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Reciprocal flexural behaviour and contrasting stratigraphies: a new basin development model for the Karoo retroarc foreland system, South Africa

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ABSTRACT

The main Karoo Basin of South Africa is a Late Carboniferous–Middle Jurassic retroarc foreland fill, developed in front of the Cape Fold Belt (CFB) in relation to subduction of the palaeo-Pacific plate underneath the Gondwana plate. The Karoo sedimentary fill corresponds to a first-order sequence, with the basal and top contacts marking profound changes in the tectonic setting, i.e. from extensional to foreland and from foreland to extensional, respectively.

Sedimentation within the Karoo Foreland Basin was closely controlled by orogenic cycles of loading and unloading in the CFB. During orogenic loading, episodes of subsidence and increase in accommodation adjacent to the orogen correlate to episodes of uplift and decrease in accommodation away from the thrust-fold belt. During orogenic unloading the reverse occurred. As a consequence, the depocentre of the Karoo Basin alternated between the proximal region, during orogenic loading, and the distal region, during orogenic unloading. Orogenic loading dominated during the Late Carboniferous–Middle Triassic interval, leading to the accumulation of thick foredeep sequences with much thinner forebulge correlatives. The Late Triassic–Middle Jurassic interval was dominated by orogenic unloading, with deposition taking place in the distal region of the foreland system and coeval bypass and reworking of the older foredeep sequences.

The out of phase history of base-level changes generated contrasting stratigraphies between the proximal and distal regions of the foreland system separated by a stratigraphic hinge line. The patterns of hinge line migration show the flexural peripheral bulge advancing towards the craton during the Late Carboniferous–Permian interval in response to the progradation of the orogenic front. The orogenward migration of the foreland system recorded during the Triassic–Middle Jurassic may be attributed to piggyback thrusting accompanied by a retrogradation of the centre of weight within the orogenic belt during orogenic loading (Early Middle Triassic) or to the retrogradation of the orogenic load through the erosion of the orogenic front during times of orogenic unloading (Late Triassic–Middle Jurassic).

INTRODUCTION

Tectonic setting

The Karoo Basin is a retroarc foreland basin developed in front of the Cape Fold Belt (CFB), in relationship to the Late Palaeozoic–Early Mesozoic subduction episode of the palaeo-Pacific plate underneath the Gondwana plate (Lock, 1978, 1980; Winter, 1984; de Wit *et al.*, 1988; Johnson, 1991; de Wit & Ransome, 1992; SOEKOR, 1996; Fig. 1). In a regional context, the CFB was part of the more extensive Pan Gondwanian Mobile Belt generated through compression, collision and terrain accretion along the southern margin of Gondwana. The associated foreland basin, subsequently fragmented as a result of Gondwana break-up, is preserved today in South America (Parana Basin), southern Africa (Karoo Basin), Antarctica (Beacon Basin) and Australia (Bowen Basin).

The Cape Orogeny developed along Late Proterozoic structural trends following the weakest and most

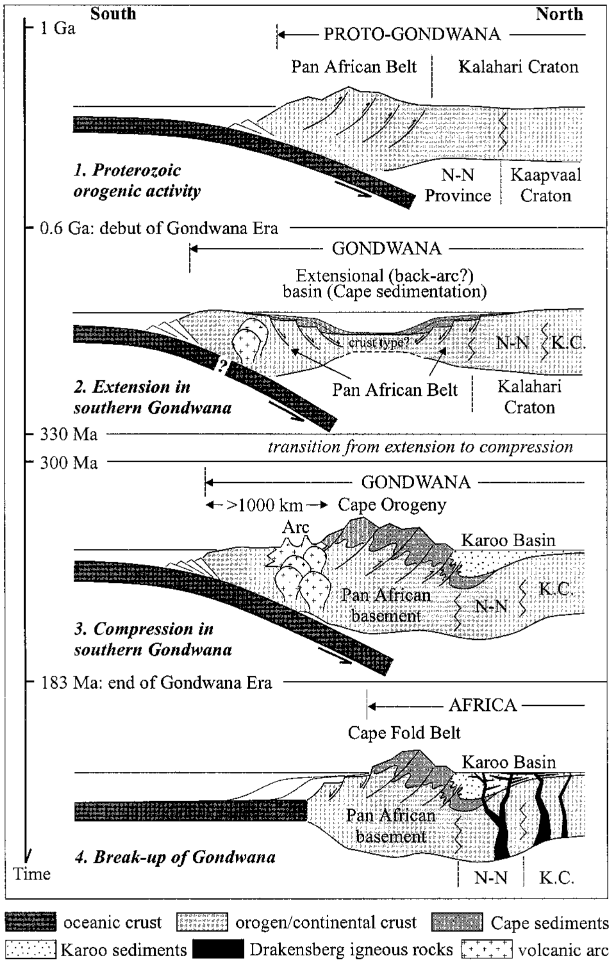

Fig. 1. Crustal evolution of southern Africa. N-N = Namaqua–Natal; K.C. = Kaapvaal Craton.

deformed zones of the continental lithosphere (Tankard *et al.*, 1982; Thomas *et al.*, 1992). Within this setting, the Karoo Foreland Basin developed in response to the supralithospheric loading generated as a result of crustal shortening and thickening in the Cape Fold Belt (Fig. 1). As the subduction took place underneath the basin, the Karoo qualifies as a retroarc (Dickinson, 1974) or retro-foreland setting (Johnson & Beaumont, 1995).

Geological background

The main Karoo Basin (Fig. 2) forms the thickest and stratigraphically most complete megasequence of several depositories of Permo-Carboniferous to Jurassic age in south-western Gondwana. The maximum preserved thickness of this megasequence adjacent to the CFB exceeds 6 km and the sedimentary succession reflects changing environments from glacial to deep marine, deltaic, fluvial and aeolian (Smith, 1990; Smith *et al.*, 1993). Basinal fill is inherently linked to the orogenesis of the CFB (Halbich, 1983; Cole, 1992), which is believed to have formed as a single-phase, multiple-event orogen (Halbich, 1983). Eight coaxial compressional deformation events are recognized and dated (Halbich & Cornell, 1983; Halbich, 1983, 1992; Gresse *et al.*, 1992). These events produced varying sedimentary responses within the foreland setting, which in the past have been interpreted as having occurred within a single subsiding basin characterized by continuous base-level rise in any of its parts (Rust, 1959, 1962, 1975; Turner, 1975; Cole, 1992). In this interpretation, all the stratigraphic features have been explained using the interplay between varying subsidence and sedimentation rates in time and across the basin (Cole, 1992), ignoring the e ects of the flexural response of the foreland lithosphere to the orogenic tectonism. Recent research and quantitative modelling of the formation and fill of peripheral and retroarc foreland basins has investigated the way in which these basins develop (Jordan & Flemings, 1991; Sinclair *et al.*, 1991; Sinclair & Allen, 1992; Watts, 1992; Beaumont *et al.*, 1993; DeCelles & Giles, 1996). Figure 3 illustrates the patterns of subsidence and uplift within the foreland system, which are controlled by stages of orogenic loading and unloading in the orogenic belt. As suggested by the evolution of the flexural profile (Fig. 3A), opposite base-level changes develop between the proximal and distal regions of the foreland system relative to the flexural hinge line. Surface profiles are obtained by adding the e ect of sedimentation to the flexural profile (Fig. 3B). They indicate that the depocentre of foreland sedimentation alternates between the foredeep and the foresag during consecutive stages of loading and unloading, respectively. The migration of the basin depocentre, together with the out of phase history of base-level changes, may generate contrasting stratigraphies between the proximal and distal regions of the foreland system as documented in recent case studies (Catuneanu *et al.*, 1997b; Catuneanu & Sweet, in press). In the light of these advances, a re-evaluation of the development of the main Karoo Basin may now be undertaken.

Aim of research

The purpose of this paper is to analyse the Late Carboniferous–Middle Jurassic sedimentary fill of the Karoo Basin in terms of: large-scale stratigraphic patterns; relationship between Cape Fold Belt tectonics and basin fill; main controls on stratigraphic sequences and distribution of depositional environments; influence of basement tectonics on basin evolution; position of the stratigraphic hinge line and the di erentiation between foredeep and forebulge settings; controls on the progradation and retrogradation of the foreland system; comparison of the Karoo Basin with other well-known retroarc foreland basins such as the Cretaceous Western Interior of North America.

