

Calcretes are also documented from humid climates (Seminuik & Searle, 1985) however, McCarthy (*pers.comm.*) believes that calcretes can form only in areas of net rainfall deficit, where carbonate may precipitate in the soil and not be redissolved by percolating groundwater.

Palaeontological evidence has also been used to define the climatic setting for the Burgersdorp Formation and as evidence for marked seasonality. Grine *et al.* (1979) used the proposed occurrence of salt glands in the cynodont *Diademodon* as evidence of aridity. Chimsamy & Rubidge (1993) used the presence of alternating layers of fibro-lamellar and lamellated bone tissue of the humerus of the dicynodont *Kannemeyeria* as evidence that the climate during the deposition of the Burgersdorp rocks was seasonal. Evidence of strong seasonality is most commonly preserved as growth rings within the trunks of trees (Bamford *pers.comm.*), however wood from the middle and upper part of the Burgersdorp Formation does not show distinct growth rings. This wood is however exclusively of the podocarp type and podocarp wood is notoriously poor in their preservation of such structure (Bamford *pers.comm.*).

Although the Burgersdorp Formation as a whole may therefore be envisaged to have been deposited under a semi-arid, seasonal climate, stratigraphic studies have shown that subtle differences may be observed throughout the Formation.

a) Lower Burgersdorp Formation

As mentioned previously in this chapter, the sedimentary rocks of the lower part of the Burgersdorp Formation are characterised by laterally continuous lag conglomerates with calcarenitic matrix, which may be traced out over many tens of kilometres. Clasts found in these lags include rounded calcareous and septarian nodules, clay pebbles, fossil bone and coprolites. The bone is often of a fragmentary nature and shows evidence of suncracking and fracturing prior to deposition. The fragmentary and suncracked nature of this fossil material indicates disarticulation and fragmentation of the remains prior to burial, with long periods

of desiccation on the floodplain, or in-channel during periods when the watercourse was inactive. Such deposits are common in arid climates, which favour increased rates of sediment supply, erosion of previous deposits and in-channel aggradation. The thin (<1m) channel sandstones overlying these lag deposits are internally structured almost exclusively by facies Sh. Such a predominance of Sh, which suggests pulsatory discharge typical of semi-arid areas with seasonal flow (Stear, 1980).

Type I palaeosols are restricted to the lower part of the Burgersdorp Formation and are characterised by having: well developed calcic horizons; abundant rhizcretionary layers; deeply infilled desiccation cracks; localised colour mottling; dark reddish brown purple colouration; and a "crumbly" weathering texture imparted by numerous clay lined fissures. These calcic vertisols are common in the lower Burgersdorp and are considered to represent pedogenesis of floodplain sediments under a seasonally semi-arid environment, being very similar to those described from the *Lystrasaurus* Assemblage Zone by Smith (1995).

The rhizcretions in the lower part of the Burgersdorp Formation are generally larger than those in the middle and upper part of the Formation and in places form large scale (up to 5cm diameter) subhorizontal accumulations that show rugose surface textures and elongate linear ridges (Fig.2.13b). These are identical to features that Smith & Kitching (1997) attribute to matted rootlet horizons within the *Tritylodon* Acme Zone, believed to have formed under arid climatic conditions.

Calcareous nodules within the lower part of the Burgersdorp Formation are in most places smooth surfaced and may form isolated sporadic clusters or laterally extensive horizons. The presence of laterally continuous calcic deposits is indicative of high evapotranspiration rates, and a net water deficit in the system (McCarthy, *pers.comm.*).

Quartz pseudomorphs of gypsum rosettes are also known from the lower part of the Burgersdorp Formation (Greenewald, 1996, *pers.comm.*), and the presence of

these features in the lower part of the Beaufort Group has previously been used to imply an arid climate (Keyser, 1966). Experimental studies (Cody & Cody, 1988) have shown that such aggregates form in warm saline muds which contain concentrations of humic compounds. Radial aggregates of barytes, enriched in strontium are also known from the lower part of the Formation in the north of the basin. The mechanism of formation of these aggregates is obscure at present, but they are thought to represent either hydrothermal or spring deposits or *in situ* growth during arid palaeosol development (Retallack, 1990).

Petrographic evidence shows that the sandstones from the lower Burgersdorp Formation plot between the plutonic and metamorphic arid climate fields of Basu (1985) (Fig.4.3). Supportive petrological evidence for an arid climate comes from the presence of abundant fresh plagioclase feldspar (Fig.2.43) within the sandstones and fines of the lower part of the Burgersdorp Formation, which if deposited under humid conditions should show alteration haloes of clay.

The presence of thick haematitic coatings to fossils from the lower part of the Burgersdorp Formation is also suggestive of an arid setting. Haematite tends to form rapidly in arid settings, whereas goethite is the iron form that usually crystallizes under humid conditions. It should be noted however that goethite may alter to haematite through time (Reading, 1996). The small size of all the fossil plant material recovered from the lower Burgersdorp Formation may be used as supplementary palaeontological evidence for a climate that was arid.

All of the above evidence therefore points to an arid climatic regime during the deposition of the lower part of the Burgersdorp Formation, being most similar to the conditions proposed for the underlying Katberg Formation (Hiller & Stavakis, 1984; Groenewald, 1996) and *Lystrosaurus* Assemblage Zone (Smith, 1995)

b) Middle Burgersdorp Formation

The sedimentary rocks of the middle part of the Burgersdorp Formation form the bulk of the descriptions in the literature regarding the Formation and the *Cynognathus* Assemblage Zone. Previously, the consolidated nature of the mudstones, and the presence of calcareous concretions and rosette shaped inclusions have been used to imply of dry conditions (Kitching, 1977).

Although common in the middle part of the Burgersdorp Formation, facies Se₁ has a very different character to that from the lower part of the Formation. In contrast to facies Se₁ in the lower Burgersdorp, Se₁ in the middle part of the Formation often contain almost complete fossil skulls, jaws and skeletal elements, and does not contain reworked *Lystrosaurus* Assemblage Zone clasts. Furthermore, the sandstone matrix is not calcareous. This may imply that this facies accumulated under slightly less arid conditions, with less residence time of fossils on the floodplain.

Palaeosols of Type II and III are common and in places form stacked sets (Fig. 2.19). These soils preserve medium sized rhizcretions, but not the large scale coalesced rugose forms found in the lower Burgersdorp Formation. Infrequent to locally frequent shallow to medium depth mudcrack infills also attest to periodic drying of the floodplain and a lowering of the water table.

In the upper part of the middle Burgersdorp, an increase in the amount and size of fossil plant material occurs, with fossil stem axes up to 20cm wide and fronds of *D. hughesii* up to 60cm in size (Anderson & Anderson 1983). The size and abundance of equisetalean casts of the genus *Calamites* also increases over those found in the lower part of the Formation. These factors, together with an increase in the size and number of large herbivorous genera, may also be indicative of conditions that were slightly less arid than in the lower part of the Burgersdorp Formation.

c) Upper Burgersdorp Formation

In the upper part of the Burgersdorp Formation, especially close to the contact with the Molteno Formation, very little evidence exists for an arid climate. The colour of the sediments is greyish red (5R 3/4), rather than the dark reddish brown (10R 3/4) of the middle Burgersdorp or the very dusky red purple (5RP 2/2) of the lower Burgersdorp Formation. The relative scarcity of calcareous nodules which occur only as laterally discontinuous, isolated smooth surfaced nodules suggests that the climate was only periodically sufficiently arid to promote calcareous development, and that there was sufficient rainfall to counter evapotranspirative loss, such that the water budget was rarely in deficit. This would imply generally wetter conditions with a higher groundwater table.

The occurrence of numerous floodbasin ponds in the upper Burgersdorp, and the lack of well preserved stacked palaeosols is also suggestive of a generally high ground water table, which would have intersected the floodplain surface at topographic lows to maintain semi-permanent to permanent standing bodies of water. This is also evidenced by the coarsening upward nature of a number of the splay sandstones, which show characters of deltaic progradation into standing bodies of water.

Although similar to the palaeosols of the upper part of the middle Burgersdorp, those of the upper part of the Formation exhibit a slightly different character being structured by rooted horizons with pervasive colour mottling, and vertically orientated micritic infill. These mottles form irregularly shaped patches of greenish grey (5G6/1) to light bluish grey (5B7/1) colouration in a sequence of greyish red (5R4/2) to greyish red-purple (5RP4/2) siltstones and mudstones. Slickensided clay lined arcuate fractures also occur, which may be interpreted as "skew planes" (Smith, 1995), formed by the swelling and shrinking of illuvial clays during seasonal fluctuations. These paleosols are interpreted as hydromorphic soils, with well developed "gley-type" mottling and are thought to represent periodic drying of a normally saturated solum (Smith, 1995). The palaeosols of the upper Burgersdorp

Formation are very similar to those described by Smith (1995) for the *Dicynodon* Assemblage Zone, which he interpreted as hydromorphic gleys, formed under wet conditions.

Petrological evidence also suggests that the climate was less arid. Figure 4.3 shows that the sandstones of the upper Burgersdorp (upper hatched area) plot just below the metamorphic source, humid climate field of Basu (1985). The overall decrease in feldspar, and the decomposed nature of the feldspar present is also more indicative of a sub-humid setting.

The upper Burgersdorp Formation also evidences a change in the palaeontological signature of the Formation, enclosing the subzone C fauna, and a more abundant and larger palaeoflora. Although not really indicative of any particular palaeoclimatic setting, this would suggest that the palaeoenvironment had changed from that of the middle part of the Formation.

d) Conclusion

To judge from the unstratified nature of the dark red mudstones and the lack of green coloured fines and the abundance of calcareous concretions and rhizocretions, the palaeoclimate of the lower part of the Burgersdorp Formation may have been sufficiently arid to warrant the term semi-desert. This conclusion is augmented by petrographic evidence and framework grain plots, which show that the sandstones of the lower part of the Burgersdorp Formation plot in the arid metamorphic source fields of Basu (1985) (Fig.4.3) and are more similar to those of the Katberg Formation. The up section decrease in: the frequency of palaeosol development; desiccation cracks; calcareous glaebules and rhizocretions; as well as the notable colour change probably indicate more poorly drained soils with a higher water table during late Burgersdorp times. This would seem to indicate a change from arid conditions in the lower part of the Formation through to less arid, more equable, and less restrictive climates in the upper part of the Formation. This change from an arid to more humid setting is also evidenced by the petrography of

the sandstones, which change from an arid to sub-humid source (Fig.4.3).

4.1.2.2 Molteno Formation

The climate for the Molteno Formation has in the past been envisaged as ranging from cold and wet with intermittent rainfall (Haughton, 1924; Anderson & Anderson, 1970; Visser, 1991; MacDonald, 1993) to humid temperate (Ellenberger *et al.*, 1964). It has even been proposed that temperature fluctuations were very severe, with palaeofrosts and palaeotemperatures as low as -30°C (Cairncross *pers.comm.*). The presence of the fern *Dicroidium* in the Molteno Formation has in the past been used to imply palaeoclimatic conditions ranging from cold temperate (Anderson & Anderson, 1979) to warm and humid (Rayner *pers.comm.*). *Dicroidium* is however known to occur in strata of Early to Late Triassic age (Anderson & Anderson, 1983; Taylor & Taylor, 1993) in deposits that range from arid to tropical settings (Retallack, 1990). Although detailed study of fossil plants may allow for refined palaeoclimatic reconstructions of the Molteno Formation, the use of *Dicroidium* must therefore be seen as limited. The proposed cold climate in the Karoo Basin during the Middle and Upper Triassic (Visser, 1991) is however anomalous to the global pattern (Frakes, 1979). Bearing in mind the two extremes proposed it is necessary to assess these proposals, based on the sedimentological and palaeontological evidence provided by the rocks of the Molteno Formation.

a) Bamboesberg and Indwe Members

The: reddish colouration of some of the fines; amount of haematitic alteration of plant material; presence of weakly developed iron rich palaeosols; and desiccation cracks in the lowermost part of the Bamboesberg Member suggests local oxidising conditions and limited subaerial exposure. This fact may in part account for the lack of coal in the lower part of the Bamboesberg Member. The dark colouration of the fines in the upper Bamboesberg and Indwe Sandstone Members, coupled to the presence of: reduction spots in the sandstones; diagenetic pyrite; and organic carbon, tends to suggest a reducing environment during the deposition of these

units. The scarcity of pedogenic carbonate, desiccation cracks and the nature of the preserved palaeosols and coals in the upper Bamboesberg and Indwe Sandstone Members is also suggestive of a high groundwater water table, with depositional sites that were permanently saturated, and a system with sufficient rainfall to prevent an evapotranspiration imbalance.

Seegard (*pers.comm.*) provides transport co-efficients from mountain belts of 100m^2 for arid climates and some 5000m^2 for humid climates. The extensive spatial distribution of the preserved fluvial system for the Bamboesberg and Indwe Sandstone Members would therefore tend to be more likely with a humid rather than an arid climate and point to them being of the humid fan, rather than arid fan or proglacial outwash type, as proposed earlier in this chapter. The abundance of coal and vegetal material may be used as supportive evidence for a humid setting (Cadle *pers.comm.*). Thin section petrography and framework grain plots fall within the humid metamorphic source fields of Basu (1985) (Fig.4.3) and are used as supplementary evidence for a humid setting.

The two genera of wood found in the Molteno, *Araucarioxylon* and *Rhexoxylon* are both assigned to the Family Araucariaceae (Bamford, *pers.comm.*) which were common in the middle latitudes with their optimal development between 40° and 50° south and north (Stockey, 1990). Araucarian forests are also known from red bed facies indicative of a low-latitude belts of summer dry climates. Stockey (1990) furthermore states that there is no evidence of a plant assemblage more thermophilous than the araucarian forest. The presence of distinct growth rings in *Araucarioxylon* is firm evidence for marked seasonality during the deposition of the lower part of the Molteno Formation (Anderson, 1974). Growth most likely occurred during the wet seasons and torpor during the drier seasons. Seasonality is further evidenced by the cyclic nature of leaf fall and the vast amounts of monogenetic foliage in restricted layers.

From the above it is evident that of the two proposed palaeoclimates for the Molteno Formation, those favouring a warm humid climate are more probably

correct. Based on the evidence presented here it seems that the anomalous cold climate for the Molteno proposed by Visser (1991) did not exist and, as proposed by Cadlo *et al.* (1993), the overall trend for the Karoo Basin during the Triassic (Fig.4.4) does not differ significantly from the global trend proposed by Frakes (1979).

Based on the available sedimentological and palaeontological evidence an overall warm and humid climate is therefore envisaged for the lower part of the Bamboesberg Member with periods of water table fluctuation, as evidenced by limited palaeosol development, rare dusky red (5R 3/4) coloured sediments and equally rare desiccation cracks in the lacustrine mudstones. The upper Bamboesberg and Indwe Sandstone Members are envisaged as having formed under a warm humid climate with high water tables which allowed for the accumulation of thick layers of plant material in standing bodies of water, and the formation of peat swamps.

4.2

Correlation

Due to the scarcity of radiometric marker horizons, the lack of seismic data, and the extensive remagnetization caused by the Drakensburg volcanics (Ballard *et al.*, 1986), biostratigraphy coupled with accurate lithostratigraphy is the most reliable method for subdividing the Karoo Supergroup, and for constraining the relative time intervals involved.

