal, Basal/Steyn and Carbon Leader placers) appear to be associated with Type 2 sequence boundaries (unconformities), where changes in stratigraphic base level are related to a slow relative sea - level fall (Figure 5.8B). Different gradients between shallow marine shelves and alluvial plains will also have an effect on the disconformity that developes during a drop in base level. The development of Type 1 unconformities (Carbon Leader Erosion Channel) are related to a higher subsidence rate than Type 2 unconformities (Carbon Leader placer), resulting in little or no stream incision. During this stage, fluvial deposition ceases and erosion becomes less prominent, while the landscape gradually becomes more denuded. The cessation of fluvial deposition in this context does not imply that there will be no fluvial deposits in the succession. It merely suggests that the existing fluvial deposits will be reworked on the peneplained disconformity surface (Wright and Marriott, 1993). Within the Central Rand Group there is compelling evidence for prolonged periods of subaerial exposure of the unconformity surface due to a sea level drop. During long periods of subaerial exposure dreikanters can develope on deflation surfaces. Some occurrences of dreikanters have been observed in placers in the Witwatersrand basin (Minter, 1991). Dessication cracks in the shelf shales are also related to such periods of subaerial exposure (Karpeta pers. comm.). Deeply weathered conditions are often shown by paleosol development, however, in terms of the Witwatersrand basin they might not be preserved.

In settings where the marine shelves have gentler slopes than the adjacent alluvial plain, base level lowering might be accompanied by a significant amount of sedimentation with no incision (Figure 5.6a & b), to give rise to sheet - like placers. The channel can not adjust to the base level change by simply changing its pattern, therefore aggradation takes place and a wedge of sediment is deposited which increases the gradient on the shelf. Simultaneously the gradient of the stream decreases on the continental shelf. The channel compensates by a decrease in sinuosity and assumes a braided planform. The changes in morphology of shelf - slope channels are important, particularly where the

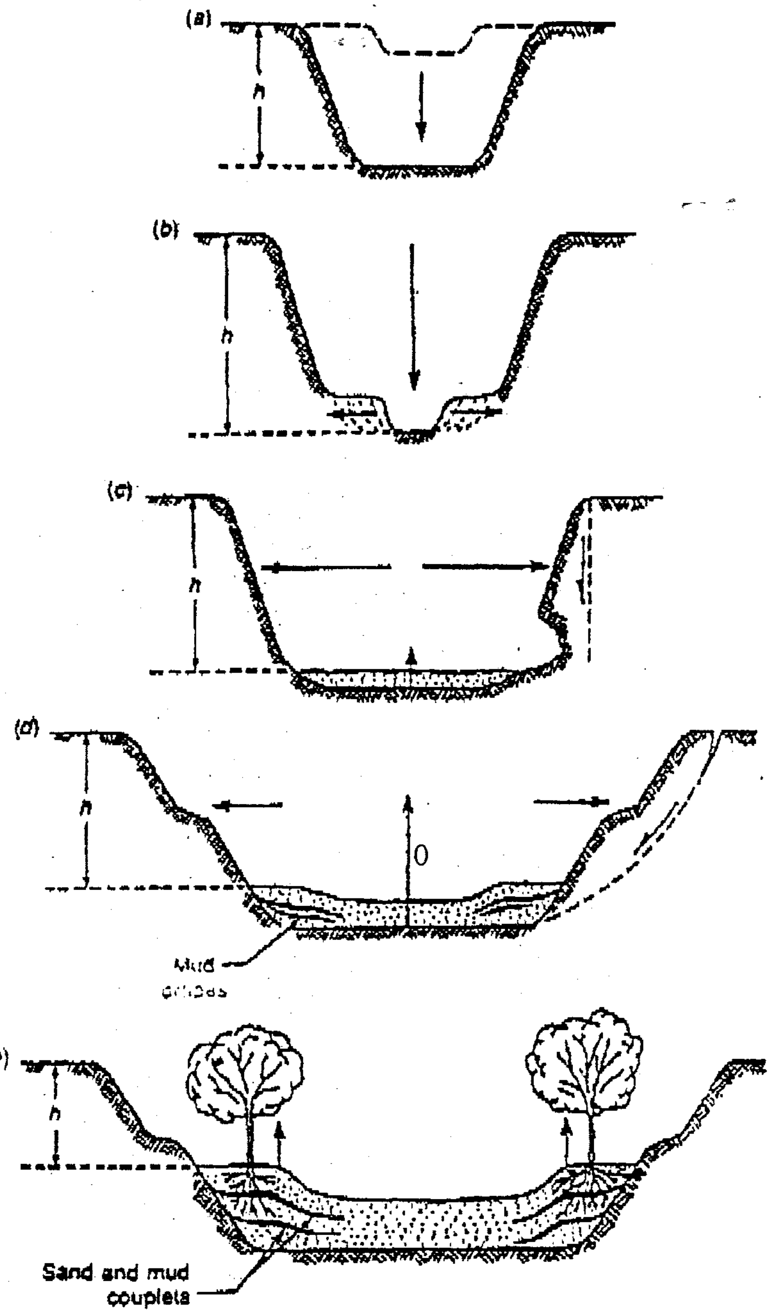


Figure 5.7 - Evolution of an incising channel from initial incision (a, b) and widening (a, b, c, d) to deposition (c, d), and eventually stability (e). This evolution occurs at one section through time, but can be observed along the channel from upstream (a) to downstream (e) (From Schumm, 1993).

gradient of the channel is maintained by a significant amount of deposition (Figure 5.6(a)). Braided rivers have the ability to carry increased sediment load, consequently an alluvial fan or fan - delta developes on the continental shelf as the unconfined channels shift laterally (Figure 5.6(a)) (Schumm, 1993; Wright and Marriot, 1993).

This setting is favourable for the concentration of economic gold mineralisation in sheet - like placers. Heavy minerals become concentrated in the initial stages of aggradation when a depositional wedge developes in order to establish a steeper gradient on the shelf. Gold concentration during this phase is expected to be highly sporadic in channels, gravels bars and zones of convergence in the braided system. As soon as the required gradient of the stream is established, this slope must be maintained. This can only be achieved by equilibrium conditions when the fan head becomes entrenched (degradation) and sediment becomes redeposited further down stream (aggradation). During these conditions a surface will develope on which gold and other heavy minerals will reach their optimum concentration levels, since there is equivalence in the state of degradation and aggradation. It is during this phase that reworking takes place of the existing sediments in the braidplain system. These conditions should prevail for long periods with large quantities of sediment passing through the system in order to concentrate an economical viable placer. The prolonged periods of subaerial exposure on the alluvial fan - delta allows for weathering and paleosol formation, deflation surfaces to develope through aeolian processes (dreikanters). Dessication cracks may form if the disconformity overlies shelf shales.

The next phase following concentration of gold is the preservation of this discrete conglomeratic layer and its gold content (Figure 5.8C). The rate at which the incised valleys and shelf are flooded is an important factor influencing the preservation potential of a transgression. A rapid rise in base level with a *sudden transgression* of the shelf will result in preservation of the underlying placer. A rapid transgression results in less time available to accumulate transgressive sediment at specific depositional sites and less destructive reworking. Less time is available for the late lowstand and transgressive deposits to be exposed to reworking by fluvial and shoreline processes (Wood et al., 1993).