We also investigate the timing of progradation of coarse-grained clastic material into the basin in relation to the periods of active tectonism or orogenic quiescence in the Cape Fold Belt. All the previous models for the evolution of the Karoo Basin stipulate a direct relationship between stages of tectonic uplift in the Cape Fold Belt

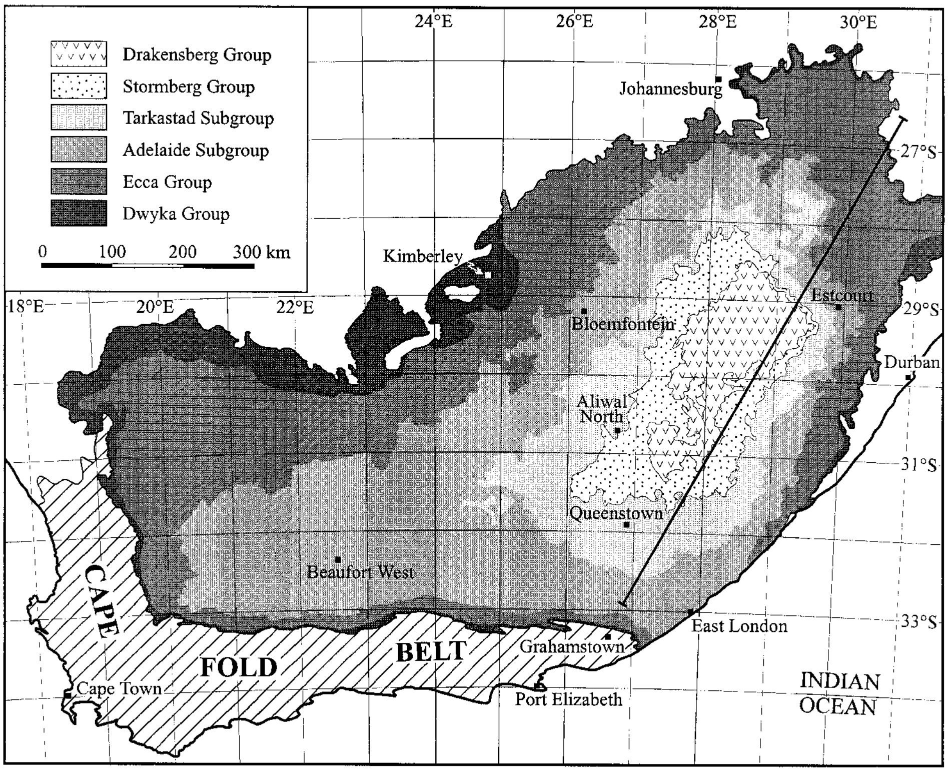

Fig. 2. Outcrop distribution of the main lithostratigraphic units of the Karoo Supergroup. The cross-section transect connects the proximal (south) to the distal (north-east) facies studied in this paper.

and the cratonward progradation of coarse sediments into the basin, although controversy surrounds the actual correlation between the coarse sedimentary wedges and various orogenic paroxysms. Our data suggest that the progradation of coarse sediments was alternately in-phase and out-of-phase relative to the orogenic paroxysms, as we refer to the southern and northern parts of the basin fill, respectively, which cannot be accounted for by the conventional models.

The time control within the Karoo Basin is achieved through assemblage zones of fossil reptilian fauna, which give a resolution averaging 2 Myr per zone (Rubidge, 1995). This allows us to perform an overall sequence analysis at the level of second-order cyclicity, which sets our target regarding the sequence stratigraphic framework of the Karoo foreland basin fill.

DATABASE

Lithostratigraphy

The Karoo Supergroup is subdivided into five main groups, i.e. the Dwyka, Ecca, Beaufort (Adelaide and Tarkastad subgroups), Stormberg and Drakensberg (Fig. 4). Apart from the igneous rocks of the Drakensberg Group, the rest of the Karoo Supergroup is composed of sedimentary rocks and is here referred to as the Karoo sedimentary sequence. A major stratigraphic gap corresponding to the late Anisian–Ladinian interval separates the Dwyka, Ecca and Beaufort groups from the overlying Stormberg strata. Along a dip-orientated profile (Fig. 5), the composite thickness of the Karoo sedimentary sequence increases significantly towards the Cape Fold Belt.

The distribution and age relationships of the major lithostratigraphic units of the Karoo Supergroup are illustrated in Figs 2, 4 and 5. Within the resolution of the reptilian assemblage zones used for biostratigraphic correlation, the relative age of the Karoo sedimentary sequences and their boundaries are well constrained (SACS, 1980; Eriksson, 1981, 1985; Smith, 1990; Cole, 1992; Visser, 1992; Rubidge, 1995; Hancox *et al.*, 1995, etc.). The age of the Drakensberg Group has been recently revised by Duncan *et al.* (in press) and attributed to the 183–179 $\pm$ 1 Ma interval based on ^{40}Ar–^{39}Ar and U–Pb methods.

From a lithostratigraphic point of view, the Karoo Basin can be subdivided into three distinct areas, characterized by di erent stratigraphies in terms of: facies, age of units, provenance and transport directions, and stacking patterns. These consist of a western area, west of the 24°E meridian; a southern area, east of the 24°E meridian (both adjacent to the CFB, representing proximal Karoo facies); and a north-eastern area, away from the CFB, representing the distal Karoo facies. These areas are separated by relatively narrow transitional zones in which significant lateral changes of facies occur. The proximal western and southern areas developed in front of the two branches of the CFB characterized by di erent orogenic strikes (NNW–SSE and W–E, respectively) and therefore individualized as distinct sources of sediment supply for the adjacent foredeep depocentres (Cole & Wipplinger, 1991). This explains the along-strike di erences between the western and southern foredeep stratigraphies. In this study we focus on the along-dip facies changes, occurring between the southern and north-eastern areas, aiming towards the di erentiation between the flexural foredeep and forebulge settings by mapping the stratigraphic hinge line position for consecutive time-slices.

A brief description of the main features of the Karoo sedimentary sequence follows.

Dwyka Group

The initiation of Dwyka sedimentation is estimated at about 300 Ma (Moscovian), following a 30-Myr stratigraphic break after the end of the Visean when the sedimentation in the Cape Basin was terminated (Visser, 1987; Cole, 1992). The Dwyka facies (tillites, cyclically grading upwards into finer-grained clastic rocks) indicates a glacial environment, with deposition from both grounded and floating ice. Although alternate deposition from grounded and floating ice has been recorded throughout the Karoo Basin, there is a clear distinction between the southern (proximal) and northern (distal) Dwyka successions. In the south, as many as nine fining-upwards cycles have been recognized (Visser, 1986), with thicknesses varying between 60 and 100 m. Each cycle displays a transition from terrestrial or subaqueous moraines, at the base, to glaciolacustrine shales at the top. A feature of the southern Dwyka succession is the uniform character and lateral continuity of the layers (Tankard *et al.*, 1982), suggesting that deposition from floating ice ('dropstones' supported by fine-grained matrix) within a large marine basin was the dominant process. In contrast, the lateral correlation of the northern Dwyka facies is very di cult due to the irregular thicknesses and complex facies relationships of the succession. Only two fining-upwards cycles have been identified in the Kimberly Britstown area of the northern Karoo, each of them comprising a basal massive till, a result of continental glaciation, grading upwards into a stratified terminal zone deposited from floating ice. The di culty in correlating the northern Dwyka layers, even between exposures only tens of kilometres apart, suggests local extension of grounded ice lobes accompanied by the development of ponds and outwash fans in adjacent areas (Tankard *et al.*, 1982). The limit between the grounded ice-dominated northern Dwyka and the floating ice-dominated southern Dwyka is placed south of Kimberley (Fig. 6). This di erence between the proximal and distal Dwyka successions, as well as the overall pattern of

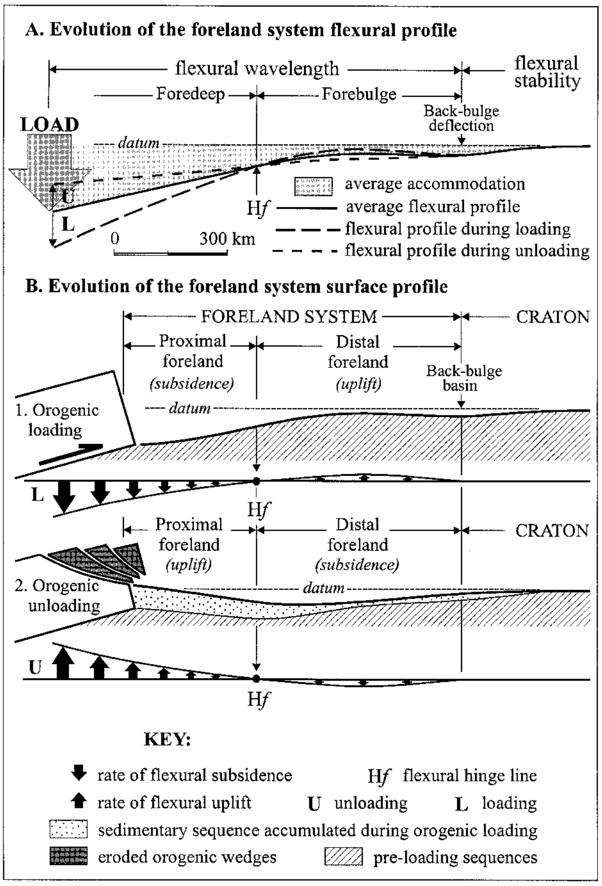

Fig. 3. Flexural and surface profiles illustrating the evolution of the foreland system during stages of orogenic loading and unloading (modified after Beaumont *et al.*, 1993; Waschbusch *et al.*, 1996; Catuneanu *et al.*, 1997a; Catuneanu & Sweet, in press). Not to vertical scale. The flexural wavelength depends on the rheology and elastic thickness of the lithosphere, basement tectonics, and mass and distribution of the applied loads. The given horizontal scale suggests a continental lithosphere with high flexural rigidity. If the foreland system develops on a less rigid and fractured basement, the foredeep may be narrower than 150 km, such as in the case of the Alps molasse basins (Homewood *et al.*, 1986; Crampton & Allen, 1995). (A) The flexural foredeep and forebulge undergoing out-of-phase subsidence and uplift in response to orogenic loading and unloading. During each flexural state, surface processes (sedimentation, erosion) tend to bring the foreland topography to the elevation of the adjacent craton (datum) allowing the subsequent mirror-image rebound of the surface profile. In (B), the proximal foreland illustrates the depositional foredeep (loading case) and a topographic slope dipping away from the orogenic load (unloading case). The distal foreland represents the topographic forebulge (loading case) and the depositional foresag (unloading case). The depocentre of the foreland systems alternates between the depositional foredeep during orogenic loading, and the depositional foresag during orogenic unloading. The coeval subsidence and uplift across the flexural hinge line may generate contrasting facies and stratigraphic patterns between the proximal and distal regions of the foreland system (Catuneanu *et al.*, 1997b; Catuneanu & Sweet, in press). The proximal and distal facies are separated by a stratigraphic hinge line whose position is not necessarily superimposed on but largely controlled by the flexural hinge line (Catuneanu *et al.*, 1997a). The migration of the flexural hinge line due to the redistribution of orogenic load is not represented.