4.2.1

Burgersdorp Formation

4.2.1.1

Interbasinal correlation

Interbasinal correlation allows for the relative age of the faunal assemblages to be determined by comparison with better dated sequences from other parts of Gondwana, Russia and China. Such correlations also show whether or not the area in question was an isolated basin, with an endemic fauna, or if there was free

passage between basins.

Capitosaurid amphibians are known from both marine and nonmarine deposits and provide the best correlative link between the marine and nonmarine realm. Kannemeyeriid dicynodonts are documented from all the Gondwanan continents, as well as Europe, Russia and China, and therefore have great potential in correlating nonmarine strata. For this reason, and due to the fact that the Triassic dicynodonts were widely studied during the course of this investigation, they are one of the key forms used for the correlations that follow. Two different methods of correlating may therefore be used for relative age determinations. The first of these is based on amphibians and the standard marine timescale and the second on kannemeyeriids and the nonmarine record.

Both capitosaurid amphibians and kannemeyeriid dicynodonts are thus important for the global correlation of the nonmarine Triassic. The faunas from Russia, India and South Africa contain both forms, whilst those from China and South America are rich only in dicynodonts, and those from Europe and Australia are dominated by amphibians, and have very few dicynodonts (Ochev & Shishkin, 1989).

The *Cynognathus* Assemblage Zone has in the past been broadly correlated to the Omingondo Formation of Namibia (Piekford, 1996), the Upper Fromou Formation of Antarctica (Himmer, 1963), the Puesto Viejo (Bonaparte, 1966a,b, 1971) and Rio Mendoza Formations of Argentina (De Juw, 1993), the Yerrapalli Formation of India (Chowdhury, 1970), the Donguz Group/Svitas of Russia (Ochev & Shishkin, 1989) and various formations in China (Cheng, 1980, 1981). As per previous workers, the *Cynognathus* Assemblage Zone is here shown to contain a fauna that is broadly correlatable with the: Omingondo Formation of Namibia; Manda, Kingori and N'tawero Formations of East Africa; Yerrapalli Formation of India; Upper Fromou Formation of Antarctica; Puesto Viejo and Rio Mendoza Formations of Argentina; Ermaying, Kalamayi and Heshangou Formations of China and Inner Mongolia; and the Donguz and Vetluga Groups/Svitas of Russia (Table 4.1).

Apart from Watson (1942) and Cooper (1982), most previous workers have failed to recognise the presence of different faunal assemblages within the *Cynognathus* Assemblage Zone, and for the purpose of correlation equate the whole of the zone with other Gondwanan faunas. The subdivision of the *Cynognathus* Assemblage Zone (Hancox *et al.*, 1995), coupled to the findings of this thesis allow for more refined correlations for the biozone.

Because many of the Gondwanan, Chinese and Russian sequences have not in the past been subdivided, nor are they clearly understood, a brief review of their biostratigraphy is necessary prior to correlation with the proposed subzones of the *Cynognathus* Assemblage Zone. It must be stressed that the following correlations are based on: personal observations of the Triassic dicynodonts; discussions with other scientists; and on the current state of the literature, which in many cases requires major taxonomic and biostratigraphic revision.

a) Namibia

In Namibia, Triassic sedimentary strata have so far only been described from the northern part of the country in the Otjiwarongo, Grootfontein and Omaruru Districts (Pickford, 1996), where they are known as the Omingonde Mudstone Formation. Keyser (1973) considered the fauna of the Omingonde Formation to be advanced over the typical *Cynognathus* Assemblage Zone fauna, and therefore correlated this unit with the Molteno Formation. The rocks of the Omingonde Formation however actually contain an assemblage of fossils closely related to those from the *Cynognathus* Assemblage Zone, including: the dicynodonts *Kannemeyeria cristatyrhynchus* and *Dolichuranus primaveus*; the trirachodontid cynodont *Trirachodon* sp; the diademodontid cynodonts *Diademodon tetragonus* and *Titanogomphodon crassus*; and a small lower jaw of an unidentified cynognathid that resembles *Cynognathus* (Keyser, 1973). Other genera include the bauriids *Sesamodon bauri* and *Herpetogale marsupialis*, an eriopid amphibian and a large erythrosuchid archosauriform cf. *Erythrosuchus africanus* (Pickford, 1996).

The exact taxonomic affinities of several of these forms are unfortunately unclear at present. *Kannemeyeria cristarhynchus* (GSP R313) was originally assigned to *Kannemeyeria simocephalus* (Keyser, 1973), but re-assigned to *K. cristarhynchus* by Keyser & Cruickshank (1979). The specific differences between these two forms are however poorly known, and the actual specific assignment of GSP R313 remains somewhat problematical. *Titanogomphodon* is considered to be an advanced diademodontid based on its overall size and the shape of its moliform teeth (Keyser, 1973), although Brink (1980) suggests that this form is congeneric with *Diademodon*. The species of *Trirachodon* is similar to *T. berryi* from subzone B, although it apparently lacks a parietal foramen, has a narrower intertemporal crest, and a large diastema in both the upper and lower jaws (Keyser, 1973). The most recent collecting in the Omingondo Formation has provided new specimens of *Trirachodon* which are larger than their South African counterparts (Smith *pers. comm.*), however their exact taxonomic position however awaits clarification.

The presence of *Kannemeyeria*, a large erythrosuchid c.f. *Erythrosuchus africanus*, *Diademodon* and a form similar to *T. berryi* indicates a close temporal relationship for at least parts of the Omingondo Formation and subzone B of the *Cynognathus* Assemblage Zone. The presence of several advanced cynodonts and the dicynodont genus *Dolichuranus* may be used to propose a slightly younger age for certain levels of the Omingondo Formation. An isolated intertemporal bar (BP/1/5536) from subzone C is identical in form and size to that of *Dolichuranus* and further collecting may show that this genus, or a very close relative occurs in subzone C, thereby allowing the direct correlation of subzone C with the *Dolichuranus* bearing levels of the Omingondo Formation.

Of the genera known from the Omingondo Formation, five are common to subzone B and for the most part this fauna may be directly correlated. The presence of *Dolichuranus* and *K. cristarhynchus*, which are believed to come from the upper part of the Formation (Keyser *pers. comm.*), may be indicative of a slightly younger age than subzone B and parts of the Omingondo Formation may represent a timespan intermediate between subzones B and C. No genera except *Diademodon* are

however shared with subzone C and the specific differences assigned to the Omingondo genera may represent isolated speciation in a sub-basin. The current stratigraphic resolution of the Omingondo Formation is not accurate enough to allow for the possibility of subdividing the sequence, although the presence of a mixed subzone B and advanced fauna suggests that this may in fact become possible in the future.

b) East Africa

Sedimentary strata containing a Karoo fauna in East Africa have been known since the investigations of Houghton (1932). Subsequent collections and descriptions by workers such as Dixey (1936), Nowack (1937), von Huene (1942), Boonstra (1953), Attridge *et al.* (1963), Drysdall & Kitching (1962, 1963), Cruickshank (1965, 1967), Bishop (1968), Cox (1969, 1991) and Kemp (1975) have greatly expanded our knowledge of these deposits and their fauna. The Kingori (K7) and Manda Formations (K8) of Tanzania and the N'tawere Formation of Zambia all contain faunal assemblages which may be correlated to the *Cynognathus* Assemblage Zone.

The presence of *Kannemeyeria* (Cruickshank, 1965) and *Diademodon rhodesiensis* (Brink, 1963) in the Kingori Sandstone Formation (K7) of Tanzania allows for the direct correlation of this fauna with subzone B of the *Cynognathus* Assemblage Zone. *Kannemeyeria* is known from only a single specimen that was originally documented as having been collected in the Manda Formation, but was subsequently re-assigned to the Kingori Formation by Cruickshank (1965). Although originally assigned to the now abandoned species *K. latirostris* (Cruickshank, 1965), this specimen has been studied by the author and has no characters that preclude it from *K. simocephalus* except for its size.

The overlying Manda Formation (K8) contains a fossil fauna which includes the: dicynodonts *Angonisaurus* (Cox & Li, 1983), *Rechnisaurus* (Cox, 1991), *Tetragonias* (Cruickshank, 1967) and *Sengusaurus* (Cruickshank, 1986);

capitosaurid amphibian *Parotosuchus pronus* (Howie, 1970); rhynchosaurian *Stenaulorhynchus* (Von Huene, 1938a); archosauriform *Mandasuchus* (Von Huene 1938); and trirachodont and traversodont cynodonts (Crompton, 1955, 1969)

The presence of *Angonisaurus* in both the Manda Formation and subzone C of the *Cynognathus* Assemblage Zone allows for their direct temporal correlation. This correlation is strengthened by the occurrence of the capitosaurid *Parotosuchus pronus* in the Manda, which is very similar to *Parotosuchus morgani* sp. nov. At the Family level the occurrence of the shansiodont genus *Tetragonias* may be used as corroborating evidence, as *Tetragonias* is closely related to the new Shansiodontid (BP/1/5532) from subzone C. The two isolated cylindrical caniniforms (BP/1/5626 & BP/1/5627) from subzone C are also identical in construction to that of *Tetragonias*. According to Dilkes (1994) *Stenaulorhynchus* is advanced over the rhynchosaurids from subzone B. The Manda Formation therefore shares two genera, as well as one family level tie with subzone C, and for this reason parts of the Manda may be correlated to subzone C with some confidence. This fauna however lacks cynognathid and diademodontid cynodonts which are common in subzone C and may therefore be slightly younger in places.

In Zambia, the N'tawere Formation may be subdivided into an Upper and Lower unit. The Lower N'tawere contains the dicynodont genus *Dolichuranus latirostris* (Crozier, 1970), the cynodonts *Trirachodon*, *Diademodon* and a *Scalenodon*-like form, as well as the amphibian genus *Parotosuchus megarhinus* (Chernin 1974) and the invertebrate *Unio kareoensis* (Drysdall & Kitching, 1963).

The presence of *Dolichuranus* and a gomphodont cynodont provisionally assigned to *Scalenodon* would allow a close correlation between the Lower N'tawere and the Omingonde Formation, and by extrapolation, subzone B of the *Cynognathus* Assemblage Zone. It should be noted however that *Dolichuranus latirostris* (BP/1/3636) bears little resemblance to the Omingonde species *D. primaveus* and *D. etienensis*, and rather shares a number of characters with the shansiodont genus *Tetragonias*.

Based on the presence of *Sangusaurus* (Cox, 1969) in both faunas, the Upper N'tawere is often correlated with the Manda Formation. Although the Upper N'tawere and the Manda both contain the genus *Sangusaurus*, the Manda is usually considered to be slightly younger due to the fact the *Sangusaurus paringtoni* from the Manda is believed to be advanced over *Sangusaurus edentatus* from the N'tawere (Cruickshank, 1986a). The Upper N'tawere also contains the dicynodont genus *Zambiasaurus* which is considered a primitive Stahleckeriid (Cox, 1969), the cynodont *Trirachodon* and the temnospondyl amphibian *Parotosuchus megarhinus*. *P. megarhinus* is very similar to *P. morgani* sp. nov. from subzone C and based on the similarities of the fauna, the majority of the Upper N'tawere is correlated to subzone C of the *Cynognathus* Assemblage Zone.

c) Antarctica

An 8.0m thick sandstone at the base of the upper member of the Fremouw Formation in the Beardmore Glacier region (Gordon Valley, Transantarctic Mountains) contains a fauna comparable with that of the *Cynognathus* Assemblage Zone (Hammer, 1988, 1990, 1995; Hammer *et al.* 1986, 1987, 1990, 1995). The assemblage is poorly known at present, but contains the cynodont *Cynognathus* sp. and a diademodontid possibly allied to *Titanogomphodon* (Hammer, 1995). The material assigned Kannemeyeriidae *incertae sedis* by Hammer (1995) seems to be similar to the *Rechisaurus* and *Kannemeyeria christarhynchus* in that the canine erupts quite medially on the stout caniniform process. Representatives of the amphibian Families Capitesauridae and Benthosuchidae are also known from this locality but are not as yet described.

Hammer (1990) notes that the diademodontid and a number of other forms are more advanced than the traditional *Cynognathus* Assemblage Zone fauna. At present this fauna cannot be directly compared to the *Cynognathus* Assemblage Zone, but based on the possible presence of *Titanogomphodon* (= *Diademodon*) and a kannemeyeriid similar to *Rechisaurus* this fauna may be correlated with parts of the Omingonde and probably equates either to the uppermost reaches of subzone

B or the lower part of subzone C.

d) India

The Yerrapalli Formation of the Pranhita-Godavari Valley of India (Jain *et al.*, 1964; Chowdhury, 1970a,b) contains a fauna including the large capitosaurid amphibian *Parotosaurus rajareddyi*, a possible brachyopid amphibian, erythrosuchid archosauriformes (cf. *Erythrosuchus*, Cooper, 1980), the rhynchosaur *Mesodapardon kuttyi*, indeterminate trirachodontid cynodonts and the kannemeyeriid dicynodonts *Wadiasaurus incus* (Chowdhury, 1970b) and *Rechnisaurus cristarhynchus* (Bandyopadhyay, 1989). Based on the evolutionary stage of the capitosaurid amphibians and kannemeyeriid dicynodonts, this fauna is slightly younger than subzone B and may be tentatively correlated with subzone C, although no genera are common to the two.

e) South America

The fossils of the Triassic sedimentary strata of South America have been known since the collecting of Von Huene (1935) and Cabrera (1943) and contain faunas broadly similar with those of other parts of Gondwana (Cox, 1965, 1968; Bonaparte, 1967; Barberena, 1977, Barberena *et al.*, 1985).

The Puerto Viejo Formation (Mendoza Province, Argentina) comprises a 325m thick sequence of mudstones and sandstones which overlies the Permian-Triassic volcanics of the Cerro Carrizal Group. The Puerto Viejo may be sub-divided into a lower (Agua de Los burros local fauna) and upper sequence. The lower sequence, proposed to be a lowland assemblage (Anderson & Anderson, 1993a), contains the shansiodont genus *Vinceria* sp. indet. (Bonaparte, 1969, 1973, 1982) and the kannemeyeriid *Rechnisaurus* (DeFauw, 1993). Bonaparte (1982) correlates the Agua de Los burros local fauna with the *Lystrosaurus* Assemblage Zone, however the presence of *Rechnisaurus* in this fauna shows that it is actually significantly younger (DeFauw, 1993).

The upper sequence is characterised by the presence of the dicynodont "*Kannemeyeria*" *argentinensis*, the shansiodont *Vinceria andina* (Bonaparte, 1966, 1967, 1982), the cynognathid cynodont *Cynognathus minor*, the traversodontid cynodonts *Pascualgnathus polanskii* (Jenkins, 1970), and a proposed galesaurid *Cromptodon mamiferoides* (Bonaparte, 1972). On the basis of the presence of *Kannemeyeria*, *Cynognathus* and diademodontids, Bonaparte (1966a,b) correlated the upper beds of the Puesto Viejo Formation with those of the *Cynognathus* Assemblage Zone.