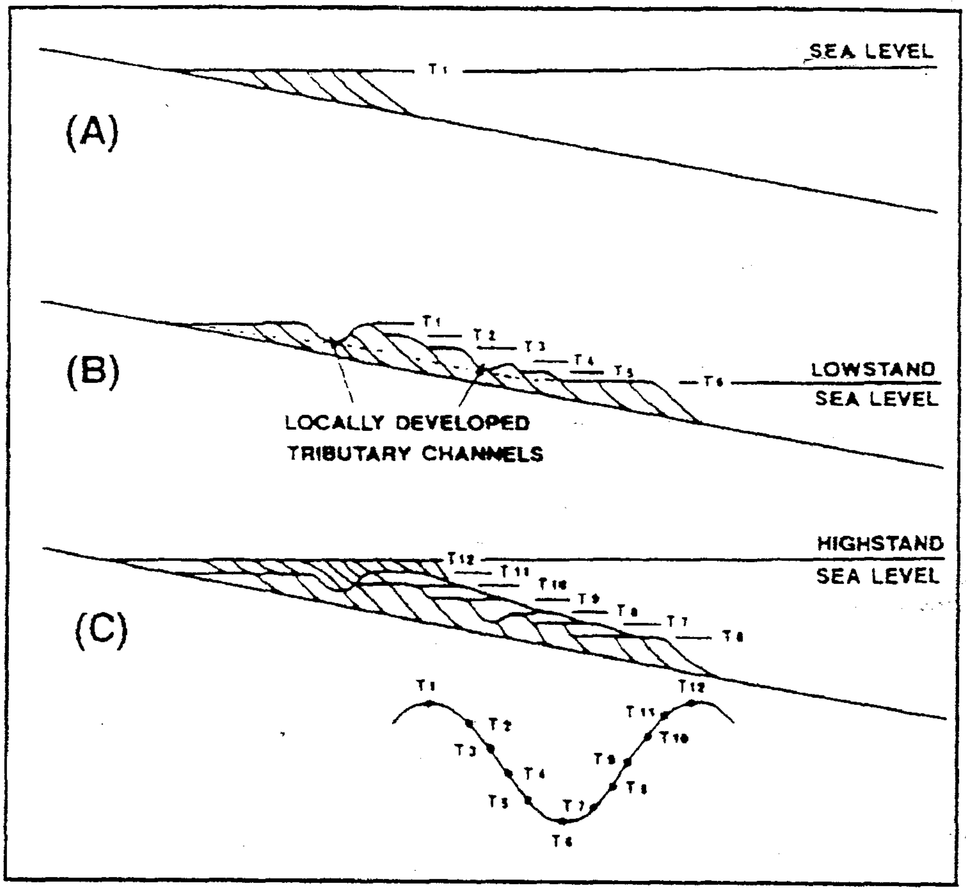


Figure 5.8 - A depositional model for a progradational basin margin which continually progrades during eustatic sea level fall (T = time). (A) The intial progradational wedge (T_1). (B) The response during the fall (T_1 - T_5) and continued progradation during lowstand (T_6). The dashed line shows the final profile for the main stream. The erosion channels are locally formed by tributaries which develop in previously submerged slope. Relief on this surface may be subtle, only a few metres. (C) The development of a time - transgressive surface (ravinement) as sea level rises (T_7 -T_{12}) and the subsequent progradation of the highstand (T_{12}) (From Wescott, 1993).

The rapid transgressions can be recognised as conspicuous, well sorted, glassy, mature shelf marine orthoquartzites, displaying bimodal to polymodal palaeocurrent directions, sometimes capped by a shelf wackestone or siltstone with wave ripple marks indicating storm wave dominated depositional environment within tidal range (eg. the Carbon Leader and Kimberley placers have hangingwall orthoquartzites).

A *slow transgression* of the shelf and incised valleys indicates that sediments have a longer period to accumulate at specific depositional sites and can accumulate thicker deposits. The previously deposited regressive and transgressive deposits become exposed for longer periods to reworking by fluvial and shoreline processes and have less chance to be effectively preserved (Wood et al., 1993).

5.5 Discussion of Contrasting Explanations of Placer Formation

Two schools of thought exist for placer formation in the Witwatersrand basin. Karpeta (1994) believes that placer formation occurs during the progradation of a alluvial fan - delta across a regressive shoreline with optimum concentration taking place during a prolonged period when degradation/aggradation is in a state of equilibrium, accompanied by braidplain reworking (lowstand systems tract). However, Beukes (1991), and Beukes and Nelson (1995) believe that gold becomes concentrated during a retrogradational phase during a marine transgression when marine transgressive lags develope (transgressive systems tract).

Beukes (1991) and Beukes and Nelson (1995) however believe that placers such as the Vaal and Carbon Leader are formed as sheet - like retrogradational fluvial lag deposits, followed by a marine transgression reworking these transgressive lags, immediately capped by marine orthoquartzites. Placers that form during a transgression are usually associated with transgressive lag deposits. Most of these transgressive lag deposits are coincident with sequence boundaries. One type of lag deposit usually consists of siliciclastic gravel and pebbles, derived from underlying strata by means of shoreface erosion during a marine transgression, subsequently concentrated as discrete beds on top of the transgressive surface, less than 60 cm thick. Transgressive lag deposits form as a result of wave and current reworking of an underlying sequence of up to 2 metres thick below the marine flooding surface. This process winnows out the finer particles while concentrating the coarser grains. There is no distinct surface separating the reworked deposits from the rest of the underlying parasequence (Van Wagoner et al., 1990).

A second type of lag deposit consists of a lag lying directly on a marine flooding surface coincident with a sequence boundary in an interfluvial area. The lag is derived from weathered paleosol material of the underlying lithologies (e.g. prior conglomeratic sequences) which formed during subaerial exposure. The subsequent transgression removes all the non durables and concentrates the residual durable material as a lag on the transgressive surface. The palaeosols are destroyed by the subsequent transgression, unless isolated remnants can be found preserved in low lying areas on the transgressed shelf (Van Wagoner et al., 1990). This type of lag deposit is similar to Karpeta`s (1994) sheet - like placers.

Regardless of which process concentrates gold on the unconformity surface, whether it is by progradation or retrogradation, placers and heavy mineral concentrations remain in close association with the subaerial exposure surface. The discrete pebble lag developed on that surface by either process, is immediately followed by a rapid transgression which effectively preserves the placer.

Other changes caused by external factors such as climate, sedimentary flux variations and tectonics may also be important. Fluvial systems are sensitive to such factors and they can have pronounced effects on the stratigraphic record and subsequent placer formation (Wescott, 1993).

5.6 Basin Subsidence Analysis of the Late Archaean Witwatersrand Supergroup

A tectonic and total subsidence curve can be generated by applying similar methods to those currently being applied to regional tectonic studies of Phanerozoic basins in Northern America and Europe. Previously various models have been proposed to explain the tectonic evolution of the Witwatersrand basin, but without any subsidence analysis to substantiate their hypotheses.

Nelson et al. (1995) provided a high resolution, well constrained basin subsidence analysis of the Witwatersrand sediments (Figure 5.9). The basin subsidence curve indicates a long term, decelerating, upward concave curve resembling the typical subsidence history characteristic of thermal subsidence basins. Overprinted on this long term curve are six convex upward curves. These convex upward trends can be equated with episodes of accelerated tectonic subsidence during increased lithospheric loading and flexuring due to compressional in - plane stresses (Cloetingh, 1988). The conglomeratic gold bearing reefs usually occur immediately after these convex upward trends during times of decelerated tectonic subsidence (Beukes and Nelson, 1995).

The subsidence curve indicates that the Witwatersrand evolved over a period of ~140 Ma, starting at ~2960 Ma and ending at ~2820 Ma. The long term concave upward curve is overprinted by six short term convex upward trends coinciding with the second order supercycle, each averaging about 20 Ma. Some of these second order supercycles can be subdivided into third order cycles based on their disconformities, spanning an average of ~4 - 5 Ma (Nelson et al., 1995).

The tectonic subsidence curve generated for the Witwatersrand basin clearly implies a foreland basin response which may have been superimposed on an inherited passive margin basin (Allen and Allen, 1991). These sudden changes on the slope of the curve can be interpreted as lithospheric loading due to advancing thrust sheets, resulting in accelerated tectonic subsidence in a foreland basin setting (Klein, 1991b; Nelson et al, 1995).