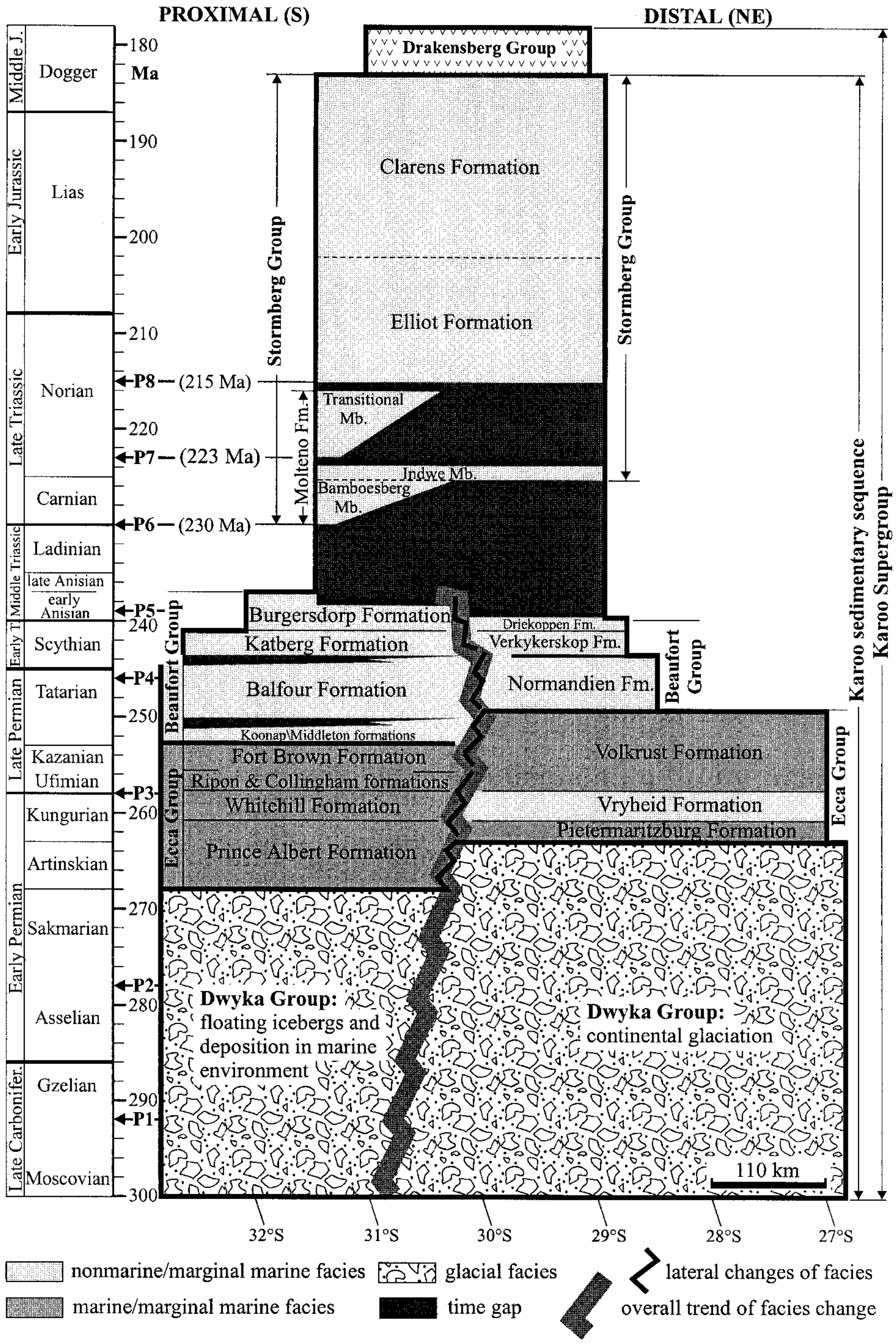

Fig. 4. Lithostratigraphy of the Karoo Supergroup, along the profile shown on Fig. 2. P1–8: tectonic paroxysms in the Cape Fold Belt, from Halbich (1983) and Gresse *et al.* (1992). The Beaufort Group includes the Adelaide Subgroup (Koonap, Middleton and Balfour formations in the south, and the Normandien Formation in the north) and the Tarkastad Subgroup (Katberg and Burgersdorp formations in the south, and Verkykerskop and Driekoppen formations in the north).

southwards ice movement (Tankard *et al.*, 1982), suggests the surface profile during the Dwyka time dipped to the south. This may also explain the diachronous age of the top of the Dwyka Group, younger in the north (Artinskian: Visser, 1989, 1992; Cole, 1992) relative to the south (Sakmarian: Visser, 1989, 1990; Cole, 1992), as the continental glaciation could have lasted longer in the north (at higher altitude) than in the south where the marine environment would have led to the rapid melting of the floating ice (Cole, 1992) during the climate warming in the Early Permian.

As a result of gradual deglaciation of the continental areas during the Artinskian, the northern Dwyka succession ends up with coal-bearing fluviodeltaic sequences (Smith *et al.*, 1993), overlain in places by the marine shales of the Pietermaritzburg Formation of the Ecca

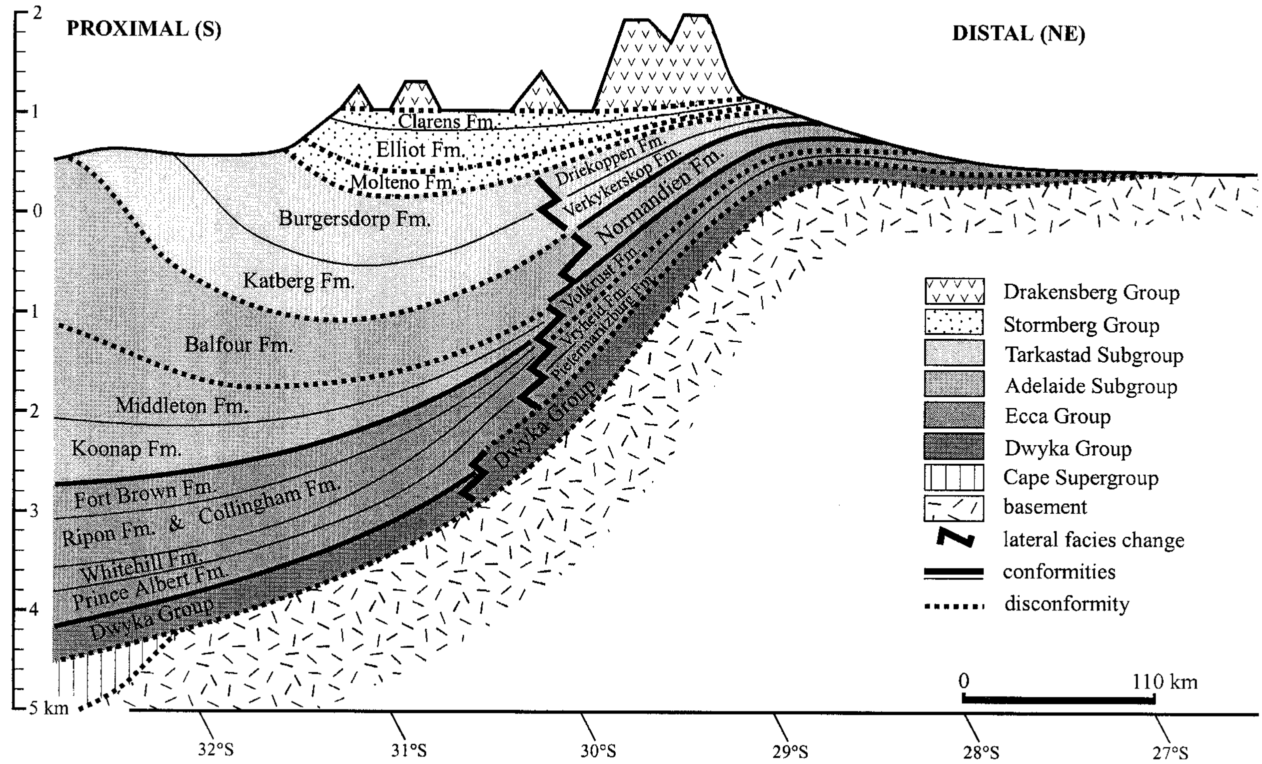

Fig. 5. Stratigraphic cross-section along the profile shown on Fig. 2.

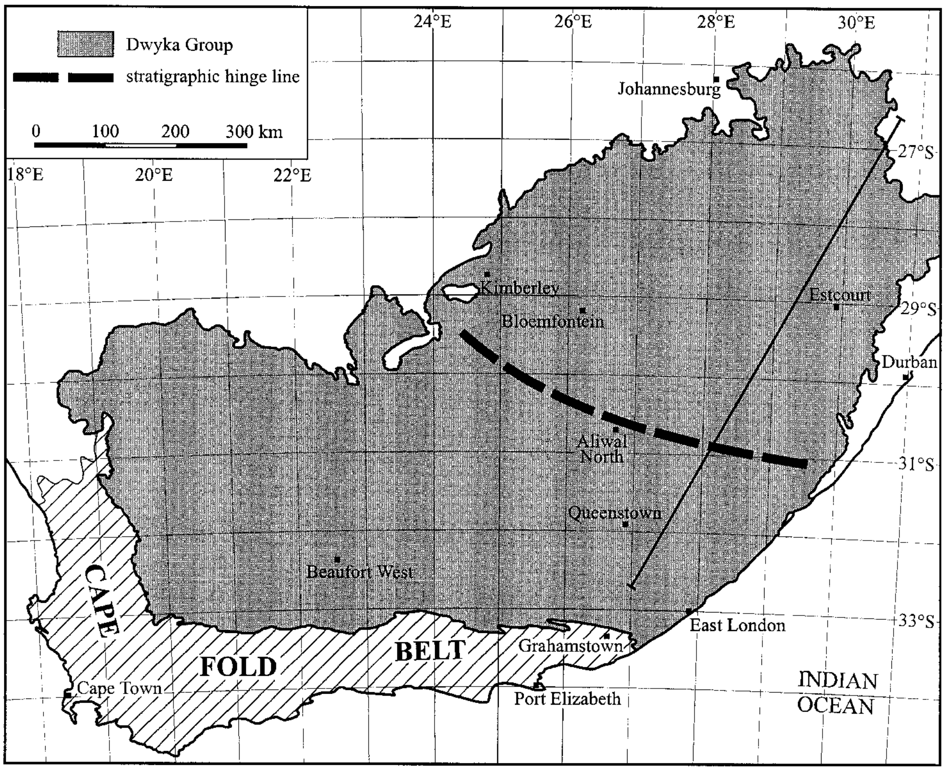

Fig. 6. Distribution map of the Dwyka Group (surface and subsurface) showing the limit between proximal (floating ice-dominated) and distal (grounded ice-dominated) facies, i.e. the position of the stratigraphic hinge line.