The presence of *Kannemeyeria* and a proposed galesaurid in the upper zone is however problematical, as galesaurids are last known in the *Lystrosaurus* Assemblage Zone in South Africa (Kitching & Groenewald, 1995). *Kannemeyeria argentinensis* is not well known and comes from the same horizon as *Vinceria andina*. Although first hand study of this material was not possible, photographs kindly supplied by Arthur Cruickshank and Barry Cox, show that certain of the features described by Bonaparte (1966a) are incorrect. The most obvious of these is that the postorbitals terminate on the skull roof approximately half way along its length and therefore do not reach to the back of the intertemporal bar as figured by Bonaparte (1966a,b). This is exactly the state found in the shansiodonts *Shansiodon* and *Tetragonias* and for this reason it seems more likely that "*K*".*argentinensis* belongs to the Family Shansiodontidae, and may in fact be congeneric with *Vinceria*.

The two faunas from the Puesto Viejo are therefore more similar than previously thought. The basal part of the upper assemblage, containing as it does the first appearance datum (FAD) of the shansiodont *Vinceria* and the cynodont *Cynognathus*, may be directly correlated to subzone C of the *Cynognathus* Assemblage Zone. The uppermost parts of the sequence record the FAD of primitive traversodontid cynodonts and may therefore be slightly younger than the subzone C fauna, in which traversodontids are unknown. The presence of a proposed galesaurid remains problematical.

The Rio Mendoza Formation (Mendoza Province, Argentina) also contains the shansiodont *Vinceria andina* and the traversodontine cynodonts *Ruscocynodon* and *Andescynodon* and is usually correlated with the Upper Puesto Viejo Formation (Battail, 1993). This fauna may therefore also be tentatively correlated to subzone C.

f) China

Nonmarine Triassic strata are widespread in northern China, occurring in two distinct areas, in a number of separate basins. In the northwest part of the country the Kalamay Formation occurs in the Junggur and Turpan (Tulufan) Basins (Xinjiang Autonomous Region). The Ordos Basin contains the Heshangou and the Upper and Lower Ermaying Formations, and occurs in the Shanxi, Gansu and Ningxia provinces in the central part of the country.

The Kalamay Formation (variously referred to as the Karamayi, (Li, 1980); Karamay (Pang, 1993) and Kelamayi (Lucas, 1993b)) forms part of the lower Xiaoguangou Group. In the Turpan Basin, this Formation contains amphibian material originally assigned to *Parotosuchus turfanelis* (Young, 1966) but re-assigned capitosaur indet by Lucas & Hunt (1993a). In the Junggur Basin the Formation contains the dicynodont *Parakannemeyeria brevirostris*. Pang (1993) assigned an archosauriform from this Formation to the genus *Garjainia*. The correlation of this Formation with the *Cynognathus* Assemblage Zone is therefore problematical, as it contains one representative of subzone A, and a dicynodont fauna which has no equivalent in South Africa. At present then this Formation cannot be correlated to the sequence in the main Karoo basin with any confidence.

In Inner Mongolia, in the Shaan-Gan-Ning Basin (Fugu County, Shaanxi Province) the upper Heshangou Formation has yielded the: dicynodont genus *Shaanbeikannemeyeria* (Li, 1980); proclophoniid genera *Eumetabolodon* (Li, 1983), *Pentaedrusaurus ordosianus* (Li, 1989); dipnoan *Ceratodus heshangouensis*; and the archosauriform *Fugusuchus* (Pang, 1993). *Shaanbeikannemeyeria* is thought to be

very closely related to, and possibly congeneric with *Kannemeyeria* (King, 1988) and *Ceratodus heshangouensis* is believed to be indistinguishable from *Ceratodus* material from subzone B (Shishkin *pers.comm.*). This would suggest a strong correlative link between the Heshangou Formations and subzone B of the *Cynognathus* Assemblage Zone. The presence of *Fugusuchus* is somewhat problematical as most authors (eg Pang, 1993) equate this genus to *Garjainia*, and if this is the case, then parts of the Heshangou Formation may be correlatable to subzone A. At present and until the stratigraphy of this unit is better documented, the Heshangou Formation is therefore equated to both subzone A and the lower parts of subzone B of the *Cynognathus* Assemblage Zone.

The Ermaying Formation occurs within the Ordos Basin in the eastern central parts of China and has been biostratigraphically subdivided into an upper and lower part based on the presence of two distinct fossiliferous horizons (Cheng, 1980; Li & Cheng, 1995). The Ermaying has in the past been equated with the *Cynognathus* Assemblage Zone (Sun, 1963, 1981), and placed as slightly older (Keyser & Cruickshank, 1979) and slightly younger (Cooper, 1982) than the biozone.

The faunal assemblage of the Lower Ermaying includes the dicynodonts *Shaanbeikannemeyeria* (Cheng, 1980; Li, 1980) and *Parakannemeyeria xingxianensis* (= *P. ningwuensis*), the bauriid *Ordosia*, a scapulasaurid, and primitive thecodonts. Three teeth assigned to the procolophonid genus *Paotodon* are also recorded (Chow & Sun, 1960), as is the procolophonid *Eumetabolodon*. The presence of *Shaanbeikannemeyeria* in this assemblage allows for its tentative correlation with subzone B.

The fauna of the upper Ermaying Formation contains three species of the dicynodont genus *Sinokannemeyeria* (Young, 1937; Sun, 1963; Cheng, 1980), as well as five species of *Parakannemeyeria* (Sun, 1963, 1978; Cheng, 1980) and four species of *Shansiodon* (Yeh, 1959; Cheng, 1980), including *S. wuhsiangensis* (Yeh, 1959). The fauna also includes the trirachodontid cynodont *Sinognathus*, the archosauriforms *Shansisuchus* and *Wangisuchus*, the captorhinid *Neoprocolophon*,

and the large bauriid *Traversodontoides* (Sun, 1981; Li & Cheng, 1995). Sun (1963, 1981) has previously correlated the Upper Ermaying Formation with the *Cynognathus* Assemblage Zone, however the presence of *Shansiodon wuhsiangensis* allows for a direct correlation to be proposed between parts of the Upper Ermaying and subzone C of the *Cynognathus* Assemblage Zone. The presence of *Shansisuchus* which is very similar in nature to *Erythrosuchus*, may indicate that parts of the upper Ermaying are of an equivalent age to subzone B as well.

g) Russia

In Russia, the Vetluga (Zone V) and Baskunchak series are known as the *Parotosuchus* fauna (Table 4.2) and are known to contain the amphibian genera *Parotosuchus*, *Trematosuchus*, *Batrachosuchus* and *Neorhaciotomi*, as well as an associated fauna including *Garjainia* (= *Vjushkovia*) (Ochev & Shishkin, 1989) and nine genera of captorhinids (procolophonids); *Phaanthosaurus*, *Tichvinskia*, *Burtensia*, *Contitosaurus*, *Orenburgia*, *Coelodontognathus*, *Kapers*, *Macrophon* and *Vitalia*.

Shishkin (*pers.comm.*) considers the European genus *Parotosuchus* to be very closely related to *Kestrosaurus* from subzone A, as is *Trematosaurus* to *Trematosuchus* (Shishkin & Welman, 1995). Coupled with the presence of *Batrachosuchus*, the primitive archosauriform *Garjainia* and the abundance of procolophonids, this would allow for the Vetluga (Zone V) and Baskunchak faunas to be directly correlated to subzone A of the *Cynognathus* Assemblage Zone. This correlation of the Vetluga and Baskunchak faunas with subzone A is further strengthened by the fact that these faunas are overlain by the Donguz Svite, shown below to contain genera common to subzones B and C. This being the case then the entire sequence known for the *Cynognathus* Assemblage Zone is also preserved in the Russian Triassic.

The *Eryosuchus* fauna of the Donguz Group broadly equates to the Heylerosaurid and Stenotosaurid fauna (Zone VI) of the Russian Platform and includes: the kannemeyriid genera *Rabidosaurus*, *Rhadiodromus* and *Uralokannemeyeria* (Danilov, 1971) the shansiodont *Rhinodicynodon* (Kalandadze, 1970); and an associated fauna including the bauriids *Notogomphodon* and *Dongosaurus*; the traversodontid cynodont *Scalenodon* (Parish, 1986); a possible traversodontid originally assigned to the genus *Antecosuchus*; and the captorhinid *Orenburgia*.

The direct correlation of this fauna is somewhat problematical as it contains *Uralokannemeyeria*, which, following (Cheng, 1980) I believe to be congeneric with *Shaanbeikannemeyeria*, and which in turn is closely related to *Kannemeyeria* from subzone B. The presence of the shansiodont *Rhinodicynodon* would however rather indicate a closer relationship with subzone C. Revision of the stratigraphic positions of these finds may prove that the equivalents of both subzones B and C are preserved in the rocks of the Donguz Group (*Eryosuchus* fauna).

4.2.2 Molteno Formation

4.2.2.1 Interbasinal correlation

Gondwanan correlatives for the Bamboesberg and Indwe Members are based on time equivalent floral, palynological and faunal evidence. The Molteno Formation is formally restricted to the main Karoo Basin and equivalents of the Bamboesberg Member are unknown outside of South Africa. Most previous work that has dealt with equivalents of the Molteno Formation outside of the main Karoo Basin has therefore been based on the Indwe Member, which has in the past been equated with deposits in Botswana, Namibia, Zimbabwe, and Kenya (Table.4.3).

In Botswana the equivalent of the Indwe Sandstone consists of a unit of coarse grained, pebbly, cross-stratified sandstones with intercalated purple or brown fines. This unit is lithologically very similar to the strata assigned to the Indwe in the Springbok Flats region. In Namibia, the Indwe is equated to a coarse grained,

trough cross-stratified pebbly sandstone approximately 80m thick (Visser, 1984).

The Escarpment Grit and the Ripple-marked Flags of Zimbabwe also consist of large sets of facies St₁ with pebbles of quartz, minor gneiss, schist and mudstone up to 3.5 cm in diameter. These are equated to the Indwe Sandstone Member by Visser (1984) and Cooper (1982), based on the occurrence of a *Dicroidium* flora in both units. In the Lebombo Region the Ntabene Formation has also been equated to the Indwe Sandstone Formation by Visser (1984). This unit consists ± 10 m of coarse grained, trough cross-stratified sandstone and subordinate shale, which rests unconformably on sedimentary strata equivalent to the Beaufort Group. In the Limpopo region, the Fripp Sandstone Formation and Bendozi Coal Series form a 240 m thick unit of medium to coarse grained sandstone overlain by thick mudstones and coal, which have been correlated with the Molteno based on their palaeontological signatures (Visser, 1984). Based on their lithological similarity, the Shimba grits of Kenya have also been equated to the Indwe Sandstone Member (Money & Prasad, 1977), and in Madagascar the Isalo Formation (Isalo II) has a similar lithological appearance and palaeontological content and is here proposed to be co-eval with the Indwe.

Although the author is of the opinion that all these sequences occur in different sub-basins, the widespread occurrence of coarse, cross-stratified pebbly sandstones in southern Africa raises the question as to the control on these sequences. If it is a tectonic control then the signature had to affect numerous different basinal settings, and be associated with some type of major tectonic re-organisation in southwestern Gondwana. The only other control on such widespread sequences is climate and sediment supply (Paola *pers. comm.*). Similar climates would probably have prevailed over much of the interior of southwestern Gondwana, and this probably accounts for the similar overprint to these sequences. Climate on its own does not seem to be a great enough influence to account for the nature of these sequences however, even if it was controlling sediment supply. It seems therefore that these widespread, coeval coarse grained sequences represent a major tectonic re-organisation of southwestern Gondwana prior to

fragmentation in the Late Triassic and Early Jurassic. This has been proposed for northern Gondwana and Pangea by Rockling & Beutler (1993) who believed that the Triassic was a period of tectonic instability.

Correlation of the Molteno Formation with other Gondwanan basins has previously been based mainly on the palaeontological aspects of the Formation. On palaeofloral evidence, Anderson & Anderson (1970, 1971) and Anderson (1974) considered the Bamboesberg and Indwe Members of the Molteno Formation to be the time equivalents of the Ipswich Coal Measures of Queensland, Australia, with several taxa common to both sequences. The palaeofauna (predominantly fish and insects) of the Bamboesberg and Indwe Sandstone Members also show a close relationship with the fauna of the Mount Crosby Formation of Australia (Rieck, 1974).

4.3

Age

Based on the above correlations the relative and absolute ages of the Burgersdorp (*Cynognathus* Assemblage Zone) and Molteno Formations may now be re-examined in light of known faunal contemporaneity. For establishing the age of the sequence, the timescales of Harland *et al.* (1989) and Ross *et al.* (*in prep.*) are used. Problems previously associated with the incompleteness of the succession and the relative dating of the two Formations is re-addressed.

4.3.1

Relative age

4.3.1.1

Burgersdorp Formation

Some dispute surrounds the ages of the *Lystrosaurus* and *Cynognathus* Assemblage Zones within the Burgersdorp Formation. This discussion deals primarily with the *Cynognathus* Assemblage Zone, but has important implications for the age of the *Lystrosaurus* Assemblage Zone as well as the development of the Karoo Basin during the Triassic. Arguments dealing with the age of the

Lystrosaurus Assemblage Zone are presented in Ochev & Shishkin (1989) and Shishkin *et al.* (1995), however this biozone is widely believed to be Lower Triassic (Induan) in age (Table 4.2).

i) ***Cynognathus* Assemblage Zone**

The *Cynognathus* Assemblage Zone has previously been assigned ages ranging from the Late Permian (Cooper, 1982) through to the Late Triassic (Broom, 1904). Most authors favour either an Early or Middle Triassic age, with some authors even suggesting both. This confusion is due to the fact that in terms of its bulk composition as listed by Kitching (1977), the biozone shares faunal similarities with both Lower and Middle Triassic faunal assemblages from other parts of Gondwana, China, Russia and Europe. The recognition of various subzones within the *Cynognathus* Assemblage Zone (Hancox *et al.*, 1995), and the correlations presented in the previous section, allows for some of these problems to be re-addressed.

a) **Subzone A**

The *Cynognathus* Assemblage Zone has long assumed to be the equivalent of the Early Triassic (TR1) upper Scythian (Spathian/Olenekian) of Europe due to the presence in the biozone of the European amphibian genus *Parotosuchus* (eg DuToit, 1954). Shishkin *et al.* (*in prep.*) show that the European genus *Parotosuchus* is not however congeneric with the South African *Parotosuchus africanus* and is in fact far more similar to, if not congeneric, with *Kestrosaurus*, the zone index fossil for subzone A (Table 4.2). *Trematosuchus* from Subzone A has also been shown to be closely related to *Trematosaurus* from the Early Triassic Middle Buntsandstein (Upper Olenekian) of Western Europe (Shishkin & Wolman, 1994).

The archiosauriform from subzone A is either congeneric with, or at the same evolutionary state as, the Russian genus *Garjainia* (Welmovskiy, 1969), known from the Vetluga Series (Zone V) of Russia. *Garjainia* is known from the

earliest Proterosuchia i.e. the Proterosuchidae (Cruickshank, 1978) which are presumed to be of lowermost Triassic in age (Zone IV). The age of the *Lystrosaurus* Assemblage Zone within the Scythian is presumed to be Lower Scythian (Griesbachian) at the latest (Ochev & Shishkin, 1989; Shishkin & Ochev, 1993). This would mean that the *Proterosuchus* bearing strata of the *Lystrosaurus* Assemblage Zone are Griesbachian (Induan) (Table 4.2) and that subzone A of the *Cynognathus* Assemblage Zone is Scythian (Spathian/Olenekian) in age.