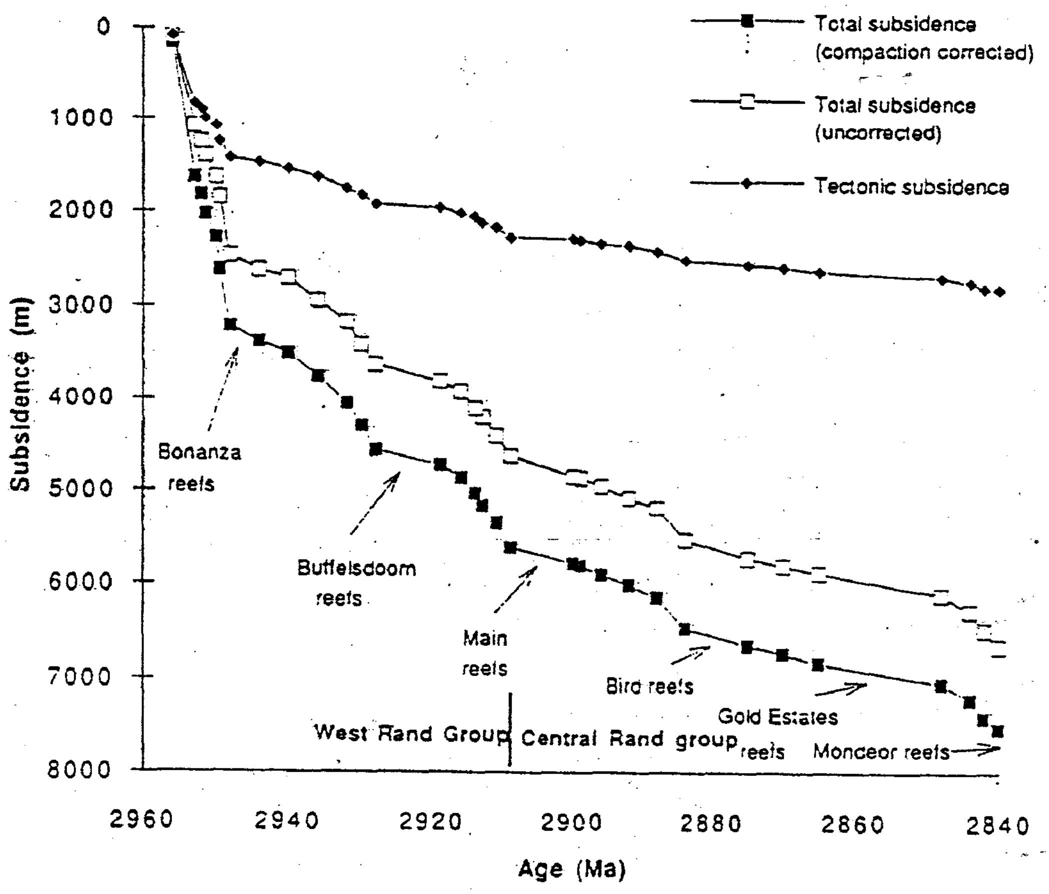


Figure 5.9 - Subsidence curve for the Witwatersrand Basin based on a profile in the Carletonville - Krugersdorp area (From Beukes and Nelson, 1995).

6 SEQUENCE STRATIGRAPHY - AN EXPLORATION TOOL FOR AURIFEROUS PLACER

The ultimate goal in exploration is to be able to recognise the sequence boundaries, the first widespread marine flooding and the maximum marine flooding surfaces in borehole core, geophysical borehole logs and outcrops. This allows for the identification of the systems tracts in both alluvial and marine strata. The next step is to distinguish between a rapid and a slow rate of relative sea - level fall, because these develope two distinctly different unconformities and two types of placers. A Type 1 unconformity results in a deeply incised, channelised placer due to a rapid fall in relative sea - level, whereas a slower and smaller sea - level fluctuation results in a Type 2 unconformity and a widespread sheet - like placer. If these placers are to become economic, the duration of subaerial exposure of the unconformities that allowed the placers to become reworked and concentrated must be determined. In order to preserve the placer, a sudden marine transgression is necessary to allow for minimal shoreline reworking and to cap the placer to prevent it from being dispersed. In the case of deep incised valley during Type 1 unconformity development, a sudden marine transgression will deposit shelf shales and diamictites within the restricted areas of the incised valleys during the early stages of the transgression followed later by more widespread deposition in the interfluvial areas (erosion Channels near the bases of the Johannesburg and Turffontein Subgroups). In the case of a Type 2 unconformity where a sheet - like placer developes on the sequence boundary surface, a sudden marine transgression will be represented by tidally - influenced fluvial deposits immediately above the placer. These deposits consist of well sorted, glassy, white, mature marine orthoquartzites displaying bimodal to unimodal palaeocurrent directions (e.g. Basal/Steyn, Vaal, Carbon Leader, Nigel and Kimberley placers).

Placers are closely associated with the development of disconformities. Therefore the development and prediction of disconformities becomes an important exploration objective (Wescott, 1993). The concepts of sequence stratigraphy provide a powerful tool to identify the unconformity surfaces, along with other surfaces of importance such as maximum flooding surfaces and transgressive surfaces (ravinement surfaces). Transgressive and maximum flooding surfaces are more readily identifiable than subaerial unconformities. The latter are subtle in fluvial strata, but have the greatest economic significance (Galloway, 1989).

In order to apply sequence stratigraphy in Witwatersrand alluvial strata, sequence boundaries must be identified, together with maximum flooding surfaces and eventually systems tracts within these non - marine strata. This allows for the identification of parasequences and parasequence sets within a systems tract.

In some cases a thin succession of alluvial strata have been deposited subsequent to incision into underlying marine strata (shelf shales) where recognition of regionally significant sequence boundaries related to changes in stratigraphic base level are relatively simple. Although alluvial plains in the Witwatersrand basin are sometimes juxtapositioned with paralic sediments, it is not always possible to trace a sequence boundary from marine - marginal sequences to its correlative sequence boundary in the non - marine sequences due to lack of distinctive sedimentary contrast (Shanley and McCabe, 1994).

Therefore one has to rely on stacking patterns of the fluvial quartzites to determine the changes in the rate at which accommodation was created (Figure 6.1). Posamentier and Vail (1988) indicated that the degree of fluvial sandstone amalgamation is controlled by rate at which alluvial accommodation space is created. Sequence boundaries can be recognised by laterally amalgamated sandy to gravelly channel - fill complexes overlying erosional surfaces i. e. Carbon Leader Erosion Channel and Kimberley channels (K8). These multilateral and multistorey channel units suggest low rates of stratigraphic rise, subsequent to channel incision, formed by repeated channel migration and cannibalising most of the finer - grained sediment within the floodplain. Above the Green Bar of the Carbon Leader Erosion Channel, one can expect more isolated meander - belt sand bodies and an increased proportion of mud overlying the widespread laterally amalgamated fluvial sandy to gravelly channel - fill complexes, indicating increasing rates at which accommodation space is created. Recognising these surfaces allows the stratal successions to be subdivided into unconformity bounded units. Other clues to sequence boundaries can also be found in interfluvial areas where careful examination is