Group. In this case, the limit between the Dwyka and Ecca groups is sharp (Du Toit, 1954), being represented by a ravinement surface generated during the marine transgression of the Ecca Sea. In the south, precise delineation of the Dwyka–Ecca boundary is di cult as the transition from glaciomarine (Dwyka) to fully marine

(Ecca) environments was gradual, encompassing the time interval in which the floating ice was completely melted. In the transition interval between the Dwyka and Ecca groups, the percentage of dropstones present in the rock gradually decreases upwards in parallel with the disintegration of the floating ice.

Ecca Group

The southern Ecca stratigraphy includes the Prince Albert, Whitehill, Ripon, Collingham and Fort Brown formations, which are partially correlative to the Pietermaritzburg, Vryheid and Volkrust formations in the north-east (Fig. 4). The base of the Ecca Group is diachronous, as shown above. The top of the Ecca Group is also diachronous, younger in the distal sector relative to the proximal sector (Cole, 1992; Rubidge, 1995), which makes the upper part of the distal Volkrust marine facies correlative to the proximal nonmarine facies of the Koonap, Middleton and lowermost Balfour formations (Rubidge, 1995; Fig. 4).

The Prince Albert Formation is represented by dark greenish-grey shale with some graded silty layers, accumulated in a deep water environment. According to Visser (1992), the age of the Prince Albert shale is Artinskian–middle Kungurian (Fig. 4). The Whitehill Formation consists of carbonaceous shale, weathering white, with chert bands and lenses, deposited in a deep water, pelagic and reducing environment. The age of the Whitehill Formation is considered to be middle Kungurian to Ufimian, which makes it the lateral correlative of the distal Vryheid Formation (Visser, 1992). The Collingham Formation is interpreted as a distal submarine fan facies associated with pelagic sedimentation and wind-blown interbedded volcanic ash (Cole, 1992). The lithofacies displays alternating siltstone and shale with yellowish layers of tu . The age is Late Permian (Cole, 1992), probably Ufimian (Fig. 4). The Ripon Formation conformably follows the Collingham Formation, and consists of graywacke, siltstone and shale arranged into Bouma sequences deposited in a deep water proximal submarine fan facies (Visser & Loock, 1978). The age of the Ripon turbidites is also Late Permian (Cole, 1992), probably Ufimian (Fig. 4). The Fort Brown Formation is represented by greenish-grey shale with subordinate sandstone becoming more prominent upwards, deposited in an overall regressive shallow marine environment. The age of the Fort Brown Formation is Late Permian (Cole, 1992), probably Ufimian to Kazanian (Fig. 4) as the base of the overlying Koonap Formation is dated as Tatarian (Smith & Keyser, 1995).

In the northern Ecca, the Pietermaritzburg Formation consists mainly of shales accumulated in a moderate to deep marine environment, of Kungurian age (Visser, 1992). The Vryheid Formation is the product of a fluviodeltaic deposition that generated interbedded sandstone, shales and subordinate coal beds. This formation contains the only nonmarine sedimentary deposits in the Ecca Group in the north of the basin, and correlates with what appears to be the deepest proximal marine environment in the southern part of the basin (Whitehill Formation, middle Kungurian to Ufimian; Visser, 1992; Aitken, personal communication). The Volkrust Formation is composed predominantly of dark shales with intercalations of fine-grained sandstone, accumulated in a deep- to shallow-marine environment. The age of this formation is estimated as Umifian to Tatarian (combined information from Visser, 1992, and Rubidge, 1995).

All the bounding surfaces in the southern part of the basin (base and top of Ecca Group, as well as the limits between formations) are conformable. In the distal sector, the base of the Ecca Group is unconformable (ravinement surface), and so is the base of the Volkrust Formation in relationship to the transgression of the Ecca Sea over the nonmarine Vryheid facies. The rest of the bounding surfaces (top of Pietersmaritzburg and Volkrust formations) are conformable. The position of the stratigraphic hinge line separating the proximal and distal Ecca facies is represented in Fig. 7.

Adelaide Subgroup (lower Beaufort Group)

By Adelaide times nonmarine conditions of sedimentation were established throughout the Karoo Basin and lasted until the end of the deposition of the Karoo sedimentary sequence. The proximal facies of the Adelaide Subgroup includes the Koonap, Middleton and Balfour formations, whereas the distal facies is represented by the Normandien Formation (Fig. 4). The limit between the proximal and distal facies is illustrated in Fig. 8. The base of the Adelaide Subgroup is diachronous (Rubidge, 1995); however, it is conformable in both the proximal and the distal sectors, i.e. a facies contact from marine to nonmarine depositional systems.

The Koonap and Middleton formations together form a single fining-upward unit. The contact between the two formations is arbitrary and is based on a general change in lithofacies. The Koonap Formation (Tatarian) is dominated by greenish silty mudstones and sandstones organized in fining-upward cycles deposited in high-energy (braided river) systems grading upwards into lower energy (meandering) systems. In contrast, the Middleton Formation (Tatarian) includes maroon and greenish-grey mudstones interbedded with sandstones in an overall fining-upward succession deposited in low-energy (meandering and lacustrine) systems. The contact between the Middleton and the overlying Balfour Formation is unconformable, marking a sharp change from low-energy meandering facies with lateral accretion surfaces and abundant mud drapes within small-scale trough cross-stratification (uppermost Middleton Formation), to high-energy braided facies with large-scale planar cross-stratification in thick sandstone units (Balfour Formation).

The Balfour Formation constitutes one distinct overall

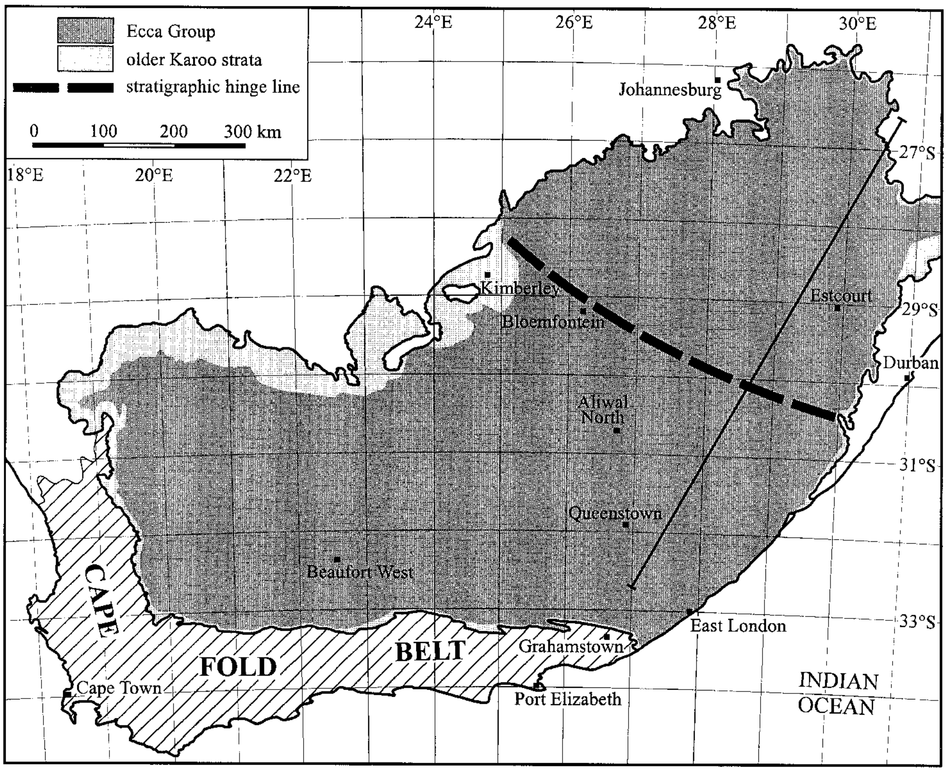

Fig. 7. Distribution map of the Ecca Group (surface and subsurface) showing the limit between proximal and distal facies, i.e. the position of the stratigraphic hinge line. The proximal facies include the Prince Albert, Whitehill, Collingham, Ripon and Fort Brown formations, whereas the distal facies refer to the Pietermaritzburg, Vryheid and Volkrust formations (Figs 4 and 5).