The similarity of *Kestrosaurus* and the European *Parotosuchus*, and *Trematosuchus* with *Trematosaurus* (Shishkin & Welman, 1994) coupled to the primitive nature of the archosauriform tentatively assigned to *Garjainia* all point to subzone A being Lower Triassic (latest Scythian; Spathian/Olenekian) in age. Further evidence that it is an older fauna than that found in subzone B or C comes from the primitive nature of the *Trirachodon* species (Welman, *pers.comm.*), coupled to the total absence of kannemeyeriids, which Ochev & Shishkin (1989) believe first appear in the Mid-Triassic.

Although based on erroneous assumptions and generic misidentifications, previous authors were therefore correct in their assignation of an Early Triassic age, although not for the entire *Cynognathus* Assemblage Zone.

b) Subzone B

Terrestrial faunas spanning the Early-Middle Triassic contact are poorly known due to the incompleteness of the fossil record and the lack of precise dating for faunas presumed to be of these ages. Cruickshank (1963, 1967), Cox (1973) and King (1988) believed that typical Middle Triassic (Anisian) faunas include assemblages comprising kannemeyeriid and shansiodont dicynodonts, archosauriformes, bauriids, cynodonts and rhynchosaurs. Ochev & Shishkin (1989) deemed the contact between the Early and Middle Triassic as corresponding to the changeover from faunas dominated by proterosuchian archosauriformes to one dominated by kannemeyeriid dicynodonts.

Euparkeria and more particularly *Erythrosuchus* show advances over the state found in *Garjainia* and the grade of evolution of the *Euparkeria* ankle is almost identical to that of *Chanaresuchus* (Romer, 1972) from the Ischichuca Formation (= Chanares Formation) which is of late Middle Triassic age. This fact was used by Cruickshank (1986) to assign the *Cynognathus* Assemblage Zone an early Middle Triassic age. The presence of Euparkerrids (archosauriformes) in the *Cynognathus* Assemblage Zone has also in the past been used as evidence for a Middle Triassic age for the biozone, based upon a literal reading of their fossil record (Dilkes, 1994). The fish fossils collected from Bekker's Kraal (Broom, 1909b) are similar to forms from the Brookvale Formation of New South Wales (Hutchinson, 1972) that is also Mid-Triassic in age.

The faunal association of subzone B, dominated in numbers by the kannemeyeriid dicynodont *Kannemeyeria*, the cynodonts *Cynognathus*, *Diademodon* and *Trirachodon* and by the archosauriformes *Erythrosuchus* and *Euparkeria* therefore shares a number of close ties with faunas proposed to be of Mid-Triassic age. One problem with the assignment of subzone B entirely to the Mid-Triassic lies the close relationship between *Kannemeyeria* and *Shaanbeikannemeyeria* from the Heshangou Formation of China, which is supposedly of Early Triassic age based on the ostracod and pollen signatures (Pang, 1993). This being the case then: either *Kannemeyeria* postdates *Shaanbeikannemeyeria*; subzone B is partly Early Triassic in age and the FAD of kannemeyeriid dicynodonts does not tie in with the base of the Mid-Triassic; or that the stratigraphic assignment of *Shaanbeikannemeyeria*, or the dating of the Heshangou Formation is incorrect.

c) Subzone C

Based on the presence of the dicynodont *Angonisaurus* and *Shansiodon*, subzone C may be directly correlated with the Manda Formation of Zambia and the Upper Ermaying Formation of China. The age of the Manda Formation is at present debatable, however it is generally regarded as being broadly of Mid-Triassic age (Parrington, 1946; Crompton, 1955; Watson, 1958). Anderson & Cruickshank

(1978) place it in the Middle Anisian, whilst Cox (1973) and Cox & Li (1983) suggest that it is approximately uppermost Anisian, based on the similarity between *Mandasuchus* and *Ticinosuchus* from Monte San Giorgio, Switzerland. Cruickshank (1986) suggests that the Manda might represent a stage later in time than the middle Anisian normally assumed for it, a view shared by Shishkin (*pers.comm.*) who feels it may even be Ladinian in age. The Upper Ermaying Formation is also considered to be Middle Triassic (Anisian) by most authors (Cheng, 1980; Shishkin *pers.comm.*). Based on the ages assigned to the Manda and Upper Ermaying Formations, subzone C may be viewed as being Mid-Triassic, probably middle to late Anisian.

ii) Conclusion

Subzone A contains a fauna that may be assigned an Early Triassic (Scythian) age with some confidence. Both subzone B and C are placed within the Middle Triassic (Table 4.2), with subzone B thought to be earliest Anisian, and subzone C slightly younger, possibly even late Anisian or earliest Ladinian. The boundary between the Early and Mid-Triassic in South Africa may therefore be seen to correspond to the boundary between subzones A and B of the *Cynognathus* Assemblage Zone. This fact accounts for the previous assignments of both an Early and Middle Triassic age to the biozone.

4.3.1.2 Molteno Formation

Based on the presence of the *Dicroidium* flora, the palynological signature, and to a lesser degree on the fish and insect palaeofaunas, the age of the Molteno Formation has been proposed to be anywhere between uppermost Lower Triassic (Scythian) (Stapleton, 1978; *in* Cooper, 1982), Middle Triassic (Keyser, 1973) and Upper Triassic (Carnian-Norian) (Dunn, 1878; Anderson & Anderson 1970; Anderson, 1974).

Stapleton's (1978) assignation of a Lower Triassic age may be disproved by the foregoing dating of the *Cynognathus* Assemblage Zone. Keyser's (1973) Middle Triassic age was based on the belief that the fauna from the Omingonde Formation was equivalent in age to the Molteno Formation. The fauna of the Omingonde Formation may however be correlated to subzones B and C of the *Cynognathus* Assemblage Zone and is therefore not the temporal equivalent of the Molteno Formation.

The best way to bracket the age of the entire Molteno Formation therefore seems to be by applying the ages determined for the uppermost *Cynognathus* Assemblage Zone, and those known for the overlying Elliot Formation. By comparison of the footprint assemblages with those of western Europe, Ellenberger *et al.* (1967) proposed an Upper Triassic age for the lower Elliot Formation. Olsen & Galton (1984) also proposed an Upper Triassic (Carnian-Norian) age for the *Euskelosaurus* Range Zone, based on correlation with North American sequences. Comparison of the fauna of the lower Elliot with that of Europe and North America suggests a very close affinity, especially between the prosauropod genera *Euskelosaurus* and *Plateosaurus*. *Plateosaurus* occurs in well dated Norian (Late Triassic) deposits (Sander, 1992) and although *Euskelosaurus* may be slightly more primitive than *Plateosaurus* (Juul, *pers.comm.*), their similar evolutionary state allows for a Norian age to be applied to the lowermost Elliot Formation.

The age of the Molteno Formation may therefore be seen to be constrained by the Middle Triassic (Anisian) age of subzone C of the *Cynognathus* Assemblage Zone, and the Norian age of the lowermost Elliot Formation (*Euskelosaurus* Assemblage Zone). These ages therefore bracket the Molteno Formation between the Ladinian and the Carnian. The age of the Bamboesberg and Indwe Members must fall within this broad band, yet little evidence is available to further refine the ages of these two Members. Anderson & Anderson (*in prep.*) believe *Dicroidium zuberi* to be restricted in its range to the Indwe Sandstone Member, and in South America this species occurs stratigraphically within strata reliably dated as Carnian (Webb, 1981). Anderson (*pers.comm.*) believes the Bamboesberg Member (unit 1) to be

barren of pollen, but assigns the Indwe Sandstone Member (unit 2, a Carnian age based on its palynological signature. The typical *Allisporites/Falcisporites* assemblage is however only broadly indicative of a Triassic age, as *Allisporites* is known from the Fremouw and Falla Formations of Antarctica, which are assigned ages ranging from the Anisian through to the Carnian (Farabee *et al.*, 1990).

The Bamboesberg and Indwe Sandstone Members also share several taxa of palaeoflora with the Ipswich Coal Measures of Australia (Anderson & Anderson, 1983), which Flint & Gould (1975) assign a Carnian age. The fish and insect palaeofauna of the Bamboesberg and Indwe Sandstone Members also show a close relationship to the fauna of the Mount Crosby Formation of Australia, which Banks (1978) dates as Carnian.

From the above it is evident that the Indwe Sandstone Member may be reliably dated as Carnian on palaeofloral, palynological and palaeofaunal grounds, and that the Bamboesberg Member must be pre-Indwe in age. The palaeontological and lithological ties between these two Members however suggests that their ages are fairly similar, and the Bamboesberg Member is provisionally placed within the Carnian.

4.3.2 Definition of a global biochron for the Middle Triassic

The standard Triassic timescale has long been based on marine biochronology (Tozer, 1967; 1984), being developed on fossil data, especially ammonoids and conodonts. Indecision still exists however regarding the adoption of an international standard for the Triassic ammonite zones (Tozer, 1984, 1993), as does the ages applied to the various stages (Lucas, 1990). Little attention has in the past been devoted to nonmarine chronology in the development of the geological timescale, mainly to a bias against the use of terrestrial fossils for long range correlation (Lucas, 1992). This bias is rooted in a number of perceptions including: that terrestrial fossils are facies specific and are not widespread; that terrestrial evolution did not proceed at the same pace as change in the marine environment,

and therefore terrestrial taxa have stratigraphic ranges that are too great to allow them to be truly useful index taxa; and that terrestrial fossils, especially tetrapods, are not sufficiently abundant to be useful for developing detailed biostratigraphies.

That the above is not the case for the Beaufort Group is very evident by the use of tetrapod fossils for the detailed biostratigraphy of the Group, and as has been shown in the section on correlation, both amphibians and dicynodonts have a global distribution. In the Burgersdorp Formation (*Cynognathus* Assemblage Zone) amphibian forms evolved as quickly as their marine counterparts, and although fragmentary they are very abundant, especially in subzone A. These biases may therefore be removed through detailed collecting and taxonomic assessment of fossiliferous non-marine sequences.

A renewed diversification of dicynodont forms occurs during the Middle Triassic, with most non-marine faunas dominated in numbers by dicynodonts, cynodonts, rhynchosaurs and advanced capitosaurid amphibians. For this reason, dicynodonts have played an important role in defining the faunal epochs, chronofaunas and biochrons proposed for the Triassic by the likes of Romer (1970), Anderson & Cruickshank (1978), Cooper (1982), Cruickshank (1985), Ochev & Shishkin (1989), DeFauw (1993), Lucas (1993a) and Shishkin & Ochev (1993).

Amphibians also form a vital correlative link between the marine and nonmarine Triassic. Prior to the works of Shishkin & Welman (1994) and Shishkin *et al* (*in prep.*) however, direct comparison of amphibian forms from Gondwana, Russia, China and Europe had not been undertaken, and their usefulness for global biochronology was not recognised. These studies have however shown that forms such as *Trematosuchus* and *Parotosuchus* have a global distribution in the Triassic, and may in future be used to define global biochrons. *Parotosuchus* in particular provides a useful index for the Early Triassic.

Although both dicynodonts and amphibians have therefore been shown to be globally distributed, certain authors still favour provinciality during the Triassic

(Battail, 1980; Thulborn, 1984). This may in part be due to what Cooper (1982) refers to as provincial taxonomy. A number of other palaeobiological studies of the late Palaeozoic to early Mesozoic tetrapod faunas (Anderson & Cruickshank, 1978; Cooper, 1982) however point to a unity of Gondwana, China, Russia and Europe during this time.

Cooper (1982) identified six intervals for Triassic time, with three of these confined to the Mid-Triassic. His *Tetragonias* zone encompasses the entire Anisian and the *Dinodontosaurus* and basal part of the *Stahleckeria* Zones covers the Ladinian. The *Tetragonias* zone is based on the fauna of the Manda Formation and is characterised by the presence of shansiodonts, advanced kannemeyeriids, chiniquodontid and traversodontid cynodonts.

Ochev & Shishkin (1989) divided the Triassic into three successive global faunal epochs (faunachrons), based on archosauriformes, amphibians and dicynodonts. Their basal Proterosuchian epoch is equated to the entire Lower Triassic and is divided into a lower *Benthosuchus-Wetlugasaurus* (neorachitome fauna) and an upper *Parotosuchus* fauna, and equated to the Lower Triassic. Their Kannemeyeroïd epoch includes the first appearance of both kannemeyeriid dicynodonts and psuedosuchian archosauriformes and equates to the Middle Triassic, and their Dinosaur epoch was equated to the whole of the upper Triassic.

Both DeFauw (1993) and Lucas (1993a) proposed global biochrons for the Middle Triassic based on dicynodont genera. The first of these is based on the kannemeyeriid *Rechnisaurus* (DeFauw, 1993) and the second on the Shansiodontid *Shansiodon* (Lucas, 1993a). Both of these proposals however have problems associated with the taxonomy of the genera in use. The discovery of a new fauna from South Africa and a revised taxonomy of the Middle Triassic dicynodonts (Hancox *in prep.*) allows for these proposals to be refined. *Rechnisaurus* is too poorly known at present to be of use in defining a global distribution and, although Lucas (1993a) synonymised a number of genera with *Shansiodon*, thereby making its distribution global, at present this genus is actually known only from China.

Based on the First Appearance Datum (FAD) of the Family Shansiodontidae, Hancox (1996) informally proposed a shansiodontid biochron for defining the Mid-Triassic. This biochron was similar to the *Shansiodon* biochron of Lucas (1993a), but reflected the advances made in the taxonomy of the Middle Triassic dicynodonts. The designation of a Middle Triassic age to subzone B of the *Cynognathus* Assemblage Zone however means that the FAD of Shansiodontids in the main Karoo Basin does not tie in with the base of the Mid-Triassic, which is rather represented by the FAD of kannemeyeriid dicynodonts as proposed by Ochev & Shishkin (1989).

To date therefore no global biochron for the Mid-Triassic has been accepted. The striking similarities between the Middle Triassic tetrapod faunas of southwestern Gondwana, China, Russia and Europe however provides evidence for unhindered migration and gene flow during this time, and holds promise that such an undertaking may be possible in the near future.

4.3.3

Absolute age

"Those ages, like those of railway timetables, are subject to change without notice"

Mason (1966)

Numerous geological timescales have been erected in the past and refinements are continually occurring. Applying absolute ages from such timescales is therefore of limited practical value, other than allowing for quantification of the duration of deposition of the two Formations, as well as the length of time represented by the contact that separates them. The following absolute ages applied to the Burgersdorp and Molteno Formations are based on the geological timescales of Harland *et al.* (1989) and Ross *et al.* (*in prep.*) (Table 4.4).

Taking the *Cynognathus* Assemblage Zone to represent the latest Early Triassic (Scythian; Spathian; 241.9-241.1Ma) to the Middle Triassic (Anisian; 241.1-239.5Ma), and applying the absolute ages of Harland *et al.* (1989), the maximum timespan represented by the entire *Cynognathus* Assemblage Zone would be 2.4Ma, with subzone A ≤ 0.8 Ma and B&C combined ≤ 1.6 Ma. These figures are in keeping with the length of time proposed for Beaufort Group biozones by Cruickshank (1986) and King (1993). Taking the Molteno Formation to represent the entire Carnian (235-223.4 Ma), gives a timespan for deposition of the entire Formation of 11.6Ma.