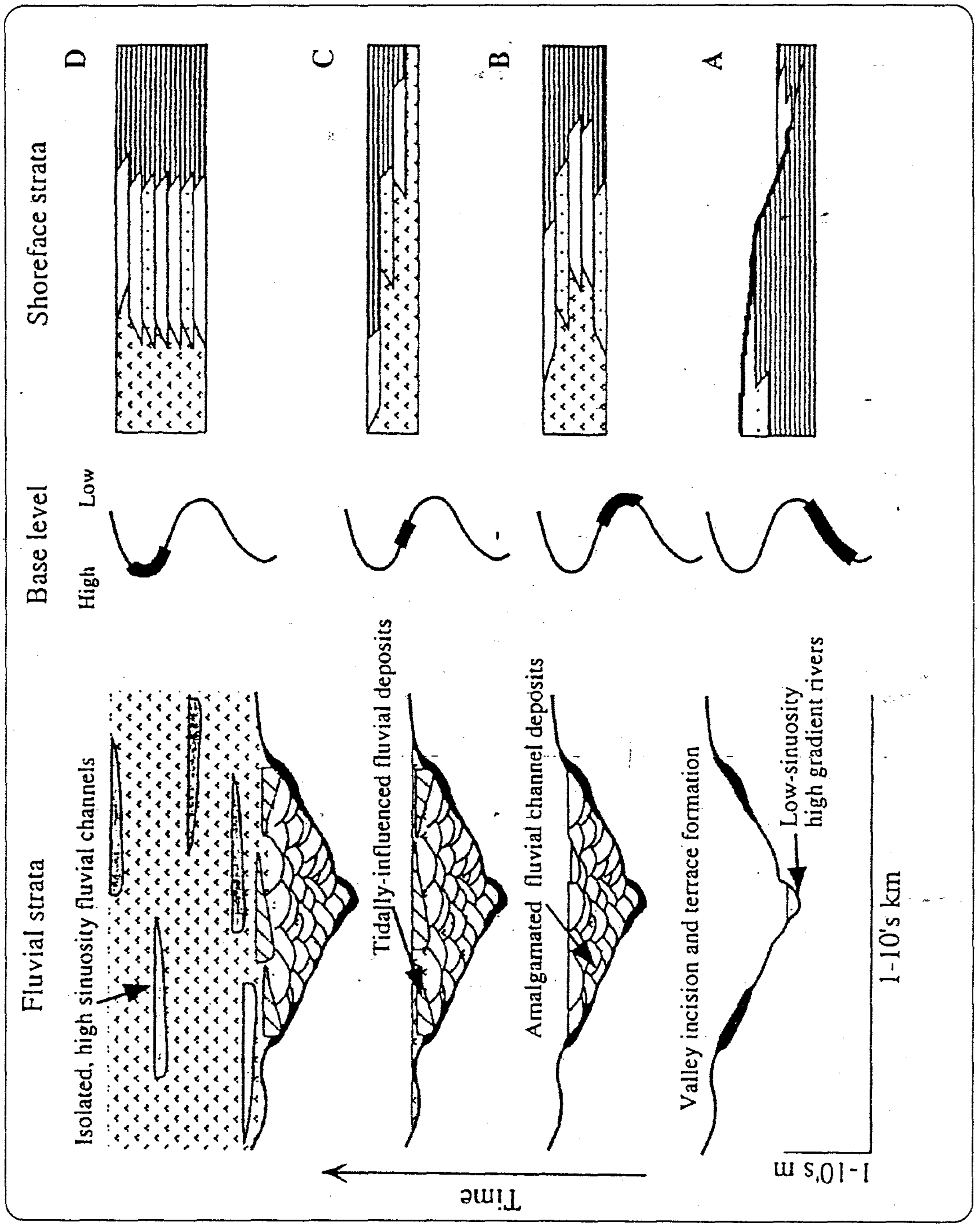


Figure 6.1 - Summary diagram illustrating the relationship between shoreface and fluvial architecture as a function of a slow base - level fall, resulting in low preservation potential of the underlying placer (**A**) Slow rates of base - level rise leading to base - level fall. (**B**) Reduced rates of base - level fall and change to slowly rising base - level. (**C**) Increased rates of base - level rise. (**D**) Reduced rates of base - level rise that are approximately balanced by rates of sedimentation (From Shanley and McCabe, 1994).

needed to identify palaeosols and since these indicate the amount of time of subaerial exposure as well as low sedimentation rates (Shanley and McCabe, 1994).

The period of maximum flooding in alluvial strata can be identified by the presence or invasion of tidal processes into areas previously dominated by fluvial processes (Figure 6.1). These tidal processes are characterised by current reversals, development of fluid - mud or turbidity maxima zones (Beukes, pers. comm.) which deposit clay drapes, rip - up clasts, flaser bedding and inclined heterolithic stratification (Shanley and McCabe, 1994). However, most of these tidal features have not been recognised above the Witwatersrand placers due to ignorance.

In an exploration borehole or outcrop, alluvial lowstand systems tracts can be recognised as sandy to gravelly bed - load deposits characterized by amalgamated, upward - coarsening and thickening channel - fill complexes (Figure 6.1). Transgressive systems tracts of fluvial strata, on the other hand will be dominated by a mixture of bed - load and suspension - load deposits organised as upward fining and upward - thinning bedsets, reflecting high accommodation space due to a rise in stratigraphic base level. During a highstand systems tract accommodation space becomes limited and the systems tract is composed of suspended - load deposits and a greater occurrence of soil profiles (Shanley and McCabe, 1994).

6.1 Exploration Guidelines Based on Deductions Made from Basin Analysis and Sequence Stratigraphic Concepts

After the initial stages of target generation i. e. locating and identifying a potential economic Witwatersrand placer by means of a multidisciplinary geophysical techniques and geological, deep diamond drilling is necessary to test the economic potential interpretation of the selected target area. The ultimate goal in deep diamond drilling exploration for Witwatersrand placers is to be able to recognise the lithologies which are associated with economic gold concentration in the orebody and to substantiate it with gold assay values.

As an exploration guide, the following features are pre - requisites for economic sheet - like placer formation of Witwatersrand type placers:-

- Economic placers should be confined to proximal, tectonically active margins.
- Optimal sheet - like placers are developed at marginal unconformities where they approach conformity.
- Placers should be formed in close association with palaeo - shoreline.
- Slope conditions of the alluvial plain should be either equal to or steeper than the adjacent shallow marine shelf prior to base level lowering.
- Sheet like placers are associated with a Type 2 unconformity. This developed due to a slow rate in relative sea - level fall accompanied with a higher subsidence rate than Type 1 sequences.
- Evidence of prolonged periods of subaerial exposure of the unconformity surface, represented by aeolian deposits, preserved palaeosols and dessication cracks in shelf shales.
- Entrenchment of the fan head, accompanied by sediment reworking on the alluvial fan.
- Evidence of a sudden transgression preserving the underlying placer, with only minor reworking. A tidally - influenced fluvial capping immediately overlies the unit with conspicuous orthoquartzites and sometimes wave rippled siltstone/shales.
- Basin tectonics rather than eustacy controlling the coastal onlap (transgressions) and relative position of the sea level.

Channelised placers are also regarded as suitable targets for gold mineralization and have been exploited successfully, but are less significant in areal extent (West Driefontein and Oryx gold mine).

Correlation of individual placers between each alluvial - fan delta, suggesting that tectonic, base level and climatic events must have had a basin wide effect compiled to a prograding palaeo - shoreline. Each individual alluvial - fan delta would have been fed from a localised, long - lived fluvial feeder system accompanied by an advancing depo - axis. Minter (1990) however, questioned the correlation of placers in the Witwatersrand basin beyond their containing fans, since sediments entering the basin could be destroying the regional chronostratigraphic framework.

This fact is not entirely true since historical mining of these placers has shown that they are laterally persistent beyond the confines of each individual goldfield. A simple fluvial/alluvial fan model is inadequate to account for this lateral continuity.

7 CONCLUSIONS

The Witwatersrand basin is unique in terms of its mineral wealth. The gold in the Witwatersrand is mainly concentrated in placers associated with two types of unconformities. By applying basin analysis and in particular sequence stratigraphic principles accompanied by subsidence analysis, the complex sedimentary basin - fill history of Witwatersrand Supergroup can be unravelled. The critical issues about the Witwatersrand basin which were addresed in this review, is the validity of basin wide correlation of placer unconformuties and whether sequence stratigraphy is applicable to fluvial systems of the Witwatersrand sequence. It is believed that the Central Rand Group was deposited as alluvial - fan deltas by fluvially dominated, braidplain systems with minor marine interaction which had a considerable impact on the preservation of economically viable placers.

Several tectonic models have been proposed for the evolution of the Witwatersrand basin and it seems as if a cratonic foreland basin accounts for many of the observed features observed the Central Rand Group basin.

Primary mechanisms play an important role in the formation of the Witwatersrand basin. They should be studied in conjunction with the sediment infill. These mechanisms include purely thermal mechanics during cooling of the lithosphere as it moves away from the spreading centres controlling the different oceanic bathymetries, changes in the lithosphere/crustal thickness involving the thinning of the crust mechanically by stretching it and loading/unloading of the lithosphere which causes flexuring leading to foreland basin formation like the Witwatersrand basin.

Isostasy becomes an important factor during backstripping calculations when generating a tectonic subsidence curve and therefore rationale behind isostasy has to be understood if sediment is loaded onto the crust/lithosphere. Backstripping procedures can be used to isolate the effects of loading, emphasising the tectonic subsidence history of the basin. In addition, the early part of the evolution of the Witwatersrand basin is controlled by thermal subsidence. This means that part of the Witwatersrand basin thermal subsidence history can be influenced by deeply buried rift basins beneath the passive margin of the early phase of the basin. It is important to understand the extentional history of the protobasinal phase of the Witwatersrand basin. Therefore one has to consider several models for thermal subsidence and crustal extension which should be related to the earlier thermal subsidence history of the basin.