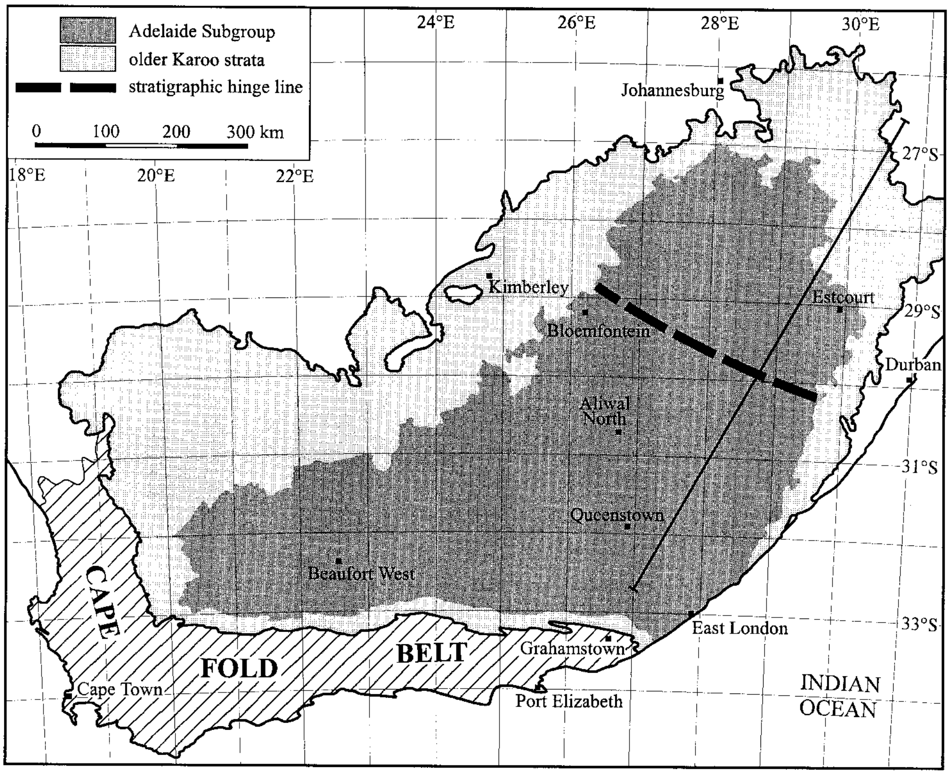

Fig. 8. Distribution map of the Adelaide Subgroup (surface and subsurface) showing the limit between proximal and distal facies, i.e. the position of the stratigraphic hinge line. The proximal facies include the Koonap, Middleton and Balfour formations, whereas the distal facies refers to the Normandien Formation (Figs 4 and 5).

fining-upward sequence bounded by subaerial unconformities both at the top and at the base. Yellowish and bluish-greenish-grey sandstones are interbedded with dark mudstones and organized in fining-upward cycles. Similar to the underlying sequence, the depositional environments changed from braided rivers grading upwards into meandering systems. The age of the Balfour sediments is Tatarian to early Scythian (Groenewald & Kitching, 1995; Kitching, 1995).

The Normandien Formation includes interbedded sandstones and mudstones deposited by meandering streams with channels flanked by wide semiarid floodplains (Groenewald, 1989). It correlates to the upper part of the Balfour Formation, and therefore the age is Tatarian to early Scythian (Groenewald, 1989; Rubidge, 1995). The top of the Normandien Formation is seen as conformable in northern Natal (Botha & Linstrom, 1978) and unconformable further to the north-east (Groenewald, 1989).

Tarkastad Subgroup (upper Beaufort Group)

A two-fold subdivision of the Tarkastad Subgroup has been proposed in both the proximal (S) and distal (NE) sectors of the Karoo Basin. The proximal facies includes the Katberg and Burgersdorp formations and is restricted to the southern margins of the basin from south of Queenstown to north of Aliwal North (Rubidge, 1995; Groenewald, 1996; Figs 4 and 9). The distal, northern facies (Groenewald, 1996) also includes two subdivisions; however, these are variously referred to di erent formations throughout the basin. In northern Natal they include the Belmont and Otterburn formations (Botha & Linstrom, 1978) whereas further north-east, in the Free State, they are termed the Verkykerskop and Driekoppen formations, respectively (Groenewald, 1989). Neither of these nomenclatures is formally recognized (SACS, 1980), but for the purpose of this paper, the terminology of Groenewald (1989) is followed for the northern facies (Fig. 4). The Tarkastad Subgroup is considered to be Scythian (Groenewald & Kitching, 1995) to early Anisian (Kitching, 1995).

The Katberg Formation unconformably overlies the Balfour Formation and is predominantly composed of thick, laterally extensive, light olive grey, coarse-grained sandstones, composed of transverse and longitudinal bar macroforms, which are internally structured predominantly by horizontal and trough cross-stratification. The nature of the sandstones and their internal fill suggest their deposition in a shallow braided environment with pulsatory discharge (Stavrakis, 1980; Hiller & Stavrakis, 1984). Thin sequences of red-olive yellow mudstones may also be preserved and probably represent abandoned channel fills and braidplain environments. The northern facies (Verkykerskop Formation) consist predominantly of thin, laterally extensive, medium- to fine-grained sandstones, dominated by transverse bar macroforms which are internally structured by planar cross-

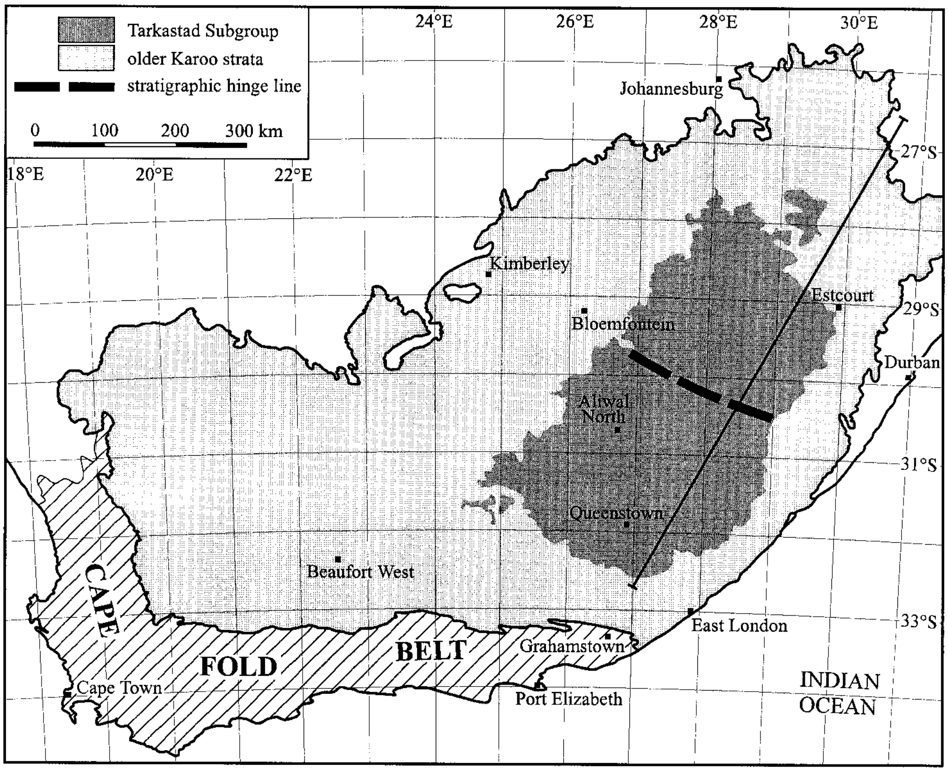

Fig. 9. Distribution map of the Tarkastad Subgroup (surface and subsurface) showing the limit between proximal and distal facies, i.e. the position of the stratigraphic hinge line. The proximal facies include the Katberg and Burgersdorp formations, whereas the distal facies refer to the Verkykerskop and Driekoppen formations (Figs 4 and 5).

stratification. The northern facies equivalents coarsen slightly to the south-west (Haycock *et al.*, 1997).

The Burgersdorp Formation conformably overlies the Katberg Formation and consists predominantly of thick fining-upward units of laterally inextensive, olive grey, fine- to medium-grained sandstones overlain by red-maroon coloured siltstones and mudstones. These fining-upward sequences are thought to represent mixed-load meandering river and floodplain deposits and preserve a fauna assignable to the *Cynognathus* Assemblage Zone (Hancox, 1998). The northern facies (Driekoppen Formation) is composed of thin fine-grained channel sandstones, internally structured by horizontal stratification, overlain by thick, massive to di usely laminated siltstones and mudstones. These deposits are believed to represent suspended-load-dominated meandering river deposits, and preserve a fauna assignable to the *Cynognathus* Assemblage Zone (Welman *et al.*, 1991). This fauna is dominated by aquatic forms including representatives of three families of amphibians, as well as by primitive archosauriformes (Welman, personal communication).

Overall, as in the case of the Koonap–Middleton and Balfour sequences, the Tarkastad Subgroup may be regarded as a single fining-upward sequence: sandstone-dominated braidplain deposits (Katberg and Verkykerskop formations), at the base, grade upwards into mudstone-dominated floodplain deposits associated with meandering systems (Burgersdorp and Driekoppen formations). The contact between the Tarkastad Subgroup and the overlying Molteno Formation is unconformable across the entire area of occurrence, whereas the limit between the lower and upper formations of the subgroup in both proximal and distal areas is conformable (gradual transition).

Stormberg Group

The Stormberg Group includes the Molteno, Elliot and Clarens formations. A major stratigraphic gap, corresponding to the late Anisian–Ladinian interval (SACS, 1980, Cole, 1992) separates these strata from the underlying Tarkastad and older sequences (Fig. 4). In contrast to the other Karoo groups and subgroups presented above, no proximal and distal facies are di erentiated within the Stormberg Group. In fact, the entire package of Stormberg deposits may be regarded as distal Karoo facies, as the Stormberg basin did not extend to the Cape Fold Belt but it was separated from it by a proximal region of syndepositional bypass and reworking of the older Karoo sequences (Fig. 10). Provenance studies on the Molteno sandstones indicate the Dwyka, Ecca and Beaufort rocks as sediment sources (Rust, 1959, 1962; Christie, 1981; Eriksson, 1984; Hancox, 1998), with the southern limit of the Stormberg basin close to the present-day preservation area (Cole, 1992).