Following the timescale of Ross *et al.* (*in prep.*) the Anisian may be constrained between 241-234Ma and the Carnian between 220-228 Ma. Taking these figures would increase the timespan of subzones B and C of the *Cynognathus* Assemblage Zone to ± 7 Ma and reduce the timespan represented by the Molteno Formation (Bamboesberg and Indwe Members) to a maximum of ± 8 Ma.

4.4

The Nature of the Contact between the Burgersdorp and Molteno Formations

As mentioned in Chapter One much debate has centred around the criteria for the definition of the base of the Molteno Formation, and there is some confusion in the literature regarding its recognition, placement and nature. Prior to the work of Turner (1975a) the Bamboesberg Member, and hence the true lower boundary of the Molteno Formation was not defined, and the lower contact was presumed to be the same throughout the basin.

Previous authors have used a number of criteria for defining the basal contact and these are first reviewed in terms of the findings of this thesis. The contact is then defined for the south of the basin where the sequence is most temporally complete, and the spatial and temporal variation of the contact is discussed.

4.4.1 Review of previous criteria used for the placement of the contact between the Burgersdorp and Molteno Formations

i) Colour

Dunn (1873) was the first to document that the colour of the Molteno Formation contrasted strongly with that of the Burgersdorp Formation. DuToit (1954), Rust (1959, 1962), Haughton (1969) and Kitching (1977) all state that the Molteno is readily distinguishable from the Beaufort Group by the absence of maroon coloured strata. Turner (1975a) felt that the sandstones of the Bamboesberg Member could be distinguished from those of the Burgersdorp Formation by their predominantly paler colour, but later noted that colour change was of little value in defining stratigraphic boundaries on either a local or regional scale (Turner, 1981).

Two main factors mitigate against colour as a reasonable criterion for the placement of the lower boundary in the south of the basin. The first of these is that the greyish red (5R 4/2) coloured fines of the Burgersdorp Formation are replaced by light brown (5YR 6/4) and greyish orange (10YR 7/4) coloured rocks between 15 and 20m below the base of the Bamboesberg Member (Fig.2.83). The second is that dusky red (5R 3/4) coloured sediments, although rare, also occur in the lowermost levels of the Bamboesberg Member.

ii) Glittering nature

The base of the "Stormberg Group" was defined by DuToit (1954) as the occurrence of the first glittering sandstone of the Molteno Formation, and a number of authors still favour this criterion. It should be noted that at this time the Bamboesberg Member was unknown and DuToit's (1954) definition therefore actually refers to the base of the Indwe Sandstone Member. This criterion does not hold in the south of the basin as most of the sandstones in the lower Bamboesberg Member do not have significant quartz overgrowths and therefore do not glitter.

iii) Grainsize

DuToit (1954), Turner (1975a) and Kitching (1977) all state that the Molteno Formation may be separated from the Burgersdorp Formation on the basis of its coarser grainsize. As noted by Christie (1981), the Bamboesberg sandstones are consistently more medium grained than those of the Burgersdorp Formation, however the two are commonly difficult to differentiate texturally in the field. Because of this problem Christie (1981) considered that it could be difficult to determine the boundary if the contact occurred within an association of sandstones. This is seen to be the case in the south of the basin at Bushmanshoek Pass, as the lowermost sandstone of the Bamboesberg Member is no coarser than the uppermost sandstone of the Burgersdorp Formation and the contact cannot be placed on grainsize alone.

iv) Sandstone/Fines ratio

Christie (1981) stated that the Bamboesberg Member was readily distinguishable from the underlying Burgersdorp Formation by its dominantly arenaceous nature, containing as little as one percent argillaceous material. He therefore placed the base of the Molteno Formation at the horizon at which sandstone succeeded siltstone as the dominant lithology. MacDonald (1993) followed Christie in using the ratio of sandstones/fines as a criteria for the placement of the contact, stating that it may be placed where this ratio changes from less than 0.25 to more than 0.50. Christie and MacDonald (*in prep.*) also defined the boundary as the horizon above which sandstone becomes the dominant lithology, but state that the lower contact is gradational and conformable where the Bamboesberg Member is developed.

The fact that the Bamboesberg Member in general contains more arenaceous material is not in question, however, in the south of the basin the first two fining upward cycles of the Bamboesberg Member may have a considerable fines content, with a sandstone/fines ratio of only 0.4. The change in the sandstone/fines ratio

as used by Christie (1981), MacDonald (1993) and Christie & MacDonald (*in prep.*) is therefore inadequate when dealing with the contact in the south of the basin.

Too often in the past, only a single, non-tenable criterion has therefore been used to define the Beaufort-Molteno contact, and at times these do not refer to the true chronostratigraphic contact in the south of the basin.

4.4.2 Definition of the contact in the south of the basin

As has been discussed above, the colour, glittering nature, grainsize and sandstone/fines ratio are not significantly diagnostic to allow for the definition of the contact between the upper Burgersdorp and the lower Bamboesberg Member in the south of the basin. Although the channel sandstone complexes of the upper Burgersdorp Formation and the lowermost part of the Bamboesberg Member share a number of gross similarities, including textural characters and the abundance of facies Sh, they differ significantly in terms of: their gross geometries; internal architectural elements; facies types and relationships; and palaeontological signatures.

A more precise and reliable definition of the contact in the south of the basin is therefore the change in channel sandstone geometry from that of a high sinuosity meandering system (for the Burgersdorp Formation) to a low sinuosity shallow braided system (for the lowermost Bamboesberg Member). This is indicated by a change in geometric form of the sandstones from laterally inextensive lenticular bodies, with deeply erosive bases, and channel width/depth ratios of between 7-20, to laterally extensive (kilometres) tabular, sheetlike bodies with horizontal bases, and width to depth ratios between 30-70.

The architectural elements of the channel fills of the upper Burgersdorp and lowermost Bamboesberg also differ. Point bars and large cut bank clasts which occur in the upper Burgersdorp Formation are not found in the Bamboesberg Member and the channel complexes of the Bamboesberg Member are furthermore

dominated by thin, minor channel fills, longitudinal and lesser transverse bars, and large sand flat macroforms, which are unknown in the upper Burgersdorp Formation.

Extraformational clasts do not occur in the Burgersdorp Formation and although rare in the lower Bamboesberg Member, are present on the basal contact surface. This criterion is probably the single most definitive indication of the placement of the lower contact. The sandstones of the upper Burgersdorp Formation and lowermost Bamboesberg Member may also be distinguished on their petrographic signatures, as the sandstones of the upper Burgersdorp Formation are predominantly lithic arenites, whereas in the lowermost Bamboesberg Member feldspathic and quartz wackes and arkosic arenites dominate.

The internal sedimentary structure of the fines may also be used to distinguish between the two Formations. The fines in the uppermost Burgersdorp Formation are predominantly massive (Fm) and bioturbated, whereas those from the Bamboesberg Member are finely laminated (Fl) and non-disturbed. The uppermost part of the Burgersdorp Formation in the south of the basin (*Cynognathus* Assemblage Zone, subzone C) also preserves an iron (high Fe_2O_3) rich clay horizon, some 2-5m below the basal Bamboesberg sandstone. This horizon is preserved throughout the aerial extent of the upper Burgersdorp Formation, and may be used as a marker horizon, when placing the contact.

The presence of coalified stringers and logs (Fig.3.70) within the contact sandstone of the Bamboesberg Member is also a useful field criterion, as coal is unknown from the Burgersdorp Formation. Care should however be taken in applying this criteria too liberally, as charcoal is preserved in small amounts in the upper reaches of the Burgersdorp Formation. The contact sandstone of the Bamboesberg Member in the south of the basin is further characterised by an abundance of plant remains (Figs.3.69 & 3.71). Here too care must be exercised in applying this criteria, as fossil plant material also becomes more prevalent in the sandstones of the upper Burgersdorp Formation. Plant fossils in the upper Burgersdorp Formation are

however predominantly poorly preserved imprints of pteridosperm axes (Fig.2.75a,b), whereas those of the lower part of the Bamboesberg Member show far better preservation. The exact taxonomic changes which occur are still under investigation.

The lithological changes noted across the boundary also correlate to a palaeontological change from a sequence dominated by a rich and diverse reptilian fauna (upper Burgersdorp; subzone C), to a well preserved, medium diversity palaeoflora for the Bamboesberg Member (Anderson & Anderson, 1989), within which tetrapod fossils are unknown.

To conclude, the contact between the upper Burgersdorp Formation and Bamboesberg Member in the south of the basin may be defined on: the geometries of the channel fill complexes; their width/depth ratios; preserved architectural element and facies associations; the presence/absence of extraformational clasts; the presence/absence of coalified logs; the nature of the fines component; and on their petrographic signatures. Supplementary palaeontological criteria include the abundance and nature of preserved fossil plant material, and the presence/absence of tetrapod fossils. These differences are summarised in Table 4.5.

4.4.3 South-North variation in the nature of the contact

The regional study of Turner (1975a) showed that the Bamboesberg Member wedged out from the south to the north of the basin (Fig.1.1). The nature of this change has been documented during this study, and the contact may be shown to be highly diachronous with rocks containing a subzone C fauna overlain by the Bamboesberg Member in the south of the basin, and subzone A overlain by the Indwe Sandstone Member in the north (Fig.4.5). Various of the criteria that are not tenable in the south of the basin, may however be of use in defining the contact in the middle and north of the basin.

At the most westerly outcrop of the Bamboesberg Member, the lower two sandstone channel complexes (BB1 & BB2) are no longer present and the contact occurs between the third channel complex (BB3) and rocks of the Burgersdorp Formation which contain a subzone B fauna. This has the affect that the characteristic lag found at the base of BB1 and BB2 in the south of the basin is not present. The contact may however be defined on all the other characters listed for the south apart from this one.

In the vicinity of Aliwal North only the top two units (BB4 & BB5) of the Bamboesberg Member occur, overlying rocks of the Burgersdorp Formation containing a subzone B fauna. Here the base of the Molteno Formation may be defined by a change in the: gross sandstone morphology; internal architecture; dominant facies type; colour; grainsize; and petrography; as well as on the abrupt palaeontological break. The sandstones of the Bamboesberg Member in this area have a higher percentage of facies Sp than in the south of the basin, but differ from the Burgersdorp Formation sandstones, which are predominantly structured by facies Sh.

Between the towns of Rouxville and Zastron, the basal four channel complex sandstones of the Bamboesberg Member do not form part of the stratigraphic succession, and the colour of the fines between the two formations may be used as an additional defining criteria when placing the contact. The fines of the Burgersdorp Formation in this area are predominantly greyish red (5R 4/2 to 10R 4/2), whilst those of the overlying Bamboesberg Member are light olive grey (5Y 6/1) and dark greenish grey (5GY 4/1). Grainsize also becomes a useful field character for defining the contact, as the sandstones of the Burgersdorp Formation are consistently fine grained, whereas those of the Bamboesberg Member are medium-coarse grained. In this area, the nature of the fining upward sequences in the Burgersdorp Formation and Bamboesberg Member may also be used to distinguish between the two sequences. The fining upward sequences in the Burgersdorp Formation consist of a basal channel-fill sandstone overlain by thick sequences of siltstone and mudstone, whereas in the Bamboesberg Member the

basal channel-fill sandstone comprises most of the sequence, being capped by only a thin sequence of siltstone, carbonaceous mudstone or coal.

To the north of Vanstadensrus, where the Bamboesberg Member is not present, the Indwe Sandstone Member unconformably overlies Burgersdorp Formation strata which contain a subzone B fauna. The contact is erosional and usually occurs between greyish red (10R 4/2) fines of the Burgersdorp Formation and moderate yellowish brown (10YR 5/4) coarse, glittery, tabular channel complex sandstones of the overlying Indwe Sandstone Member. The base of the Indwe Member in this area often co-incides with a concentration of small extraformational clasts of vein quartz and jasper.

In the northernmost extent of the basin in the vicinity of the town of Senekal, greyish red (10R 4/2) fines of the Burgersdorp Formation contain a subzone A fauna, and are overlain by coarse grained, moderate brown yellow (10YR 5/4) sandstones of the Indwe Sandstone Member. The basal part of the Indwe Sandstone Member consists of facies G₀, with grains of vein quartz and jasper up to 8mm in diameter. The overlying channel sandstone complex is dominated by large and small scale sets of facies G_p and S_p (Fig.3.62). This is in contrast to the underlying Burgersdorp Formation sandstones which are internally structured by fine grained sandstones of facies S_h and S_t.

iii) Discussion

Due to the change in the nature of the lower contact, and the similarities between the upper Burgersdorp and lowermost Bamboesberg in the south of the basin, many authors consider it is more convenient and logical to place the boundary between the "Stormberg" and Beaufort Groups at the base of the Indwe Member (Stavrakis, 1980; Cooper, 1982; Greenowald *pers.comm.*). Cooper (1982) actually includes the Bamboesberg Member at the top of the Beaufort Group, and places the major hiatus surface in the basin between the Bamboesberg and Indwe Sandstone Members. The Bamboesberg Member however displays sufficient differences from

the underlying Burgersdorp Formation to warrant separation from it, and significant similarities to the Indwe Sandstone Member to warrant inclusion within the Molteno Formation. The most striking similarities with the Indwe Sandstone Member being: the nature of the sandbody geometries; the presence of extraformational clasts of CFB origin; the presence of coal; and the palaeontological signature.

As presently defined by Johnson & Hiller (1990) and MacDonald (1993), the contact is believed to be a transitional, conformable zone, some 100m thick. Although the upper Burgersdorp Formation has certain similarities to the lower Bamboesberg Member, at no locality in the basin was the contact transitional. Although there is some sedimentological evidence for an unconformity separating the Bamboesberg Member and Burgersdorp Formation in the south of the basin, no diagnostic evidence, such as an angular discordance or the development of thick palaeosol horizons (Bown & Kraus, 1987) exists. The palaeontological signatures of the two Formations on the other hand differ significantly and such a difference may be used as evidence of an erosional unconformity when there is considerable temporal difference in the assemblages (Crompton & Allen, 1995). The recognition of a regional subaerial unconformity (SU) at the base of the Molteno Formation, and the abrupt palaeontological break therefore provide firm evidence against a conformable nature for the contact sequence.

4.4.1 Time represented by the contact

4.4.4.1 Relative time represented by the contact

DuToit (1954) held the view that the contact between the Beaufort and "Stormberg" Groups was one of the two major temporal breaks in the main Karoo Basin. Estimates of the length of time represented by this contact however range from conformable (Christie, 1981), to tens of millions of years (Visser, 1991). Bearing in mind the different datings of the proposed subzones of the *Cynognathus* Assemblage Zone, and the fact that the nature of the contact changes from south-north through the basin, the relative amounts of time actually represented by the

contact in the different parts of the basin may now be discussed.

The placement of the entire *Cynognathus* Assemblage Zone in the Lower Triassic (Spathian) and the Molteno within the Upper Triassic (Carnian) (Anderson & Anderson, 1993a), implied a considerable hiatus between the two, incorporating all of the Middle Triassic (Anisian and Ladinian). This would almost account for the proposed 50Ma hiatus between the Molteno and Burgersdorp Formation as proposed by Visser (1991), as well as the Mid-Triassic lacuna of Cole (1992). On the other hand, applying an Anisian age to the upper Burgersdorp (Kitching, 1995), and a Ladinian age to the Bamboesberg Member (Keyser, 1973; Smith *et al.*, 1993)), significantly reduces the hiatal surface in the south, such that the sequence would be almost conformable, as proposed by Christie (1981).