It is likely that the Witwatersrand Basin was superimposed an older, deep seated extensional basin. These compressive tectonics can be superimposed on extensional basins, where the shift from extensional to compressional tectonics lead to inversion processes. Usually these inversion effects are likely to be preserved in sedimentary basins along weakened zones, prone to amplify the tectonic stress change by preserving a thicker sedimentary sequence ie. along the northwestern and western active margins of the Witwatersrand Basin.

Although sequence stratigraphic models have been rigorously applied to marine sequences, they still have to be applied and tested to non - marine, fluvial strata, such as the late Archaean Witwatersrand basin. The failure of the present sequence straigraphic models to accurately predict sedimentary facies and important surfaces in fluvial strata, can be ascribed to inadequacies of modern principles of geomorphology.

Most important to the exploration geologist is the recognition of stacking patterns of the fluvial strata to determine change in the rate at which accommodation was created. Identifying sequence boundaries and other relevant surfaces important for identifying these stacking patterns of the sequences, depends entirely on the recognition of a hierarchy of stratal units including beds, bedsets, parasequences, parasequence sets and the surfaces bounding sequences.

Placers are closely associated with the development of disconformities and therefore become important to recognise in fluvial strata. If these placers are to become economic, the duration of subaerial exposure of the unconformities that allowed the placers to become reworked and concentrated must be determined. In order to preserve the placer, a sudden marine transgression is necessary to allow for minimal shoreline reworking and to cap the placer to prevent it from being dispersed.

The placers in the Witwatersrand basin occur in four major gold - bearing placer zones in the Central Rand Group. Accordingly they can be assigned to four supercycles, which are cyclical and therefore predictive. It is the predictive nature of these rocks and the ability of sequence stratigraphy to enhance this aspect, which is a pre - requisite for an effective exploration tool in the search for new ore bodies or their extension in the Witwatersrand basin.

ACKNOWLEDGEMENTS

The management of Gold fields of South Africa are thanked for allowing me the opportunity to attend this Master of Science in Exploration Geology course at Rhodes University on a full -time basis. Grateful thanks are extended to Prof John Moore and Clyde Mallinson for their support and encouragement during the years proceedings.

I also wish to take this oppertunity of thanking the staff of the Rhodes University Geology Department and my fellow MSc students for their contributions and assistance during the 1995 Exploration Geology course.

Thanks are extended to my colleagues, Mike Buxton and Paul Karpeta for their helpful comments and suggestions throughout the allowed period to complete the dissertation. Their support during this time is highly appreciated.

Most of all, I would like to thank my wife and two sons for their encouragement and support throughout the year, because without such devoted support it would not have been possible to complete the studies.

REFERENCES

Allen, P.H. and Allen, J.R., 1990. Basin Analysis - Principles and Applications. Oxford, Blackwell Scientific Publications, 451.pp.

Antrobus, E.S.A., 1986. Witwatersrand Gold - 100 Years. *Geol. Soc. S. Afr.*, 298pp.

Angevine, C.L. and Heller, P.L., 1987. Quantitative Basin Modelling. Geological Society of America Short Course Notes, 80 pp.

Armstrong, R.A., Retief, E., Compston, W. and Williams, I.S., 1990. Geochronological Constrains on the Evolution of the Witwatersrand Basin, as Deduced from Single Zircon U/Pb Ion Microprobe Studies. *Abstr. Geocongress '90, Geol. Soc. S. Afr.*, Cape Town, 24 - 27.

Armstrong, R.A., Retief, E., Compston, W. and Welke, H.J., 1986. Ages and Isotopic Evolution of the Ventersdorp Volcanics. *Abstr.Geocongress '86, Geol. Soc. S. Afr.*, 89 - 92.

Armstrong, R.A., Compston, W., Retief, E., Williams, I.S., and Welke, H.J., 1991. Zircon Ion Microprobe Studies Bearing on the Age and Evolution of the Witwatersrand Triad. *Precambrian Reasearch,* **Vol 53**, 243 - 266.

Bally, A.W. and Snelson, S., 1980. Realms of Subsidence. In: Miall, A.D. (Ed). Facts and Principle of World Petroleum Ocurrences. *Canadian Society of Petroleum Geologists*, **Memoir 6**, 9 -94.

Barton, E.S., Compston, W., Williams, I.S., Bristow, J.W., Hallbauer, D.K. and Smith, C.B., 1989. Provenance Ages for the Witwatersrand Supergroup and the Ventersdorp Contact reef. Constraints From Ion Microprobe U - Pb ages of Detrital Zircons. *Economic Geology*, **Vol 84**, 2912 - 2019.

Beukes, N.J., 1990. Interaction of fluvial and Marine Sedimentary Processes in the Formation of Some Kerogen - Bearing Auriferous Palaeoplacers of the Witwatersrand Supergroup. *Abstr. Geocongr. '90,* Geol. Soc. S. Afr., Cape Town, 46 - 49.

Beukes, N.J. and Buxton, M., 1991. Genetic Stratigraphy of the West Rand Group, Witwatersrand Supergroup: Annual Report, March 1990 to February 1991. *Anglo American Internal Report*, Johannesburg, 60 pp.

Beukes N.J. & Cairncross, B., 1991. A lithostratigraphic-sedimentological reference profile for the late Archaean Mozaan Group, Pongola Sequence: application to sequence stratigraphy and correlation with the Witwatersrand Supergroup. S. Afr. J. Geol. , **Vol 94** (1), 44 - 69.

Beukes, N.J. and Nelson, J.P., 1995. Sea - Level Fluctuation and Basin Subsidence Controls on the Setting of Auriferous Palaeoplacers in the Archaean Witwatersrand Supergroup: A Genetic and Sequence Stratigraphic Approach. Ext *Abstr. Centennial Geocongr. '95, Vol II, Geol. Soc. S. Afr.*, Johannesburg, 860 - 863.

Bickle, M.J. and Eriksson, K.A., 1982. Evolution and Subsidence of Early Precambrian Sedimentary Basins. *Royal Society (London) Philosophical Transactions*, **Vol 305A,** 225 - 247.

Borchers, R., 1964. Exploration of the Witwaterrand System and its extensions. In: Haughton, S.H. (Ed.). The Geology of Some Ore Dposits in Southern Africa, **Vol I**: Johannesburg, Geol. Soc. S. Afr., 2 5 - 61.

Burke, K., Kidd, W.S.F. & Kudsky, T.M., 1986. Archean basin foreland tectonics in the Witwatersrand, South Africa: *Tectonics*, **Vol 5** (3), 439 - 456.

Christie-Blick, N., Grotzinger, J.P. and Von Der Borch, C.C., 1988. Sequence Stratigraphy in Proterozoic Successions. *Geology*, **Vol 16**, 100 - 104.

Clendenin, C.W., Charlesworth, E.G. and Maske, S., 1988. Tectonic Style and Mechanism of Early Proterozoic Successor Basin Development, Southern Africa. *Tectonophysics*, **Vol 156**, 275 - 291.

Cloetingh, S., 1988. Intraplate Stresses: A New Element In Basin Analasis. In: Kleinspehn, K.L. and Paola, C. (Eds). New Prespective in Basis Analysis. Springer - Verlag, New York, 205 - 230.

Davidson, C.F., 1955. The Mineralization of the Witwatersrand. *Min. Mag.*, **Vol 92,** 152 - 156.

De Wit, M.J., Roering, C., Hart, R.J., Armstrong, A., De Ronde, C., Green, R.W.E., Tredoux, M., Peberdy, E., Hart, R.A., 1992. Formation of an Archaean Continent. *Nature*, **Vol 357**, 553 - 562.

Eidel, J. J., 1991. Basin Analysis for the Mineral Industry. In: Force, E.R., Eidel, J.J., and Maynard, B.J., (Eds). Sedimentary and Diagenetic Mineral Deposits: A Basin Analysis Approach to Exploration. *Reviews in Economic Geology*, **Vol 5**, 1 - 13.

Els, B.G., 1991. Placer Formation During Progradational Fluvial Degradation: The Late Archaean Middelvlei Gold Placer, Witwatersrand, South Africa. *Economic Geology*, **Vol 86**, 261 - 277.