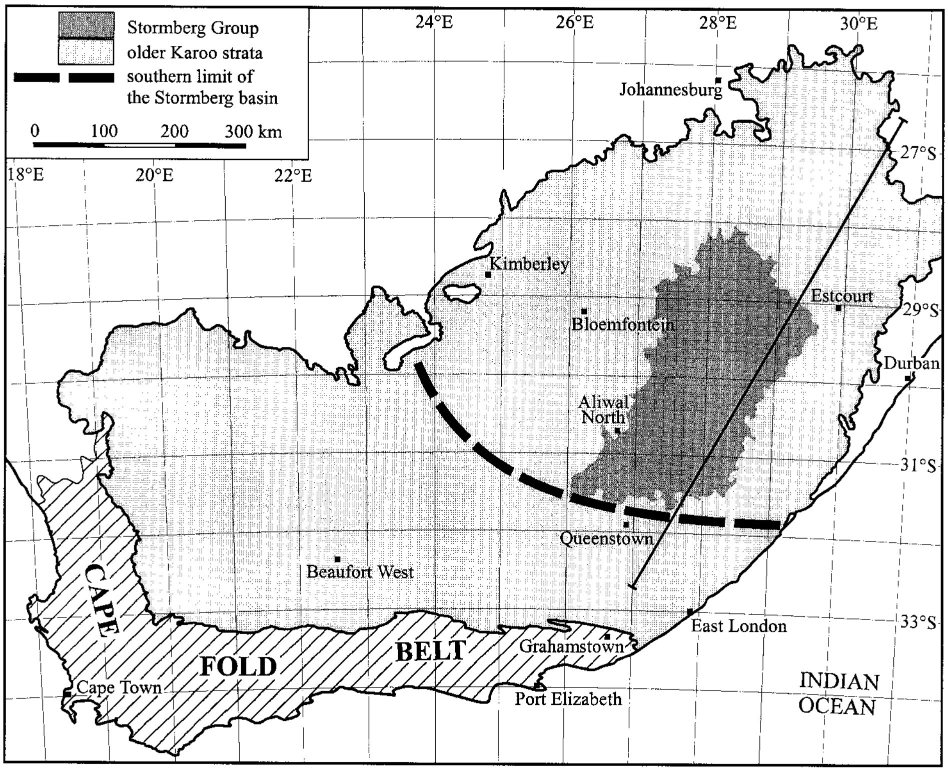

Fig. 10. Distribution map of the Stormberg Group (surface and subsurface) showing the southern limit of the Stormberg basin. This limit separates the area of Stormberg sedimentation, to the north, from the area of syndepositional erosion to the south. In this situation, the stratigraphic hinge line may be placed at the limit between the distal depositional area (base-level rise/subsidence and positive accommodation) and the proximal erosional area (base-level fall/uplift and negative accommodation).

The Molteno Formation (Carnian–Norian: SACS, 1980, Anderson & Anderson, 1993; Fig. 4) is composed of two major coarsening-upward sequences (Hancox, 1998). The basal of these is formed by the Bamboesberg and Indwe Sandstone members and the upper sequence by the Transitional Member. The formation is composed predominantly of tabular sheets of medium- to coarse-grained sandstone internally structured by horizontally and cross-stratified macroforms. These laterally continuous sheet sandstones are interpreted as having been deposited by braided streams on a vast braidplain. Siltstone, mudstone and coal deposits also occur but are far less abundant. These deposits are interpreted as the fills of abandoned channel tracts and within ponded bodies of water on the braidplain (Turner, 1975).

The Bamboesberg Member is dominated by olive grey fine- to medium-grained sandstones, internally structured equally by horizontal and trough cross-stratification. The overlying Indwe Sandstone Member is also dominated by sandstone, internally structured predominantly by trough cross-stratification. Other distinctive features of the two members include the presence of dm-scale clasts of Witteberg quartzites (Rust, 1959, 1962) within the sandstones (especially in the uppermost Bamboesberg and Indwe Sandstone members), and the large (mm- to cm-scale) crystals of feldspar within the Indwe sandstones (Turner, 1975; Christie, 1981). The Transitional Member also coarsens upwards with the contact between the Molteno and overlying Elliot Formation marked by the top of the uppermost coarse sandstone. This contact also marks a sharp palaeontological break in that it coincides with the first occurrence of fossils assignable to the lowermost biozone of the Elliot Formation, the *Euskelosaurus* Assemblage Zone (Kitching & Raath, 1984).

The Elliot Formation (Norian–Early Jurassic; Gau re, 1993) is dominated by reddish floodplain mudstones with subordinate channel and crevasse splay sandstones, believed to represent deposition within mixed-load-dominated meandering systems (Visser & Botha, 1980). In addition to the fluvial fine sediments, wind-blown sediment input is also added to the system introducing a loessic dust component (Eriksson, 1985). The aeolian influences increase upwards, making the transition towards the desert environment that dominated the deposition of the Clarens Formation. The occurrence of aeolian sandstones starts as m-scale thick intercalations in the upper part of the Elliot Formation, before the definitive establishment of the aeolian environment, showing cyclical changes in climate or aeolian sediment input in the transition interval between the Elliot and Clarens environments.

The Clarens Formation (Early to Middle Jurassic; Olsen & Galton, 1984) consists of cream or yellow fine-grained sandstones, sandy siltstones and mudstones with subordinate coarse-grained sandstones (Eriksson, 1984). Deposition took place in a desert environment, generating wind-blown dunes, as well as in wetter, less severe climate with numerous shallow playa lakes (Smith, 1990). Towards the end of Clarens deposition, the climate moderated to some extent and wet desert processes of stream and sheet flood became more dominant (Smith, 1990).

Considering the stratigraphic conformity and the gradual character of the transition between the fluvial-dominated (Elliot) and the aeolian-dominated (Clarens) environments, the Elliot and Clarens formations may be taken together as one coarsening-upward sequence.

Tectonic paroxysms in the Cape Fold Belt

During the evolution of the Karoo Basin, eight tectonic paroxysms have been documented in the CFB and dated using K–Ar and Ar–Ar techniques (Fig. 4): P1 (292 ± 5 Ma), P2 (278 ± 2 Ma), P3 (258 ± 2 Ma), P4 (246 ± 2 Ma), P5 (239 Ma), P6 (230 ± 3 Ma), P7 (223 Ma) and P8 (215 ± 3 Ma) (combined data from Halbich *et al.*, 1983; Halbich, 1983, 1992; Gresse *et al.*, 1992). The ages of the orogenic paroxysms have been obtained by dating the final phases of compression and metamorphism associated with each pulse of orogenic activity (cooling age of metamorphic minerals), which means that the paroxysms indicate the end of active stages of tectonism. Paroxysms P1 and P2 occurred during Dwyka deposition, P3 is coeval with the distal fluviodeltaic Vryheid sedimentation (the only nonmarine system of the Ecca Group), P4 occurred during Balfour sedimentation, P5 was coeval with the deposition of the Tarkastad Subgroup, P6 coincides with the base of the Molteno Formation, P7 equates to the base of the Transitional Member, and P8 occurred at the base of the Elliot (Fig. 4). We investigate the relationship between the CFB orogenic stages and the Karoo stratigraphy in the following sections of the paper.

INTERPRETATION

Flexural profile of the Karoo foreland system

An important issue in the analysis of the Karoo Basin is the identification of the flexural foredeep and forebulge settings, as well as their migration trends through time. The evolution of the foreland system flexural profile involves coeval subsidence and uplift relative to the flexural hinge line in response to stages of orogenic loading and unloading (Fig. 3A). Flexural processes combined with sedimentation determine the position of the basin depocentre in the proximal region (i.e. depositional foredeep) during orogenic loading, and in the distal region (i.e. depositional foresag) during orogenic unloading (Fig. 3B). The out of phase base-level changes between the proximal and distal regions result in contrasting stratigraphies, which may explain the stratigraphic hinge lines emphasized for the Karoo succession at consecutive time-steps (Figs 6–10). The five stratigraphic hinge lines are plotted together in Fig. 11.

Data presented under the *Lithostratigraphy* section indicate that the Karoo sedimentary sequence may be split into two distinct successions separated by the late Anisian–Ladinian disconformity: the Dwyka, Ecca and Beaufort groups, in the lower part of the sequence, and the Stormberg Group in the upper part (Fig. 4). The main stratigraphic feature that makes the distinction between the two successions is the degree of lateral facies change, much more pronounced in the former (Fig. 4).

The lower succession (Dwyka–Ecca–Beaufort groups)

Within the pre-Stormberg strata, significant lateral changes of facies occur between the proximal and distal reaches of the basin. For successive time-slices, the locus of the zone of facies change is not placed randomly within the basin, but it consistently follows a trend which records a pattern of NE-ward migration with time that turns around into a southward migration trend during the Beaufort time (Fig. 4). These changes of facies have been attributed to the interplay between varying subsidence and sedimentation rates across the basin (e.g. Cole, 1992). However, many stratigraphic features of the Karoo sequences point towards a more complex geodynamic evolution of the basin, with simultaneous manifestation of base-level rise and base-level fall within the foreland setting (proximal vs. distal) as the model of flexural response of the lithosphere to episodes of orogenic loading and unloading (Fig. 3) would predict.