Taking the lower Bamboesberg as Carnian as proposed in this thesis, the hiatal surface in the south of the basin, where EB1 disconformably overlies the uppermost Burgersdorp Formation (*Cynognathus* Subzone C; late Anisian), represents the entire Ladinian. In the middle of the basin where the uppermost Bamboesberg (BB4 or BB5) overlies middle Burgersdorp strata (*Cynognathus* Subzone B; early Anisian), the hiatus probably represents only the late Anisian and the Ladinian. In the north of the preserved limit of the basin, where the Indwe Sandstone Member disconformably overlies the lowermost Burgersdorp (*Cynognathus* Subzone A) the unconformity represents both subzones B & C. This period of time spans the entire Middle Triassic (Anisian and Ladinian) and possibly even the early Carnian. The relative ages applied to the Burgersdorp Formation (*Cynognathus* Assemblage Zone) and Bamboesberg and Indwe Members of the Molteno Formation in this thesis therefore suggest a strongly diachronous nature to the basal hiatal surface of the Molteno Formation.

It is interesting to note that all over Gondwana a Ladinian tetrapod palaeontological "gap" is known to occur (Anderson & Anderson, 1993a). This "gap" has in the past been related to a period of eustatic sea-level fall and global marine lowstand, with erosion in the nonmarine continental interiors (Anderson & Anderson,

1993a,b). Such an inter-regional unconformity may however also be produced by major orogeny, especially when coupled to periods of changes in sea-floor spreading rates (Crampton & Allen, 1995). Considering the world-wide lack of Ladinian age faunas and deposits, and the ages applied to the erosional unconformity at the base of the Molteno Formation, it is possible that the hiatal surface between the Burgersdorp and Molteno Formations may be tied to a Gondwana-wide event. Such an phenomenon is further evidenced by the widespread occurrence of "Indwe" style sedimentation in all the sub-basins of south-western Gondwana following the Ladinian event.

4.4.1.2 Absolute time represented by the contact

The absolute ages of the two formations are discussed purely to try and quantify the estimates of the time represented by the contact between the Burgersdorp and Molteno Formations. The timescales of both Harland *et al.* (1989) and Ross *et al.* (*in prep.*) (Table 4.4) are used to show how different absolute dates would affect the magnitude of time represented by the contact. It should be noted here that as the geological timescale becomes more refined, so too will the estimates of the absolute time represented by the contact.

Previous estimates of the absolute time represented by the contact are limited to the works of Visser (1984, 1991) who provided an age estimate of between 10-50 Ma for the basal hiatus. Looock (*pers.com.*) however believes that the hiatus at the base of the Molteno Formation represents a period of only 1-2 Ma. The timespan represented by the basal hiatal surface of the Molteno Formation therefore ranges from between near conformable (1 Ma) to some 10 Ma for the south and 50 Ma or more for the north of the basin (Visser, 1984).

Based on the ages of Harland *et al.* (1989), it may be shown that in the south of the basin the upper Burgersdorp Formation (subzone C: Anisian; 241.1-239.5Ma) is overlain by the Bamboesberg Member (Carnian: 235-223.4Ma), the disconformity between the two Formations represents at least 4.5Ma and at most 6.1Ma.

Applying the same ages to the north of the basin where lower Burgersdorp Formation (Subzone A: Spathian; 241.9-241.1Ma) are overlain by the Indwe Sandstone Member (Carnian: 235-223.4Ma), the magnitude of the hiatal surface ranges from a minimum of 6.1Ma (Anisian & Ladinian) to a maximum of 18.5Ma (beginning of the Scythian to the end of the Carnian). Extending the age of the Indwe Sandstone Member into the late Norian (223.4 - 209.5Ma), as suggested by Anderson & Anderson (1993a) extends the maximum to 32.4Ma. Even the most extended range therefore falls far short of Visser's (1991) 50Ma hiatal surface.

Applying the timescale of Ross *et al.* (*in prep.*) the hiatus in the south of the basin (Ladinian: 234-228) is of the order of 6 Ma, whereas in the north of the basin (Anisian-Ladinian: 241-228Ma) it is at least 13Ma. Applying this timescale to a Norian age (220-212Ma) for the Indwe Sandstone Member increases this maximum figure to 29Ma.

Based on a best age estimate of late Anisian for subzone C and early Carnian for the Bamboesberg Member, the timespan represented by the hiatus in the south of the basin therefore ranges from 4.5-6Ma. This timespan differs significantly to the north of the basin where Spathian aged rocks of the Burgersdorp Formation are overlain by the Carnian aged Indwe Sandstone Member. Here the hiatal surface is of greater magnitude, ranging from between 6.9-13Ma to a maximum of 29-32Ma dependant on the timescale followed. The close of Beaufort Group sedimentation was therefore followed by a long period during which non-deposition or erosion occurred. The unconformity between the Burgersdorp Formation and Indwe Sandstone Member in the north of the basin represents the greatest magnitude unconformity surface developed in the main Karoo Basin during the Triassic, and may possibly be linked to a major period of change in tectonic style in southwestern Gondwana.

Such inter-regional erosional or non-depositional subaerial unconformities (SU) are believed to be controlled either by tectonism or eustacy (Posamentier & Vail, 1988; Posamentier *et al.*, 1988; Vaill *et al.*, 1991; Ettensohn, 1994; Crampton & Allen,

1995). The recognition of such a widespread diachronous unconformity surface in the upper Karoo has important implications for the development of the basin during the Triassic.

4.5 Basinal development of the upper Karoo during the Triassic

The data presented in Chapters Two and Three, along with the proposed palaeoenvironments, intrabasinal correlations, and ages of the Burgersdorp and Molteno Formations, are here combined with the current literature to present a model for the development of the main Karoo basin during the Triassic.

Many of the palaeogeographic factors that control the evolution of sedimentary basins, including the: basins shape; size; subsidence behaviour; water depth; and the location and elevation of sediment sources, are controlled by the plate tectonic setting (Bally & Snelson, 1980; Allen *et al.*, 1986; Condio, 1989). The tectonic framework of any particular depositional basin and its source area also has a direct and major bearing on the quantity and type of sedimentary fill (Krumbein & Sloss, 1963). To understand the fill of a basin, or part thereof, it is therefore first necessary to have an understanding of the tectonic framework within which the basin occurs.

4.5.1 Tectonic setting of the Karoo Basin

The main Karoo Basin was originally the largest and deepest of several separate, or partly connected depositories which existed on a large intracratonic plate that formed southwestern Gondwana (DeWit, 1977; Johnson, 1991; Smith *et al.*, 1993; Veevers *et al.*, 1994). As evidenced from the foregoing correlations, the stratigraphic fills of these basins are now found in the southern half of Africa, South America, Antarctica, India and Australia.

The Karoo Basin is a retro-arc foreland basin (Johnson, 1991) bordered by the Cape Fold Belt (CFB) in the south and the Kaapvaal Craton (KC) in the north. Johnson & Beaumont (1995) prefer the usage of the term retro-foreland to distinguish basins formed on the stable cratonward lithosphere, and which preserves no evidence of a distinct magmatic arc. Because no direct evidence occurs for a magmatic arc source during the Triassic, this usage is followed here. A distinction must also be drawn between the thrust belt and adjacent foreland basin, yet there is significant control imparted by one on the other, and both the formation and fill of the Karoo Basin is inherently linked to the orogenesis of the CFB (Hälbich, 1977, 1983a-d, 1992; Cole, 1992). The orogenesis of the CFB was controlled by the collision of the palaeo-Pacific and Gondwanan plates, and originally produced an extensive retro-foreland basin (RFB) with its depocentre situated in the southwest, however during the Permian-Triassic the depocentre migrated northeast (Thomas *et al.*, 1993).

The majority of RFB development is furthermore related to thin-skinned thrust belts (Stockmal & Boutilier, 1986; Jordan, 1995), where pre-existing sedimentary cover rocks are shortened by folding and thrusting above undeformed basement (Allmendinger *et al.*, 1983). This is also the case for the Karoo Supergroup where the Cape Supergroup rocks formed a pre-existing, thick pile of layered sedimentary strata.

a) Mechanism of Cape Fold Belt formation

The Cape Fold Belt (CFB) is part of the Pan Gondwanian mobile thrust belt which extended from South America through to Australia. Preserved remnants of this belt occur today in Bolivia, Peru, Argentina, Antarctica and eastern Australia (Lock, 1980). The CFB is currently believed to have developed as a single phase, multiple event orogen during the Permian and Triassic (Theron, 1961; Hälbich, 1983a-c, 1992; Hälbich *et al.*, 1983; Veevers *et al.*, 1994), with up to five co-axial compressional deformation events, accompanied by low grade metamorphism, recognised and dated (Hälbich *et al.*, 1983; Hälbich & Cornell, 1983; Hälbich, 1992; Gresse *et al.*, 1992). These compressional tectonic events are believed to

produce varying structural and sedimentary responses within the Karoo Basin (Rust, 1959, 1962; Turner, 1975a; Colo, 1992; Hancox & Rubidge, 1995; Groenewald, 1996).

Plate tectonic theory emphasizes horizontal lithospheric movements which induce vertical movements due to changes in crustal thickness, thermal character and isostatic adjustment. The necessary conditions for thrust belts to form and widen are horizontal shortening and the existence of topographic highs (Davis *et al.*, 1983). Horizontal compression may be created by coupling at the contact between the subducting and over-riding plate or by traction at the base of the lithosphere, caused by mantle circulation above the subducting plate (Mitrovica *et al.*, 1989; Wdowinski *et al.*, 1989).

The application of plate tectonic models to account for the development of the CFB and the main Karoo Basin is complicated by the position of the proposed convergent margin, some 1000km away from the thrust belt (Lock, 1978; deWit, 1992). Because of this problem, three main models have been proposed for the formation of the CFB and Karoo RFB. These are: the continent-continent collision model (Winter, 1984; Rames, 1986); the northward subduction of oceanic crust under the palaeo-Pacific margin of Gondwana (Lock, 1980; Smellie, 1981), which led to the flat plate subduction model (Lock, 1980); and orogeny without subduction or the ensialic model (Hülbich, 1983d). Opinions for the development of the CFB and Karoo Basin therefore differ in respect of the presence of a distinct magmatic arc, the site and nature of the subduction zone, and even as to whether subduction was involved at all.

Two other examples of large retro-foreland basins (North American Cordillera and Andes of South America) are both known to have developed while they were underlain by subducted plates with very low angles of descent into the mantle (flat-subduction) (Allmendinger *et al.*, 1983). Cateneanu *et al.* (*in prep.*) furthermore believe that flat plate subduction is actually a pre-requisite for the formation of large scale retro-foreland basins and it therefore seems most likely that this was

also the case for the Karoo Basin. The cause and effect relationship between the angle of subduction and foreland shortening is not however clear at present (Lyon-Caen *et al.*, 1985).

Documentation and mathematical modelling (Quinlan & Beaumont, 1984; Stockmal & Beaumont, 1987) of RFBs have advanced dramatically in the past few years and the effects of basin flexure relative to thrust advance, sedimentary loading and source unloading, are now fairly well constrained (Johnson & Beaumont, 1995). Due to the vast amounts of work undertaken on retro-foreland basins in the past two decades, and particularly in the past five years, a review of the current status of these basins, their principal physical behaviours, and stratigraphic fills, is essential to a re-evaluation of the basin development of the upper Karoo.

4.5.2 Overview of retro-foreland basins

RFBs constitute one of the two principal occurrences of nonmarine strata in the rock record (Johnson & Beaumont, 1995). Such basins form along the continental interior of continental-margin orogenic belts and are large-scale (hundreds of kilometres wide and thousands of kilometres long), long duration (10-100 Ma) basins, which accumulate sedimentary strata at rates ranging from 0.1-1m/1000 years (Johnson & Beaumont, 1995). The thrust belt, foreland basin and stratigraphic fill is typically wedge shaped, being thickest adjacent to the core of the orogenic system, and tapering towards the cratonic interior. This usually leads to an asymmetric fill in cross sectional profile. Because of the nature of the subsidence and sediment supply, RFBs usually preserve an exceptionally complete stratigraphic record and geological time is represented primarily by accumulated strata. This however decreases down dip in a cratonward direction.

a) Physical principles of basin flexure and subsidence

The formation of a RFB results from the deformation of a foreland fold-thrust belt. In such thrust belts, horizontal shortening produces crustal thickening, which in

turn is responsible for a number of responses. Firstly isostatic adjustment to the increased mass causes topographic elevation within the thickened region (orogen), and a down-flexed foreland basin adjacent to this region. This is followed by an additional transfer of mass during the simultaneous erosion of the topographic highs, and the sedimentary filling of the foreland basin. The erosional transfer of mass from the thrust belt to the basin brings about isostatic uplift in the mountain belt and load driven subsidence in the basin, which amplifies the tectonically driven subsidence. This genetic link between foreland thrust systems and basin subsidence was first recognised by J. Barrell in 1917 (Price, 1973), however for nearly sixty years this finding was largely ignored. Price (1973) renewed interest in the concept of thrust-load-driven subsidence and following this paper, Beaumont (1981) and Jordan (1981) independently established the quantitative relationship between the mass distribution in a RFBs and the flexural and isostatic compensation of the thrust-plate mass.

As mentioned previously, horizontal movements are governed by plate motion, and vertical movements (both in the orogen and in the RFB) are governed by local isostatic adjustments. Such localised isostatic adjustment is only strictly true for a weak lithosphere, in which each column behaves independently of its neighbour. In most retro-foreland settings, the underlying continental lithosphere has lateral strength and isostatic compensation occurs over a broad area as regional or flexural compensation. A beam of greater flexural rigidity distributes the isostatic compensation over a broader distance and hence causes less local subsidence and a broader basin. Under viscoelastic conditions, the response of the lithosphere approximates the bending of an elastic beam under a vertical load (Beaumont, 1981).

This flexural model, first proposed by Beaumont (1981), simply states that lithospheric flexure caused by overthrust loading is sufficient to generate subsidence in a foreland basin. Subsequent works by Quinlan & Beaumont (1984), Stockmal *et al.* (1986) and Stockmal & Beaumont (1987) have geodynamically modelled foreland subsidence, as well as the resulting stratigraphic thicknesses of

foreland basin fill. These models quantitatively show that lithospheric flexure brought about by overthrust loading, coupled with the resultant topographical expression across the thrust belt, is an adequate mechanism for generating RFBs and their stratigraphic fills. Typically, only homogenous lithosphere and vertical loads (orogenic wedge and basin fill sediments) are considered in modelling retro-foreland geometries (Quinlan & Beaumont, 1984; Stockmal *et al.*, 1986; Flemings & Jordan, 1989, 1990) but more recently this scope has been increased to include lithospheric heterogeneity in the rootwall plate (Yang & Dorobek, 1995).

At some distance from the load, an upward deflection occurs known as the peripheral bulge or forebulge. Such flexural bulges normally have only limited topographic expression and rarely exceed tens of metres in elevation, yet they may move many kilometres laterally during their evolution. The position of the forebulge in active basins can only be confirmed by the position of a high gravity value, which reflect the upwards deflection of the base of the crust. In modern RFB settings, the forebulge rarely forms a topographic high and is therefore not expressed in the surface sedimentary facies patterns. There is however often a marked thinning over the forebulge (Johnson & Beaumont, 1995). The wavelength of the foreland flexural profile varies from basin to basin and is a function of: the strength of the underlying lithosphere (rheology), expressed as its flexural rigidity; the thickness and age of the lithosphere; basement tectonics; and the mass of the applied supracrustal load (Quinlan & Beaumont, 1984).