Force, E.R., 1991a. Placer Deposits. In: Force, E.R., Eidel, J.J., and Maynard, B.J., (Eds). Sedimentary and Diagenetic Mineral Deposits: A Basin Analysis Approach to Exploration. *Reviews in Economic Geology*, **Vol 5**, 131 - 139.

-------------- 1991b. Sedimentary Processes as Ore - Forming Processes. In: Force, E.R., Eidel, J.J., and Maynard, B.J., (Eds). Sedimentary and Diagenetic Mineral Deposits: A Basin Analysis Approach to Exploration. *Reviews in Economic Geology*, **Vol 5**, 17 - 20.

Galloway, W.E., 1989. Genetic Stratigraphic Sequences in Basin Analysis I: Architecture and Genesis of Flooding-Surface Bounded Depositional Units. *AAPG Bulletin*, **Vol 73**, (2), 125 - 142.

Grotzinger, J.P., 1986. Evolution of Early Proterozoic Passive - margin Carbonate Platform. Rocknest Formation, Wopmay Orogen, NWT., Canada. *Journal of Sedimentary Petrology*, **Vol 56**, 831 - 847.

Grotzinger, J.P., 1990. The Evolution of Sedimentary Basins. Short Course Notes, Univ of the Randse Afrikaanse, Johannesburg, 85 pp.

Hallbauer, D.K., Klemd, R. and Von Gehlen, K., 1986. A Provenance Model for the Witwatersrand Gold and Uraranium Mineralization and its Implications in the Recognition of Gold - Distribution Patterns in Reefs. *Abstract Geocongress `86*, Johannesburg, Extended Abstracts, 133 - 137.

Hallbauer, D.K. and Barton , M.J., 1987. The Fossil Gold Placers of the Witwatersrand: A Review of their Mineralogy, Geochemistry and Genesis. *Gold Bull.*, **Vol 20**, 68 - 79.

Haq, B.U., Hardenbol, J. & Vail, P.R., 1988. Mesozoic and Cenozoic Chronostratigraphy and Cycles of Sea Level Change. In: Wilgus, C.K., Hastings, B.S., Kendall, C.G. St. C., Posamentier, H.W., Ross, C.A. and Van Wagoner, J.C., (Eds). Sea level changes - An Integrated Approach. *SEPM Special Publication*, **No. 42**, 71 - 108.

Haq, B.U., Hardenbol, J. & Vail, P.R., 1987. Chronology of Fluctuating Sea Levels Since the Triassic. *Science*, **Vol 235**, 1156 - 1167.

Haq, B.U., (1991). Sequence stratigraphy, sea level change, and significance for the deep sea. *Spec. Publs. Int. Ass. Sediment.*, **Vol 12**, 3 - 39.

Harnady, C.J.H. and Stowe, C.W., 1991. The Archaean - Proterozoic Transition: A Review of the Randian erathem in Southern Africa With Reference to the Accretionary Tectonic Evolution of the Kaapvaal and Zimbabwe Provinces. *Chamber of Mines Precambian Research Unit,* Univ. of Cape Town, **Circ 2** , 18 pp.

Hutchinson, R.I., 1975. The Witwatersrand System as a Model for Sedimentation in an Intracratonic Basin. *D.Sc. Thesis (Unpubl.)*, Univ. Orange Free State, Bloemfontein, 255 pp.

Hutchinson, R.W. & Viljoen, R.P., 1988. Re-evaluation of gold source in Witwatersrand Ores: *S. Afr. J. Geol.*, **Vol 91**(2), 157-173.

Jackson, M.C., 1992. A Review of the Late Arcaean Volcano - Sedimentary Dominion Group and Implications for the Tectonic Setting of the Witwatersrand Supergroup, South Africa. *Journal of African Earth Sciences,* **Vol 15**, 169 - 186.

Jervey, M.T., 1988. Quartitative Geological Modelling of Siliclastic Rock Sequences and their Seismic Expression. In: Wilgus, C.K., Hastings, B.S., Kendall, C.G. St. C., Posamentier, H.W., Ross, C.A. and Van Wagoner, J.C., (Eds). Sea level changes - An Integrated Approach. *SEPM Special Publication,* **No. 42**, 1988.

Karpeta, W.P., Gendall, I.R. & King, J.A., 1991. Evidence for marine marginal and submarine canyon sedimentation in the Central Rand Group: Implications for the geometry of the Witwatersrand basin. *Terra Nova Abstr., Conf. of Precambr. Sedim. Basins of Southern Africa,* Sedim. Div., Geol. Soc. S. Afr., Pretoria, Suppl. 3., 16 pp.

Karpeta, W.P., 1994. Sea Level Changes and Placer Formation: The Application of Sequence Stratigraphy to Heavy Mineral Concentrations in Fluvial Settings. *Abstract Volume, Liverpool Sequence Stratigraphy Conference.* High Resolution Sequence Stratigraphy: Innovations and Applications, Univ of Liverpool, 316 - 319.

Kingsley, C.S., 1987. Facies Changes from Fluvial Conglomeratesto Braided Sandstone of the early Proterozoic Eldorado Formation, Welkom Goldfield, South Africa. In: Ethridge, F.G. Flores, R.M., and Harvey, M.D. (Eds). Recent developments in Fluvial *Sedimentology. Spec Publ. Soc. Econ. Paleontol. Mineral.*, **Vol 39**, 389 pp.

Klein, G. de V., 1991a. Rationale for Modern Basin Analasis Applied to Ore Deposits. In: Force, E.R., Eidel, J.J., and Maynard, B.J., (Eds). Sedimentary and Diagenetic Mineral Deposits: A Basin Analysis Approach to Exploration. *Reviews in Economic Geology*, **Vol 5**, 21 - 23.

-----------------, 1991b. Basin - Forming Processes. In: Force, E.R., Eidel, J.J., and Maynard, B.J., (Eds). Sedimentary and Diagenetic Mineral Deposits: A Basin Analysis Approach to Exploration. *Reviews in Economic Geology*, **Vol 5**, 25 - 36.

-----------------, 1991c. Basin Sedimentology and Stratigraphy - The Basin Fill. In: Force, E.R., Eidel, J.J., and Maynard, B. J., (Eds). Sedimentary and Diagenetic Mineral Deposits: A Basin Analysis Approach to Exploration. *Reviews in Economic Geology*, **Vol 5**, 51 - 87.

Klemd, R. and Hallbauer, D.K., 1987. Hydrothermally altered Peraluminous Archaean Granites as a Provenance Model for Witwaterrand Sediments. Mineralium Deposita, **Vol 22**, 227 - 235.

Krapez, B., (1993). Sequence stratigraphy of the Archaean supracrustal belts of the Pilbara Block, Western Australia. *Precambrian Research*, **Vol 60**, 1 - 45.

Mayer, J.J and Albat, H.M., 1988. Tectono - Sedimentary Setting of the Area of the Vredefort Structure During Deposition of the Upper Quartzite Member of the Hospital Hill Subgroup. *S. Afr. J. Geol.*, **Vol 91**, 239 - 247.

Maynard, J.B. and Klein, G. de V., 1995. Tectonic Subsidence Analysis in the Characterization of Sedimentary Ore Deposits: Examples from the Witwatersrand (Au), White Pine (Cu), and Molango (Mn). *Economic Geology*, **Vol 90**, 37 - 50.

McCarthy, T.S., Stanistreet, I.G., Cadle, A.B. and Durrheim, R., 1990. The Geology of the Ventersdorp Supergroup (Bezuidenhout Valley) Outlier to the East of Johannesburg and its Relevance to the Tectonics of the Witwatersrand Basin Margin. *S. Afr. J. Geol.*, **Vol 93 (2)**, 289 - 309.

McKenzie, D.P., 1978. Some Remarks on the Development of Sedimentary Basins. *Earth and Planetary Science Letters*, **Vol 40**, 25 - 32.

Miall, A.D., 1991. Stratigraphic Sequences and Their Chronostratigraphic correlation. *Journal of Sedimentary Petrology*, **Vol 61**, 497 - 505.

Minter, W.E.L., 1978. A Sedimentological Synthesis of Placer Gold, Uranium, and Pyrite Concentration in Proterozoic Witwatersrand Sediments. In: Miall, A.D., (Ed). Fluvial Sedimentology. *Canadian Society of Petroleum Geologists*, Calgary, Alberta, 801 - 829 pp.