The alternative correlation between proximal marine (Whitehill Formation)–distal nonmarine (Vryheid Formation) and proximal nonmarine (Koonap and Middleton formations)–distal marine (Volkrust Formation) facies argues in favour of the flexural model. In addition to that, the high diachroneity of some major lithostratigraphic boundaries, i.e. top of Dwyka and Ecca groups, suggests di erent geodynamic histories for the proximal and distal sectors. Even more diagnostic is the lateral extent of some of the major unconformities, restricted to either the proximal or the distal settings and with correlative conformities on the other side of the basin. For example, important distal unconformities occur at the base of the Pietermaritzburg and Volkrust formations, both of them having correlative conformities on the proximal side of the basin. Reciprocally, the unconformable contact between the Middleton and Balfour formations (proximal sector) has a correlative conformity in the distal sector within the Volkrust Formation. Similarly, the unconformable top of the proximal Balfour Formation correlates to a distal conformity (Figs 4 and 12). This pattern illustrates the concept of 'reciprocal stratigraphies' typical for retro-foreland systems subjected to superimposed supra- and subcrustal loading, with the proximal and distal regions corresponding to the out of phase flexural foredeep and forebulge settings (Catuneanu *et al.*, 1997a). Many stratigraphic features of the pre-Stormberg strata may now be explained using the model

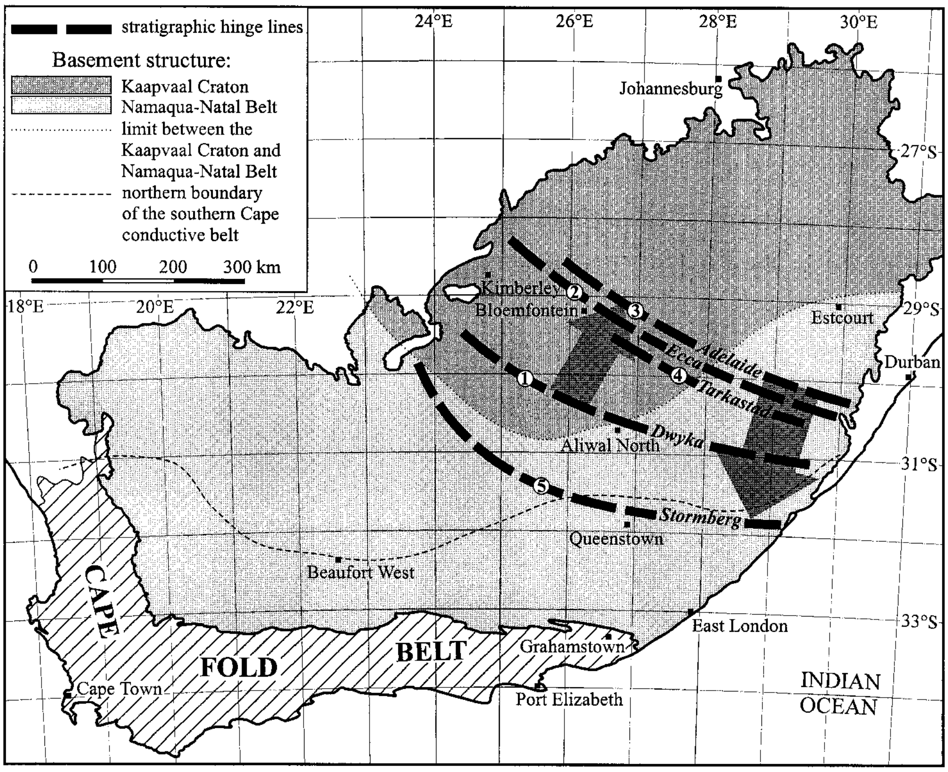

Fig. 11. Position of the stratigraphic hinge line at consecutive time-slices (1–5, in chronological order), pattern of hinge line migration (arrows), and basement tectonics. The limits between the Precambrian structures in the basement are from Visser (1995). The southern Cape conductive belt is a basement region defined by three major geophysical anomalies: a large positive static magnetic anomaly, a negative isostatic anomaly and high electrical conductivity (Pitts *et al.*, 1992).

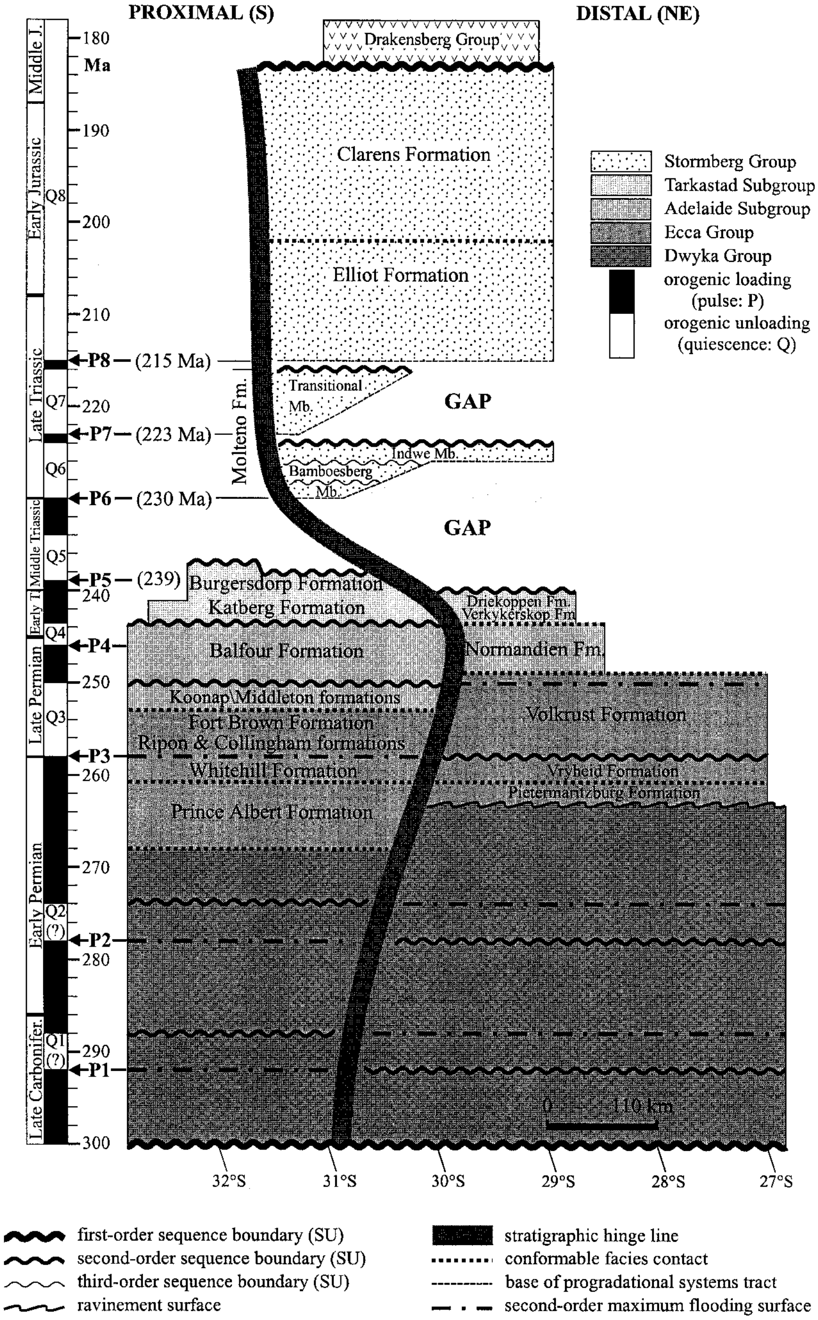

Fig. 12. Sequence stratigraphic framework of the Karoo first-order sequence in relationship to the orogenic cycles in the Cape Fold Belt, and the pattern of stratigraphic hinge line migration along the profile shown in Fig. 2. P1–8: tectonic paroxysms in the Cape Fold Belt, from Halbich (1983) and Gresse *et al.* (1992). SU: subaerial unconformity.

of reciprocal flexural behaviour of the foreland lithosphere (Fig. 3).

1 Dwyka Group (Figs 4, 6): the dominance of the ground ice in the north-east vs. the floating ice and the marine environment to the south could be accounted for by the initiation of the Cape Orogeny in the Late Carboniferous (Fig. 1) leading to a dominantly compressional regime and supracrustal loading in the orogenic area, which in turn implies proximal base-level rise (subsidence) coeval with distal base-level fall (uplift) in the foreland basin.

2 Dwyka-Ecca contact (Fig. 4): older in the proximal sector (Sakmarian–Artinskian limit) relative to the distal sector (Artinskian–Kungurian limit). This may be explained by proximal subsidence (base-level rise) coeval with distal uplift (base-level fall) as a result of orogenic loading. During such a stage, the initial Ecca Sea transgression would be restricted only to the foredeep, the forebulge area continuing to host the continental glacial environment.

3 Ecca Group (Figs 4, 5 and 7): lateral facies change between proximal Price Albert, Whitehill, Collingham, Ripon and Fort Brown formations, and distal Pietermaritzburg, Vryheid and Volkrust formations. Di erences between the proximal and distal settings are also suggested by the E–W trend of the upper Ecca isopachs in the proximal sector as opposed to NW–SE (parallel to the stratigraphic hinge line) in the distal sector (Cole, 1992). For the entire Ecca Group, sediment sources have been located within the CFB for the proximal deposition, with northward and north-eastward

transport directions, and within the Witwatersrand Arch (north of the basin) and the hypothetical 'Eastern Highlands' (east of the basin) for the distal deposition, with southward and westward transport directions (Cole, 1992). It is interesting to note that the position of the sediment sources was on the same side of the hinge line as the place of deposition, as the case of the uppermost Cretaceous Western Interior (fig. 16 in Catuneanu *et al.*, 1997b). Also noteworthy is the direct correlation between the proximal Whitehill Formation (deepest marine facies of the southern Ecca) and the distal Vryheid Formation (the only nonmarine facies of the northern Ecca, Fig. 4).

4 Ecca–Adelaide contact (Fig. 4): of Tatarian age, but older in the proximal sector than in the distal sector based on biostratigraphic evidence (Rubidge, 1995). This may be explained by proximal uplift (base-level fall) coeval with distal subsidence (base-level rise) as a result of orogenic unloading. During such a stage, the transition from marine to nonmarine environment would take place sooner in the foredeep area (Koonap and Middleton formations), whereas the marine environment would persist longer within the subsiding foresag (Volkrust Formation).