Jordan (1995) however warns that despite the theoretical justification of the flexural model, it remains unclear as to whether changes in flexural response occur on a sufficient vertical scale, and at short enough timescales, to be a factor in basin dynamics. She furthermore points out that most theoretical models to date consider flexural response only to tectonic loads, with the dynamic processes of erosion and sedimentation left out, and that flexural models do not fully approximate real systems. This is because they are designed based on homogenous materials, and do not account for lateral variation in the composition of the lithosphere and crust, which may have inherited zones of weakness. Royden

(1992) showed that areas of pre-existing weakness in the lithosphere may cause the forebulge to remain fixed in location, even whilst the load shifts in position. Yang & Dorobek (1995) further showed that intra-foreland structural discontinuities may partition strain during basin development, and influence regional subsidence and sedimentation patterns.

Flemings & Jordan (1989), Beaumont *et al.* (1993), Ettensohn (1994) and Cateneanu *et al.* (1997) have demonstrated that foreland basins actually include two distinct tectonic settings which behave differently during periods of tectonic loading (Pulse) and erosional unloading (Quiescence) in the orogenic belt. These are the foreland syncline (foredeep or forethick) adjacent to the orogenic load and the peripheral bulge (forebulge), away from the orogenic load. These authors have also shown that these basinal settings undergo reciprocal flexure in response to tectonic activity (Fig.4.6). Cateneanu *et al.* (1997) suggest an alternative terminology of proximal foreland for the foredeep, and distal foreland for the peripheral bulge (Fig.4.6).

During tectonic loading (P stage) the proximal foreland undergoes subsidence, whereas the distal foreland (peripheral bulge) is uplifted. During the following unloading (Q stage) the proximal foreland is uplifted by isostatic adjustment, whereas the distal foreland undergoes subsidence (anti-peripheral bulge or foresag). The proximal and distal foreland therefore shows reciprocal responses to tectonics, which is reflected in the fill. The area that separates these two tectonic settings is known as the hingeline (Fig.4.6).

In view of the overall hypothetical relationship of thrust systems to foreland basins, the subsidence history of the basin and its fill should reflect the history of thrust loading, and the cyclicity of movements in the thrust belt should be expressed as variations across the hingeline. The stratigraphic fill of the RFB therefore represents the preserved long term reciprocal responses of the depositional surface to prolonged subsidence.

In this new modelling, during the P stage, flexure creates subsidence in the proximal foreland and distal foreland uplift (peripheral bulge). This produces more pronounced unconformities at some distance from the orogen and across the peripheral bulge (Crampton & Allen, 1995). During the Q stage, uplift occurs in the proximal foreland and subsidence in the distal foreland (foresag). The Q stage is therefore predicted to produce unconformities within the proximal foreland which merge into the distal foreland. This is because the proximal foreland would be isostatically uplifted in response to erosion of the mountain belt (Beaumont, 1981) and sedimentary loading in the foredeep.

b) Surface modification and basinal fill

During the stratigraphic history of their fill, RFB's may be in an underfilled or overfilled condition. In an underfilled basin, subsidence is greater than sediment supply and the basin will take the form of a topographic trough, bound on one side by the orogenic belt and on the other by the peripheral bulge. In this instance drainage is usually longitudinal (parallel to the orogenic belt), with input from the elevated forebulge area as well. Underfilled basins usually occur where there is insufficient sediment supply from the source area, or if the transport system is too ineffectual to distribute the sediment across the basin. Overfilling may result from: a high rate of sediment flux; efficient transport of sediment across the basin (which is dependant on the type of transportational system and its discharge); a weak lithosphere; and/or a slow rate of thrust front advance. In this instance drainage is usually perpendicular to the strike of the orogenic belt and the sedimentary cover extends across the forebulge. In such a case the foreland basin is a lowland region, but is not a valley. Such overfilling commonly produces a basin that is wider than the zone of tectonic subsidence.

As has been shown above, long-term accumulation in a RFB requires tectonic subsidence, which may be generated by mass loading in the orogen during thrusting, or by isostatic adjustment during unloading. Flemings & Jordan (1989) made a significant advancement to the understanding of the fill of subsiding RFBs,

with the introduction of linear diffusive mass transport to represent the redistribution of mass via surface processes. Flemings & Jordan (1989), and Waschbusch *et al.* (1996) have also investigated whether intrinsic surface processes, or extrinsic tectonic controls are more important as primary stratigraphic controls. Surface processes redistribute the vertical orogenic load by erosion and may impart such a strong signature on the basinal fill that a tectonic signal may be obscured (Flemings & Jordan, 1989). A number of controls on stratigraphic fill (depositional environments) are furthermore independent of tectonic activity. These controls include: bedrock lithology; climate (both in the source area and in the depocentre); orographically controlled precipitation; geologic age (particularly in relation to the evolution of land plants); drainage patterns; river power; the balance between tectonic and erosional mass fluxes; and eustatic sea-level changes (Johnson & Beaumont, 1995). Flemings & Jordan (1989) indicate that on a theoretical level, intrinsic surface processes may be the major controls on stratigraphic fills, as the values of sediment supply and transport within the basin can control whether a basin is in an underfilled or overfilled state.

Another key variable in retro-foreland basins is the geometry of the sediment supply to the basin. In underfilled basins, sediment supply occurs from both the orogenic belt and the elevated forebulge. In overfilled basins on the other hand, there is no topographic expression over the forebulge and the only detrital supply is from the orogenic belt. Most RFB fill models are also based on two dimensional line sources, however sediment supply may also be from a number of point sources, both within the orogenic belt and the RFB.

From the preceding background it is evident that the first order control on basin subsidence behaviour and architectural fill in RFBs is complicated, and is dependant on a number of interlinked variables. Basin subsidence behaviour (base level change) is dependant on the applied load, the flexural rigidity and heterogeneity of the lithosphere (Beaumont, 1981; Yang & Dorebek, 1995) and intra-plate stresses (Poper, *et al.*, 1995) and may show reciprocal behaviour across the hingeline (Cateneanu *et al.*, 1997). The sedimentary fill of the basin is strongly dependant

on the creation of accommodation space, as well as on the nature of the drainage system and prevailing climate. Accurate basinal modelling therefore requires numerous inputs including factors such as: the lithospheric strength; the duration and timing of thrust activity; the attitude and rate of thrust advance; the depositional topography; the nature of the drainage system; the volume of sediment supplied; and the transport co-efficient from the mountain belt. Of the above factors, few are well constrained for the upper Karoo Basin at present, and for this reason the model which follows is somewhat simplified, and will doubtlessly require refinement with subsequent research. This model is however based on all available data from the literature as well as evidence accumulated during the course of this study.

4.5.3 New modelling of the development of the upper Karoo Basin during the Triassic

4.5.3.1 Introduction

Research into the formation and fill of retro-foreland basins has completely revolutionised the way in which RFBs are perceived, and in light of these new findings, the nature of the process-response models previously proposed for the main Karoo Basin during the Triassic will be re-evaluated.

In the models previously proposed for the upper part of the Karoo Basin (Rust, 1959, 1962; Visser & Dukas, 1979; Watson 1984; Colo, 1992; Smith *et al.*, 1993; Hancox & Rubidge, 1995; Greenewald, 1995), phases of active tectonism were believed to bring about increased relief and an immediate flux of coarse sediment into the basin (process-response model). Interbedded fines were considered to indicate times of tectonic quiescence. In these models the Katberg and Molteno Formations were proposed to represent the progradational response of the system to increased gradients associated with source area tectonism, whereas the Burgersdorp Formation was believed to represent a retrogradational (or sourceward) facies shift during a period of tectonic quiescence (Hancox & Rubidge, 1995).

The recognition that many modern sedimentary basins have accommodation space that requires filling led to the querying of the dogma that sedimentary responses to tectonic activity and the creation of accommodation space was rapid. This in turn produced the question of whether much of recorded basin fills actually post-date tectonic events (Soegard *pers.comm.*; Flemings & Jordan, 1989). Flemings & Jordan (1989) have shown that coarse grained clastic sediment may only prograde cratonward during tectonically quiet periods (Q), when the supply of sediment exceeds tectonically induced subsidence. This then contradicts the long held belief by many authors, that wedges of coarse clastic sediment within a sequence may be directly related to thrust/fault activity.

Applying this modelling (Fig.4.7), the arenaceous Katberg and Molteno Formations represent Q stage progradation (Q1 & Q2), whereas the predominantly argillaceous Balfour and Burgersdorp Formations form as a response to P stage tectonism (P1 & P2). In terms of this flexural model, the Triassic development of the Karoo RFB may be seen to be controlled by periods of limited tectonic activity or quiescence (Q), during which time facies prograded basinward, punctuated by pulses of tectonic activity in the source area (P), which created accommodation space and shifted facies sourceward (Fig.4.8). It also becomes evident that this modelling is exactly one phase out (180°) on the timing of the sedimentary response model of Hancox & Rubidge (1995) and the tectonostratigraphic model of Groenewald (1996).

The application of the flexural model to the fill of the upper Karoo Basin cannot however account for the presence of large extraformational clasts, fresh foldspar and coarse detritus in the distal sector, as the foredeep should preferentially retain coarse detritus. To account for the presence of these features in the Bamboesberg and Indwe Sandstone Members, the reciprocal stratigraphic model of Flemings & Jordan (1989), Ettensohn (1994), Cateneanu *et al.* (1997) needs to be applied.

4.5.3.2 Model for the development of the Triassic Katberg, Burgersdorp and Molteno Formations

This model interprets the Katberg, Burgersdorp and Molteno Formations in terms of modern sequence stratigraphic concepts, the flexural model for lithospheric bending (Beaumont, 1981; Quinlan & Beaumont, 1984) and the reciprocal stratigraphic architectural approach of Flemings & Jordan (1989), Ettensohn (1994), Cateneanu *et al.* (1997) and Cateneanu *et al.* (*in prep.*).

The only previous attempt at applying modern sequence stratigraphic concepts to the upper Karoo Basin during the Triassic is that of Groenewald (1996). In this work he adopted a tectonostratigraphic approach to model the fill of the basin during deposition of the Tarkastad Subgroup. This approach is not followed here as: it requires a totally new terminology; is based on process response models; and is applicable only to the nonmarine parts of the Karoo Basin, and may therefore not be readily integrated into the overall Karoo sequence. For a review of tectonostratigraphy and its application to the South African stratigraphic record, the reader is referred to Johnson & Rust (1988) and Winter (1990). Following Cateneanu *et al.* (1997) the Transgressive-Regressive (T-R) sequence terminology of Embry (1993, 1995) is preferred for nonmarine sequences.

Of the factors controlling basin development listed in the previous section: the lithospheric strength is inferred from previous work and basinal architecture; timing of thrust activity is based on the work of Hülbig *et al.* (1983), Hülbig (1992), and Gresse *et al.* (1992), as well as on the ages of the proposed sedimentary responses to these events; and the basin's depositional topography is determined from intra-basinal stratigraphic correlations and sedimentological studies. The facies genesis, climate and duration of fill follow the proposals of this thesis. Following Hülbig *et al.* (1983) and Gresse *et al.* (1992) four episodes of tectonic activity are of importance to this model. These episodes (P1-4) (Fig.4.7) equate to the second, third and fourth paroxysms of Hülbig *et al.* (1983), and to the second, third and fourth events of Gresse *et al.*, 1992.

With the introduction of the reciprocal flexural model for the generation of subsidence in RFBs, it has been shown that upward coarsening sequences in the distal foreland are evidence of periods of tectonic quiescence (Heller *et al.*, 1989; Cateneanu *et al.*, 1997).

Applying this model (Fig.4.7) the Katberg Formation may therefore be viewed as a basinward progradational response (TST) during a major period of quiescence (Q1) following cessation of compressional tectonic activity (P1) at 246 ± 2 (250) Ma. Towards the end of this period of quiescence, Burgersdorp Formation strata (*Cynognathus* Assemblage Zone, subzone A) were being deposited on the distal cratonward side of the basin. The next major period of dated tectonic activity in the CFB at ± 239 Ma (P2) (Gresse *et al.*, 1992), created accommodation space in the foredeep and elevated the foreland bulge such that during subzone B times distal, cratonward sedimentation was shifted sourceward (RST), and erosion occurred on the uplifted cratonward side. This trend continued with the base of subzone C restricted to the southern margins of the basin by orogenward foreland bulge migration. The decrease in the aerial extent of the basin during *Cynognathus* Assemblage Zone times (Fig.4.8) may therefore be seen as a response to tectonics in the CFB creating accommodation space in the proximal sector, and to the orogenward movement of the foreland bulge.

The 229 ± 5 Ma paroxysm of Hålbich (1981) is slightly anomalous, but may have been responsible for the formation of the regional erosional unconformity (SU) at the base of the Molteno Formation. Following the timescale of Harland *et al.* (1989) it would be dated at lowermost Carnian and following Ross *et al.* (*in prep.*) late Anisian or Ladinian. This event therefore seems to be dated either within the uppermost Burgersdorp Formation, in which case it may be responsible for the final deepening of the basin and orogenward shift of the forebulge, or within the period of development of the hiatal surface at the base of the Molteno.

The coarsening upward nature of the Bamboesberg and Indwe Members of the Molteno Formation is proposed to represent deposition entirely within the distal

sector (foresag or anti-peripheral bulge) (Fig.4.9). The formation of accommodation space in the foresag occurred during a major period of quiescence (Q2), during which time isostatic uplift occurred in the foredeep. The uplifted foredeep was therefore an area of sedimentary bypass (non-deposition) and erosion (Fig.4.9). During this period the distal sector was subsiding and gradients in the proximal sector increased due to isostatic unloading and erosional mass transfer from the CFB. These factors may account for the presence of coarse detritus and extraformational clasts of CFB and Lower Beaufort Group strata, in the distal sector, some 200km from their source area. That the sediment transport across the rising proximal sector was rapid (fluvial systems of high sediment transport efficiency), is evidenced by the presence of fresh feldspar in the distal sector.

The subsiding foresag therefore created accommodation space into which the Bamboesberg system prograded through time (TST) (Fig.4.9). During Bamboesberg Member times the extent of this subsiding foresag was constrained between the northern boundary of the Southern Cape Conductive Belt (SCCB), the southern boundary of the Kaapvaal Craton (KC) (Fig.1.2) and the Colesberg lineament to the west. As the Bamboesberg Member does not extend any further north than the southern boundary of the KC, either the supply of sediment was insufficient to outstrip foresag subsidence, or the lithospheric heterogeneity between the KC and the lithosphere underplating the foresag created a topographic high against which the Bamboesberg system onlapped, but did not overtop. Although fault controlled, a similar feature is seen today for the distal reaches of the Okavango Losimean system (Stannistreet & McCarthy, 1993) which terminates against the Tamalakane Fault. During Indwe Sandstone Member deposition (peak quiescence and proximal sector uplift) the supply of sediment probably exceeded subsidence in the foresag and the system prograded over the southern margin of the KC, to significantly increase the spatial extent of deposition of the Molteno Formation.