Minter, W.E,L., Hill, W.C.N, Kidger,R.J., Kingsley, C.S. and Snowden, P.A., 1986. The Welkom Goldfield. In: Anhausser, C.R. and Maske, S. (Eds). Mineral Deposits of South Africa, **Vol I**. *Geol. Soc. of S. Afr.*. 497 - 539.

Minter, W.E.L., Feather, C.E. and Glatthar, C.W., 1988. Sedimentological and Mineralogical Aspects of a Newly Discovered Witwatersrand Placer Deposit that Reflect Proterozoic Weathering, Welkom Gold Field, South Africa. *Economic Geology*, **Vol 83**, 481 - 491.

Minter, W.E.L., 1990. Palaeoplacers of the Witwatersrand basin. *Mining Engineering*, **A2**, 195 - 199.

Minter, W.E.L., 1991. Ancient Placer Gold Deposits. In: Foster, R.P. (Ed). *Gold Metallogeny and Exploration*. Blackie, London, 283 - 308.

Minter, W.E.L. and Loen, J.S., 1991. Palaeocurrent Dispersal Patterns of Witwatersrand Gold Placers. *Precambrian Research Unit*, University of Cape Town. Circl. No 3, 23 pp.

Mitchum, R.M., Vail, P.R. and Thompson, S. III, 1977. Siesmic Stratigraphy and Global Changes of sea level, Part 2: The Depositional Sequence as a Basic Unit for Stratigraphic Analysis. In: Payton, C.E., (Ed). Seismic Stratigraphy - Applications to Hydrocarbon Exploration. *AAPG*, **Memoir 26**, 53 - 62.

Myers, R.E., McCarthy, T.S. and Stanistreet, I.G., 1990. A Tectono - Sedimentary Reconstruction of the Development and Evolution of the Witwaterarand Basin, with Particular Emphasis on the Central Rand Group. *S.Afr. J. Geol.*, **Vol 93**, 180 - 210.

Nelson, J. P. , Beukes, N. J. & van Eeden, J., 1995. Sequence Stratigraphic and Basin Subsidence Analysis of the Late Archaean Witwatersrand Supergroup. Ext *Abstr. Centennial Geocongr. '95, Vol II, Geol. Soc. S. Afr.*, Johannesburg, 875 - 877.

Nummedal, D., Riley G.W. and Templet, P.L., 1993. High - Resolution Sequence Architecture: A Cronostratigraphic Model Based On Equilbrium Profile Studies. In: Posamentier, H.W., Summerhayes, C.P., Haq, B.U. and Allen, G.P., (Eds). Sequence Stratigraphy and Facies Associations. *Spec. Publs. Int. Ass. Sediment.*, **Vol 18**, 55 - 68.

Olivier, H.J., 1965. The Tectonics of the Upper Division of the Witwatersrand System in the Loraine Area of the Orange Free State goldfield. *Trans. Geol. Soc. S. Afr.*, **Vol 68**, 143 - 175.

Phillips, G.N. & Myers, R.E., 1987. Mineralogy of the Witwatersrand reef components. A review and some implications for exploration. Univ. Western Australia Geology Dept. Univ. Ext. Pub. **11**, 341-354.

Phillips, G.N., Myers, R.E., & Palmer, J.A., 1987. Problems with the placer model for Witwatersrand gold: Geology, **Vol 15**, 1027-1030.

Phillips, G.N. and Myers, R.E., 1989a. The Witwatersrand Gold Fields: Part II. An Origin for Witwatersrand Gold During metamorphism and Associated Alteration. *Economic Geology,* **Monograph 6**, 598 - 608.

Phillips, G.N., Myers, R.E., Law, J.D.M., Bailey, A.C., Cadle, A.B., Beneke, S.D. and Guisti, L., 1989b. The Witwatersrand Gold Fields: Part I. Postdepositional Histor, syndepositional Processes, and Gold Distribution. *Economic Geology,* **Monograph 6**, 585 - 597.

Pitts, P.A., (1990). Aspects of Shear Strain in the East Rand Basin. MSc. Thesis (unpubl), Rand Afrikaans Univ., Johannesburg, 144 pp.

Plint, A.G., Eyles, N., Eyles, C.H. and Walker, R.G., 1992. Control of Sea Level Changes. In: Walker, R.G. and James, N.P. (Eds). Facies Models - Response To Sea Level Change. *Geological Association of Canada*, 15 - 27.

Posamentier, H.W., Allen, G.P., James, D.P. & Tesson, M., 1992. Forced Regressions in a Sequence Stratigraphic Framework: Concepts, Examples and Exploration Significance. *AAPG Bulletin,* **Vol 76** (11), 1687 - 1709.

Posamentier, H.W., Jervey, M.T. & Vail, P.R., (1988). Eustatic Controls On Clastic Depositional I - Conceptual Framework. In: Wilgus, C.K., Hastings, B.S., Kendall, C.G. St. C., Posamentier, H.W., Ross, C.A. and Van Wagoner, J.C., (Eds). Sea Level Changes - An Integrated Approach, *SEPM Special Publication* **No 42**, 109 - 124.

Posamentier, H.W. & Vail, P.R., (1988). Eustatic Controls On Clastic Deposition II - Sequence and Systems Tract Models. In: Wilgus, C. K., Hastings, B. S., Kendall, C.G. St. C., Posamentier, H.W., Ross, C.A. and Van Wagoner, J.C., (Eds). Sea Level changes - An Integrated Approach. *SEPM Special Publication* **No 42**, 125 - 154.

Posamentier, H.W. & James, D.P., (1993). An Overview of Sequence Stratigraphic Concepts: Uses and Abuses. In: Posamentier, H.W., Summerhayes, C.P., Haq, B.U. and Allen, G.P., (Eds). Sequence Stratigraphy and Facies Association. *Spec. Publs. Int. Ass. Sediment.*, **Vol 18**, 3 - 18.

Pretorius, D.A., 1975. The Depositional Environment of the Witwatersrand Gold Field: A chronological Review of Speculations and Oberservations . Minerals Sci. Eng., **Vol. 7,** 18 -47.

Pretorius, D.A., 1981. Gold and Uranium in Quartz - Pebble Conglomerates. *75th Anniv. Vol., Economic Geology*, 117 - 138.

Pretorius, D.A., 1984. The Proterozoic Basin of Southern Africa: Tectonic Settings, Patterns of Evolution, and Distribution of Gold Mineralization. *M.Sc. Mineral. Expl. Course, Course Notes,* Rhodes Univ., Grahamstown, 213 pp.

Pretorius, D.A., Brink, W.C.J. and Fouché, J., 1986. A Map of the Witwatersrand Basin: Surface and Subsurface Geology and Structure. Johannesburg, Geol. Soc. South Africa.

Pretorius, D.A., 1991. The Source of Witwatersrand Gold and Uranium: A Continued Difference of Opinion. *Economic Geology*, Monograph, **Vol 8**, 139 - 163.

Ramdohr, P., 1958. New Observations on the Ores of the Witwaterrand in South Africa and their Genetic Significance. *Trans.Geol. Soc. S. Afr.*, Annex 61, 50 pp.

Reimer, T.O., 1984. Alternative Model of the Derivation of Gold in the Witwatersrand Supergroup. *Geol. Soc. London Jour.*, **Vol 141**, 263-272.

Robb, L.J. & Meyer, M., 1985. The nature of the Witwatersrand hinterland: Conjectures on the source area problem. *Econ.Geology Research Unit Inf. Circ.*Univ. Witwatersrand **178**, 25 pp.

--------------, 1987. The nature of the Archaean basement in the hinterland of the Witwatersrand Basin: Part 1 - The Rand anticline between Randfontein and Rysmierbult: *South African Jour. Geology*, **Vol 90** (1), 44 - 63.

Robb, L.J. and Meyers, F.M., 1990a. The Nature of the Witwatersrand Hinterland: Conjectures on the Source Area Problem. *Economic Geology,* **Vol 85,** 511-536.

Robb, L.J., Davis D.W. and Kamo, S.L., 1990b. U - Pb Ages on Single Dentrital Zircon Grains from the Witwatersrand Basin, South Africa: Constraints on the Age of Sedimentation and on the Evolution of Granites Adjacent to the Basin. *Jour. of Geol.*, **Vol 98**, 311 - 328.

Robb, L.J., Meyer, M., Drenman, G.R. & Ferraz, M.F., 1990c. The distribution of radioelements in Archaean granites of the Kaapvaal craton, with implications for the source of uranium in the Witwatersrand basin: *South African Jour. Geology*, **Vol 93** (1), 5 - 40.

-------------, 1991. Chronological framework for the Witwatersrand Basin and Environs: Towards a Time Constrained Depositional Model. *S. Afr. J. Geol.*, **Vol 94 (1)**, 86 - 95.

Royden, L. Sclater, J.C. and Von Herzen, R.P., 1980. Continental Margin Subsidence and Heat Flow: Important parameters in Formation of Petroleum Hydrocarbons. *AAPG Bulletin*, **Vol 64**, 173 - 187.

South African Committee for Stratigraphy (SACS), 1980. Stratigraphy of South Africa. Part I (Comp. L.E. Kent) Lithostratigraphy of the Republic of South Africa, South West Africa/ Namibia, and the Republics of Bophuthatswana, Transkei and Venda: *Handb. geol. Surv. S. Afr.*, **8**, 690 pp

Sarg, J.F., 1988. Carbonate Sequence Stratigraphy. In: Wilgus, C.K., Hastings, B.S., Kendall, C.G. St. C., Posamentier, H.W., Ross, C.A. and Van Wagoner, J.C., (Eds). Sea level changes - An Integrated Approach, *SEPM Special Publication*, **No. 42**.

Sclater, J.G. and Christie, P.A., 1980. Continental Stretching: An Explanation of the Post - Mid - Cretaceous Subsidence of the Central North Sea Basin. *Journal of Geophysical Research*, **Vol 85**, 3711 - 3739.

Schumm, S.A., 1993. River Response to Base Level Change: Implications for Sequence Stratigraphy. *Journal of Geology*, **Vol 101**, 279 - 294.

Shanley, K.W. & McCabe, P.J., 1994. Perspectives on the Sequence Stratigraphy of Continental Strata. *AAPG Bulletin*, **Vol 78** (4), 544 - 568.

Stanistreet, I.G. and McCarthy, T.S., 1991. Changing Tectono - Sedimentary Scenarios Relevant to the Development of the Late Archaean Witwatersrand Basin. *Jour. of African Earth Science*, **Vol 13** , 65 - 81.

Tankard, A.J., Jackson, M.P.A., Erikson, K.A., Hobday, D.K., Hunter, D.R. and Minter, W.E.L., 1982. Crustal Evolution of Southern Africa; 3.8 Billion Years of Earth History. Springer Verlag, Heidelberg, 523 pp.

Tainton, S., 1994. A Review of the Witwatersrand Basin and Trends in Exploration. *XVth CMMI Congress. SAIMM*, Johannesburg, 19 - 45.

Vail, P.R., Mitchum, R.M. Jr and Thompson, S., III, 1977. Siesmic Stratigraphy and Global Changes of Sea Level, part 3: Relative Changes of Sea Level From Coastal Onlap. In: Payton, C.E., (Ed). Seismic Stratigraphy - Applications to Hydrocarbon Exploration. *Mem. Am. Assoc. Petrol. Geol.*, **Vol 26**, 63 - 81.

Van Biljon , W.J., 1980. Plate - Tectonics and the Origin of the Witwatersrand Basin. In: Ridge, J.W. (Ed.). *Proc. 5th IOGOD Symposium*, **Vol 1**, Schweizerbartsche Verlagbuchhandlung, Stuttgart, 796 pp.

Van Hinte, E., 1978. Geohistory Analysis - Application of Micropaleontology in Exploration Geology. *AAPG Bulletin*, **Vol 82**, 201 - 222.

Van Wagoner, J.C., Posamentier, H.W., Mitchum, R.M., Vail, P.R., Sarg, J.F., Loutit, T.S. & Hardenbol, J., 1988. An overview of the fundamentals of sequence stratigraphy and key definitions. In: Wilgus, C.K., Hastings, B.S., Kendall, C.G. St. C., Posamentier, H.W., Ross, C.A. and Van Wagoner, J.C., (Eds). Sea- level changes - An Integrated approach, *SEPM Special Publications,* **No. 42**.

Van Wagoner, J.C., Mitchum, R.M., Campion, K.M. & Rahmanian, V.D. 1990. Siliciclastic Sequence Stratigraphy in Well Logs, Cores and Outcrops: Concepts for High-Resolution Correlation of Time and Facies. *AAPG Methods in Exploration Series*, **No 7,** 52 pp.

Vermaakt, D.T. and Chunnett, I.E., 1994. Tectono - Sedimentary Process Which Controlled the Deposition of the Ventersdorp Contact Reef Within The West wits Line. *XVth CMMI Congress*, Johannesburg, SAIMM, **Vol 3,** 117 - 130.

Vos, R.G., 1975. An Alluvial Plain and Lacustrine Model for the Precmbrian Witwatersrand Deposits of South Africa. *J. Sediment. Petrol.*, **Vol 45**, 480 - 493.

Walker, R.G., 1992. Facies, Facies Models and Modern Stratigraphic Concepts. In: Walker, R.G. and James, N.P. (Eds). Facies Models - Response To Sea Level Change. *Geological Association of Canada*, 1 - 15.

Watchorn, M.B. and O`Brien, M.F., 1991. The Significance of Marine Modification in Some Witwaterrand Placers - An Example from the Lower Witwatersand West Rand Group. *S. Afr. J. Geol.*, **Vol 94 (5/6)**.

Watso, D.C. and Klein, D de V., 1989. Origin of Cambrian - Ordovician Sedimentary Cycles of Wisconsin Using Tectonic Subsidence Analysis. *Geology*, **Vol 19**, 879 - 881.

Wernicke, B., 1985. Uniform - sense Normal Simple Shear of the Continental Lithosphere. *Canadian Journal of Earth Science*, **Vol 22**, 108 - 125.

Whiteside, H.C.M., Glasspool,K.R., Hiemstra, S.A., Pretorius, D.A. and Antrobus, E.S.A., 1976. Gold in the Witwatersrand Triad. In: Coetzee, C.B. (Ed.). Mineral Resources of the Republic of South Africa. *5th Ed., Pretoria, South Africa, Geol. Surv.*, 39 - 73.

White, N. and McKenzie, D., 1988. Formation of the Steer's Head Geometry of Sedimentary Basins by Differential Stretching of the Crust and Mantle. Geology, **Vol 16**, 250 - 253.

Winter, H de La R., 1987. A Cratonic Foreland Model for Witwatersrand Basin Development in a Continental Back - Arc, Plate Tectonic Setting. *S. Afr. J. Geol.*, **Vol 90** (4), 409 - 427.

Winter, H de La R., 1994. Foreland depobasin results of the Witwatersrand Supergroup in the Rietfontein - East Rand Region: Eustatic Marine Parallels and Tectonic Continental Contrast s around the proximal Rim. *S. Afr. J. Geol.*, **Vol 97 (2),** 119 - 134.

Wescott, W.A., 1993. Geomorphic Thresholds and Complex response of Fluvial System - Some Implications for Sequence Stratigraphy. *AAPG Bulletin*, **Vol 77** (7), 1208 - 1218.

Wood, L.J., Ethridge, F.G. and Schumm, S.A., 1993. The Effects of Rate of Base - Level Fluctuations on Coastal - Plain, Shelf and Slope Depositional Systems: An Experimental Approach. In: Posamentier, H.W., Summerhayes, C. P., Haq, B. U. and Allen, G. P., (Eds). Sequence Stratigraphy and Facies Association. *Spec. Publs. Int. Ass. Sediment.*, **Vol 18**, 45 - 53.

Wright, V.P. and Marriott, S.B., 1993. The Sequence Stratigraphy of Fluvial Depositional Systems: The Role of Floodplain Sediment Storage. *Sedimentary Geology*, **Vol 86**, 203 - 210.

Ziegler, P.A., 1987. Late Cretaceous and Cenozoic Intra - Plate Compressional Deformation in the Alpine foreland - A Geodynamic Model. *Tectonophysiscs*, **Vol 137**, 389 - 420.