The tectonic event (P3) dated at 223Ma (approximately end Carnian) by Gresse *et al.* (1992) would have re-activated the upwarping at the southern boundary of the KC, restricting deposition to the southern sector, thereby causing cessation of

Indwe deposition and erosion in the uplifted northern part of the basin (SU). This re-activation of the structural feature at the southern boundary of the KC restricted the lower part of the "Transitional" Member (Mayaputi Shale Member; Turner 1975a) to the more proximal reaches of the foresag. During the following period of quiescence (Q3) the upper part of the "Transitional" Member (Glen Rossa, Transitional and Gubenxa Members; Turner, 1975a) prograded into the basin and as for the Bamboesberg Member, the aerial extent of the "Transitional" Member was initially controlled by heterogeneity between the lithosphere underplating the foresag and the Kaapvaal Craton. If the upper of the two coarse sandstones of the Moltene Formation in the northern preserved extremities of the basin is considered to be the upper "Transitional" Member (Gubenxa equivalent), then it may be seen that the progradation of the "Transitional" Member (TST) into the basin is identical to that modelled for the Bamboesberg-Indwe sequence. Following this scenario, during Gubenxa times sediment supply was greater than subsidence and the system once again prograded into the northern sector of the foresag. The last dated tectonic event (P4) at 215Ma (Hälbich *et al.*, 1983), brought about the cessation of Moltene deposition, and erosion in the uplifted northern part of the basin (SU).

4.5.3.3 Factors affecting the fill of the upper Karoo Basin during the Triassic

A central goal in stratigraphic studies is to discern the cause and controls on the sequence, be they extrinsic factors such as tectonism (thrusting and subsidence), eustasy or climate, or intrinsic changes in the depositional system. As has been shown above, the primary control on the fill of RFBs is the creation or destruction of accommodation space. Base level changes may however be brought about by either eustatic sea-level changes (Posamentier & Vail, 1988; Van Wagoner *et al.*, 1988) or tectonism (Johnson, 1995). Eustatic sea-level changes are the primary control on whether a RFB is marine or non-marine during its fill (Jordan, 1995). At times of high eustatic sea-level RFBs tend to be marine, with maximum water depths in the region of 500m (Jordan, 1995), whereas during periods of low eustatic sea-level, RFB's tend to be entirely non-marine. This is due to the fact that thrust-load driven subsidence is insufficient to submerge normal thickness

continental lithosphere during low eustatic sea-level (Jordan, 1995).

Stratigraphic sequences generated by episodic tectonism in nonmarine environments also differ significantly from those produced by eustatic sea-level changes. In completely nonmarine basins with variable thrusting, stratigraphic sequences are bound by major erosional unconformities in the distal foreland (Flemings & Jordan, 1989), whereas in classic eustatic controlled sequence stratigraphy, major subaerial erosional unconformities occur predominantly in the marginal and shallow marine realm during periods of eustatic lowstand. Flemings & Jordan's (1989) model shows that in nonmarine environments, subaerial unconformities occur only locally in the proximal areas, and regionally in the distal margins of the basin.

The fact that the upper part of the Karoo Basin is entirely non-marine, that the stratigraphic sequences in the distal foreland are bound by major subaerial unconformities, and that the Triassic was a period of relatively low eustatic sea-level, argues against eustatic changes as the primary control on the preserved sequences in the basin. Floodplain sediment storage and preservation also requires the creation of accommodation space (Wright & Marriot, 1993), and that this is tectonically driven for the Burgersdorp Formation is very evident from: the timing of thrusting in the CFB (Fig.4.7), nature of the sourceward facies shift during tectonism (Fig.4.8), and the diachronous nature of the unconformity at the top of the Formation. In a basin isolated from eustatic influence, the major control on the stratigraphic fill is tectonically generated changes in accommodation space (Johnson, 1995), and the creation/destruction of accommodation space in the Karoo Basin during the Triassic may be seen to be primarily controlled by reciprocal flexural responses to tectonism in the CFB. Other factors including inherited topography, basinal scale structural discontinuities, sediment flux and climate (Posamentier & James, 1993; Wasehbusch *et al*, 1996), may however have also played a significant role.

i) Inherited topography

LeRoux (1993) proposed that pre-existing topography may be created by the fact that deposited sands subside faster than muds and impart a created load. For this reason major channels often use the same path after tectonic adjustment. The fact that there is a continuous south-north trend for the sandstones of the upper part of the Karoo Basin (Katberg-Molteno Formations), coupled to the fact that many of the channel complexes of the Katberg and Molteno Formations cut down into one another, may be evidence for some type of inherited topography. The continuous south-north trend of the palaeocurrents is however more probably due to the fact that the basin was in an overfilled state during the Triassic, with drainage parallel to the strike of the orogen, and that the basinal axis was re-activated along inherited preferential zones of crustal weakness.

ii) Control by heterogeneity of the footwall

Various authors have shown that the footwall plate is not a homogeneous unit and that it may have different flexural rigidities (Quinlan & Beaumont, 1984), lateral variation and inherited zones of weakness (Royden, 1992), and intra-foreland structural discontinuities (Yang & Dorobek, 1995).

On a regional structural scale the thickest development of sedimentary strata of the upper Karoo during the Triassic (Katberg-Molteno Formations) occurs between the northern limit of the Southern Cape Constrictive Belt (SCCB) (and Cape Supergroup rocks) and the southern margin of the Kaapvaal Craton. The area between these two structural features formed the fore-sag during Bamboesberg and lower "Transitional" Member times, and is underlain by the eastern extension of the Natal-Namaqua Mobile Belt. These structural features therefore seem to impart a strong control on the areal extent and nature of the preserved fill of the main Karoo Basin during the Triassic.

The thinning and wedging out of the Bamboesberg Member in particular is most probably a reflection of the original basinal topography. This was primarily controlled by the differences in rheology between the KC and the SCCB, as well as the Colesberg Lineament in the west and the Natal embayment in the east. The east-west thinning may therefore be partly erosional, and partly due to the control imparted by the position of pre-existing zones of weakness. If pre-rift doming was active at this time, the eastern margin overlying the Cape Supergroup rocks of the Natal embayment may well have served as a structural control, as well as a source of sediment supply.

iii) Sediment supply and climate

From Figures 2.51, 3.51 and 3.66 it is evident that the nature of the source area supplying sediment to the Karoo basin changed during the Triassic. Sedimentological and palaeontological data also strongly suggest that this was accompanied by a change in the climate as well (Figs. 4.3; 4.4). This change from a semi-arid to humid climate would have affected the rate of breakdown of material in the source area, and the transport co-efficient to the depobasin. Because the efficiency of transport is dependant on the type of fluvial system and its discharge, the humid climate proposed for the lower part of the Moltene Formation would allow for more efficient sediment transport across the basin than would the semi-arid climate proposed for the Katberg and Burgersdorp Formations. Coupled to the change in tectonic setting, a change in climate may therefore have played an important role in allowing coarse sediment to prograde into the basin, and may account for certain of the differences between the depositional style and composition of the Katberg and Moltene Formations.

The fact that the limits and position of the forebulge and foresag remain semi-fixed in space during the deposition of the Burgersdorp and Moltene Formations however suggests that this area was an inherited zone of weakness, at the suture between the lithospheric plates that formed the Kaapvaal Craton and the younger, more viscoelastic lithosphere that underplated the Karoo Basin. This in turn would imply

that the tectonically controlled flexural profile strongly influenced the fill, and that climatic and sedimentological factors were not the primary control of the fill of the main Karoo Basin during the Triassic (Waschbush *et al*, 1996).

4.5.4 The question of a paraconformity in the upper Burgersdorp Formation

Rust (1959, 1962) and Stavrakis (1980a) both proposed that a major "paraconformity" existed within the Burgersdorp Formation, which separated the Katberg tectonosedimentary cycle from the proto-Molteno cycle. This conclusion was based on the fact that these authors believed the upper Burgersdorp and Molteno Formations to have been deposited under similar conditions.

Rust (1959, 1962) was however basing this similarity on the section at Bushmanshoek Pass, where his placement of the contact was incorrect, and he was therefore actually comparing the lowermost channel complexes of the Bamboesberg Member. Stavrakis (1980) based his conclusion on the similarity of sandstones within the upper Burgersdorp and lowermost Molteno Formation, an observation which is in line with the dominance of facies Sh in both sequences.

Although the Burgersdorp Formation sequence lacks basin-wide or regional concretionary layers, palaeosols or an angular discordance of strata, the difference in sedimentary style between the two Formations are significant. This is reflected in the differences of their channel complex geometries, architectural styles and facies associations (Table.4.5). Coupled to the difference in palaeontological signature between the two formations, there is sufficient evidence to show that a hiatal surface of between 4.5-6.1Ma actually separates the two Formations in the south of the basin, and that there was not continuous deposition across the contact.

Stavrakis's (1980) conclusion was however based on a conformity of succession from the upper Burgersdorp Formation through to the Bamboesberg Member, which as has been shown is not actually the case. What Stavrakis (1980) actually

recognised is that the deposits of the upper part of the Burgersdorp Formation differ significantly from those of the middle part of the Formation, a fact which may be accounted for by tectonically or climatically controlled changes in gradient and sediment supply.

4.6

Conclusions

This study has shown that the stratigraphy, sedimentology and palaeontology of the Burgersdorp and Molteno Formations are far more complex than was previously believed, and these findings stress the importance of combined stratigraphic, sedimentological and palaeontological studies in elucidating basinal fills.

The sedimentological studies of the Burgersdorp Formation undertaken for this work have shown that the Formation is broadly fluvial in its genesis and is dominated by high sinuosity mixed load meandering systems. Subordinate sheetwash, shallow ephemeral, and bedload dominated low sinuosity systems with tracts of in-channel braiding also occur in places. The overbank environments were dominated by suspended load deposition following flooding. A low water table is envisaged for the lower part of the Formation, with a net water deficit for the system giving rise to widespread calcrete formation. The presence of hydromorphic paleosols in the upper part of the Formation attests to slightly higher water tables and less net evaporative loss. Climatic conditions also changed through the deposition of the Burgersdorp Formation from semi-arid to sub-humid, and this change may well have controlled faunal and floral diversity within the basin.

Collecting in the Burgersdorp Formation has allowed for a better understanding of the palaeoecology of the *Cynognathus* Assemblage Zone, as well as allowing for refinements to the biostratigraphy. The recognition of faunal turnover in the *Cynognathus* Assemblage Zone allowed for the threefold biostratigraphic subdivision proposed by Hancox et al (1995) to be better understood. Although currently believed to contain faunal assemblages of both the *Lystrosaurus* and *Cynognathus* Assemblage Zones, the fauna represented in the rocks of the

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Burgersdorp Formation seems to be restricted to fossils of the *Cynognathus* Assemblage Zone. The palaeoflora of the Burgersdorp Formation is still not well known, but seems to be more variable and diverse than previously documented, with a number of new forms discovered during collecting. This palaeoflora seems to be transitional between that of the underlying *Lystrosaurus* Assemblage Zone and the overlying Molteno Formation.

Interbasinal correlation of the proposed subzones with faunas from other parts of Gondwana, Russia and China has shown that a number of these sequences also have more complex histories than previously documented. Study of the distribution of amphibian and dicynodont families have shown that provinciality first occurs only in the Late Triassic, and that during the Early and Middle Triassic, there was free gene flow between the southern and northern continents. The refined correlations allow for the age of the Burgersdorp Formation to be better constrained, and for the confusion regarding the age of the Formation to be resolved. The Burgersdorp Formation strata containing a *Cynognathus* Assemblage Zone fauna are Early (TR1; Scythian) to Middle Triassic (TR2; Anisian) in age, thereby encompassing the Early-Middle Triassic boundary.

The spatial and temporal extent and variability of the Bamboesberg and Indwe Members of the Molteno Formation have been documented and it has been shown that they both thin dramatically from south to north. The stratigraphic positioning of coals in the Bamboesberg Member has been assessed and their importance for understanding the temporal changes that occur in the Bamboesberg discussed. The lower part of the Molteno Formation is envisaged as having been deposited on a broad braidplain by bedload dominated, low sinuosity systems, under seasonally warm to humid climatic conditions, which were similar to the global conditions (Fig.4.4). Overbank environments were limited in their extent, and for the upper part of the Bamboesberg Member were dominated by peat swamps. The poor quality of the coals in the Bamboesberg and Indwe Sandstone Members may be directly attributable to the input of significant amounts of detrital elastic material, to fluctuations in the water table, and to the nature of the palaeoflora which

formed the peat. Based on interbasinal correlation and the age of the underlying *Cynognathus* Assemblage Zone (subzone C), and overlying lower Elliot Formation (*Euskelosaurus* Assemblage Zone) the age of the lower part of the Molteno Formation has been constrained as Carnian.

The nature of the basal contact of the Molteno Formation in the main Karoo Basin has been defined, and it has been shown to be highly diachronous. The basal hiatus represents at least the entire Ladinian and estimates of the magnitude of this surface range from 4.5-6Ma in the south of the basin, to between 13-33Ma for the north of the basin. The hiatal surface between the Burgersdorp Formation and the Indwe Sandstone Member of the Molteno Formation in the north of the basin therefore represents the greatest magnitude subaerial unconformity in the main Karoo Basin during the Triassic. This surface and the overlying fill of the Molteno Formation, may be linked to a major period of change in the tectonic style of southwestern Gondwana, prior to fragmentation in the early Jurassic.

Spatial and temporal distribution studies of the amphibians of the *Cynognathus* Assemblage Zone, and their associated faunas, evidences a gradual north to south decrease in the aerial extent of the basin during deposition of the Burgersdorp Formation. This was concomitant with the deepening of the proximal foreland and the orogenward migration of the peripheral bulge during a period of thrusting (P2) in the CFB.

The cratonward progradation of the lower part of the Molteno Formation may be linked to a period of quiescence in the thrust belt (Q2), and isostatic uplift in the proximal sector (Fig.4.9). Sedimentation during Bamboesberg and Indwe Sandstone Member times therefore occurred exclusively in the distal foreland (foresag). This model may for the first time explain the presence of very coarse sandstones, large extraformational clasts of CFB origin, and fresh foldspurs at a distance of hundreds of kilometres from the source area. Previous models for the development of the upper Karoo Basin which linked orogenic activity in the CFB to the progradation of coarse detritus into the basin were therefore 180° out of phase.

The fact that the drainage is perpendicular to the strike of the orogenic belt for the entire Triassic suggests that the main Karoo Basin was overfilled during this time. Sediment supply was predominantly from the orogenic belt, but the basin may also have received detritus from point sources within, and on the flanks of the basin. The effects of doubly vergent orogens (Willet *et al.*, 1993), and the interplay of compressional and extensional tectonics in different parts of the basin, both require further investigation.

The model presented for the development of the main Karoo Basin during the Triassic shows that the primary cyclicity is the periodicity of tectonism in the CFB, which controlled flexurally induced reciprocal subsidence. It is also recognised that the heterogeneity of the footwall lithosphere may have played an important role in the positioning of the forebulge/foresag region. Modifying controls on sedimentation include the existing surface topography, continental scale drainage patterns, climate, sediment flux and changing source area geology.

To conclude, this study has shown that the combined use of stratigraphic, sedimentological and palaeontological data, allows for far more precise interpretations of the physical and biological events which occurred during the Triassic development of the main Karoo Basin, than could any study undertaken in isolation.

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