5 Adelaide Subgroup (Figs 4, 5 and 8): lateral facies changes between proximal Balfour Formation and distal Normandien Formation. Same comments related to sediment sources and transport directions as for point 3 apply here.

6 Tarkastad Subgroup (Figs 4, 5 and 9): lateral facies changes between proximal Katberg and Burgersdorp formations and distal Verkykerskop and Driekoppen formations, respectively. Isolated areas within the distal sector acted as intrabasinal sources for sediment supplied to the proximal sector (Cole, 1992), which is interpreted here as a result of forebulge uplift during times of orogenic loading. The extrabasinal sediment sources, as well as the transport directions, were the same as in the case of the Ecca sequence, supporting a di erentiation between the proximal and distal settings. The upper Tarkastad (Burgersdorp Formation and distal correlatives) is characterized by an overall finer-grained sedimentation throughout the basin, which is interpreted to be due to a generally low influx of sediments into the basin (Hiller & Stavrakis, 1984).

The upper succession (Stormberg Group)

Following the late Anisian–Ladinian stratigraphic gap, the post-Beaufort succession is distinctly di erent from the underlying Karoo sequences. Except for gradual decreases in the grain size with increasing distance from the sediment sources, no major or sharp lateral changes of facies occur in this upper succession (Fig. 4). The accumulation of the Stormberg sediments was coeval with the erosion of the older Karoo strata in the more proximal region of the Karoo Basin, as indicated by the Dwyka, Ecca and Beaufort sourced lithoclasts within the Stormberg strata (Rust, 1959, 1962; Christie, 1981; Eriksson, 1984; Hancox, 1998). The southern limit of the Stormberg basin was close to the present-day preservation area (Cole, 1992; Fig. 10), and probably separated regions subjected to base-level fall (proximal erosion) and base-level rise (distal deposition). This interpretation is consistent with the decrease in orogenic activity in the CFB during the Stormberg time (Halbich, 1983), which implies the dominance of orogenic unloading coupled with proximal uplift and distal subsidence in the foreland basin. In this light, the entire Stormberg Group may be interpreted as the product of foresag deposition during orogenic unloading (Fig. 3B). This conclusion is also supported by the correlation of the major stratigraphic gaps in the Stormberg succession with orogenic paroxysms in the CFB (pulses P7 and P8, Fig. 12), as stages of orogenic loading trigger uplift and erosion in the distal region of the foreland system.

Palaeocurrent analysis indicates sediment transport from southern and eastern sources (Cole, 1992), with the proximal area dominated by sediment bypass and erosion. The existence of cm- to dm-scale pebbles and cobbles of Witteberg quartzites derived from the Cape Fold Belt throughout the Stormberg occurrence area (Fig. 10), as well as the preservation of large crystals of feldspar (up to cm scale) in the sandstones of the Bamboesberg and Indwe Sandstone members of the Molteno Formation, confirms rapid bypass of an uplifted and steep proximal sector sloping towards the north, followed by deposition in the subsiding distal sector. The presence of the three major coarsening-upward sequences within the Stormberg succession (Bamboesberg–Indwe, Transitional and Elliot–Clarens) may be attributed to the gradual steepening of the proximal topographic slope during stages of orogenic unloading, leading to a corresponding increase in the energy level of the fluvial systems feeding the distal depositional area (Catuneanu & Sweet, in press).

The decrease in the aerial extent of the Stormberg Group relative to the lower Karoo succession has previously been interpreted to be due to a decrease in the size of the Karoo Basin through time (Cole, 1992), although proper explanations for this geodynamic behaviour have not been supplied. The recognition of the Stormberg strata as a result of foresag deposition may provide an alternative explanation for this phenomenon.

First-order sequence stratigraphy

Within the Karoo, sedimentation related to the tectonic setting of the foreland basin was closely controlled by the tectonic regimes manifested within the CFB. The tectonic history of the CFB commenced with a Late Carboniferous–Middle Triassic compressional stage (mainly thrusting and folding in the orogen) followed by a Late Triassic–Middle Jurassic orogenic relaxation/ quiescence (mainly normal faulting in the orogen; Halbich, 1983), and it was terminated with the initiation of Gondwana break-up in the Middle Jurassic. The CFB tectonic cycle is referred to as a first-order episode (de

Wit & Ransome, 1992), as being related to the assembly of the Pangaea supercontinent. The corresponding sedimentary succession accumulated within the Karoo Basin during the CFB cycle therefore represents a first-order stratigraphic sequence (Fig. 12), which is also compatible with the hierarchy system based on the magnitude of base-level changes generating the sequences, proposed by Embry (1995).

The Karoo first-order sequence correlates to a major shift in the tectonic setting, and is genetically related to the CFB-foreland system. According to this ranking system, the Karoo first-order sequence does not include the Drakensberg Group, which succeeds the CFB orogeny and relates to the Gondwana break-up extensional setting (Fig. 1), although from a lithostratigraphic point of view the Drakensberg Group is part of the Karoo Supergroup (Fig. 4). The lower boundary of the Karoo first-order sequence is represented by the major unconformity separating the Cape and Karoo supergroups, whereas the upper first-order sequence boundary is the contact between the Clarens Formation and the Drakensberg Group (Fig. 12).

The initial compressional stage (Late Carboniferous–Middle Triassic), characterized by building up of supracrustal load and in-plane stress, corresponds to a first-order orogenic loading and it is defined by the accumulation of thick foredeep sequences. The postcompressional stage (Late Triassic–Middle Jurassic), characterized by extensional and erosional o -loading in the CFB, corresponds to a first-order orogenic unloading and it is defined by foresag deposition (Fig. 3B). These two stages of the first-order CFB cycle are separated by the P6 tectonic paroxysm that marks the limit between the lower and upper Karoo successions (Figs 4 and 12).

Second-order sequence stratigraphy

The tectonic paroxysms dated in the CFB on the basis of the cooling age of metamorphic minerals (Halbich, 1983; Gresse *et al.*, 1992) signify the end of major compressional stages (resulting in thrusting, folding and supracrustal loading) identified here as second-order tectonic pulses as they provide the basic subdivision of the first-order CFB cycle (Fig. 12). The timing of cessation of these second-order pulses is known from radiometric age determinations (Halbich, 1992; Halbich *et al.*, 1983; Gresse *et al.*, 1992; see Database section), but their initiation can only be interpreted on the basis of foreland stratigraphy. The orogenic pulses are followed by stages of orogenic quiescence, so a number of eight second-order orogenic cycles of loading and unloading may be separated during the first-order CFB cycle (Fig. 12). Figure 12 also provides the second-order sequence stratigraphic interpretation of the Karoo sedimentary fill in relation to the orogenic tectonism. Sequence boundaries are placed at the end of uplift stages, which are out of phase between the proximal and distal regions of the foreland system (stages of orogenic unloading and loading, respectively; Fig. 3B).

The paroxysm P1 (292 ± 5 Ma) ends the first second-order tectonic pulse that initiated the supracrustal loading in the CFB and implicitly the sedimentation within the foreland basin: the base of the Dwyka Group is estimated at about 300 Ma (Cole, 1992), and therefore the first second-order compressional stage probably lasted about 8 (± 5) Myr. There is no control yet on the duration of the second and third second-order orogenic pulses, but it is likely that the debut of both of them took place during the Dwyka time (Fig. 12). The configuration of the Karoo Basin during the pulses P1, P2 and P3 is suggested in Fig. 13(A). Time-step (1) in Fig. 13 explains the marine and continental facies of the Dwyka glacial deposits, as well as the position of the Ecca Sea initially restricted to the proximal region of the basin. Time-step (2) shows the Ecca Sea transgression over the entire foreland area, which may be related to the melting of the continental ice caps. Time-step (3) explains the coeval deposition of the Whitehill proximal marine and Vryheid distal nonmarine facies (Figs 4, 12 and 13).

The base of the fourth second-order orogenic pulse is constrained by the erosional limit between the Middleton and Balfour formations as renewed foredeep subsidence was required to start the Balfour deposition, so a 4 (± 2) Myr duration is inferred for the pulse P4 (Fig. 12). Pulses P3 and P4 are separated by the stage Q3 of orogenic quiescence (Fig. 12), which explains the gradual regression of the Ecca Sea from the proximal region and the coeval deposition of proximal nonmarine and distal marine facies (Fig. 13B). The renewed orogenic loading during the pulse P4 can be associated with a new stage of clastic progradation within the basin and the accumulation of the Balfour and Normandien formations (Fig. 3C). Subsequent orogenic quiescence in the CFB (stage Q4 in Fig. 12), generating proximal uplift and distal subsidence within the foreland system, may explain the unconformable contact between the Balfour and Katberg formations (Figs 12 and 14D).

The base of the fifth second-order orogenic pulse is constrained by the debut of progradation of the Tarkastad sedimentary wedge (proximal Katberg and Burgersdorp, and distal Verkykerskop and Driekoppen formations; Fig. 12). The basin configuration during this stage is suggested in Fig. 14(E). The unconformable top of the Tarkastad sequence may be related to the subsequent stage Q5 of orogenic unloading, for the proximal region (Fig. 14F), and to the pulse P6 for the distal region (Fig. 14G).

The deposition of the preserved Stormberg strata is exclusively related to the foresag depocentre individualized during stages Q6, Q7 and Q8 of orogenic unloading (Figs 12 and 14H). This prolonged regime of orogenic quiescence (first-order orogenic unloading) was only interrupted by two short orogenic pulses (P7 and P8, Fig. 12) that generated forebulge uplift and erosion within the Stormberg basin, i.e. the subaerial

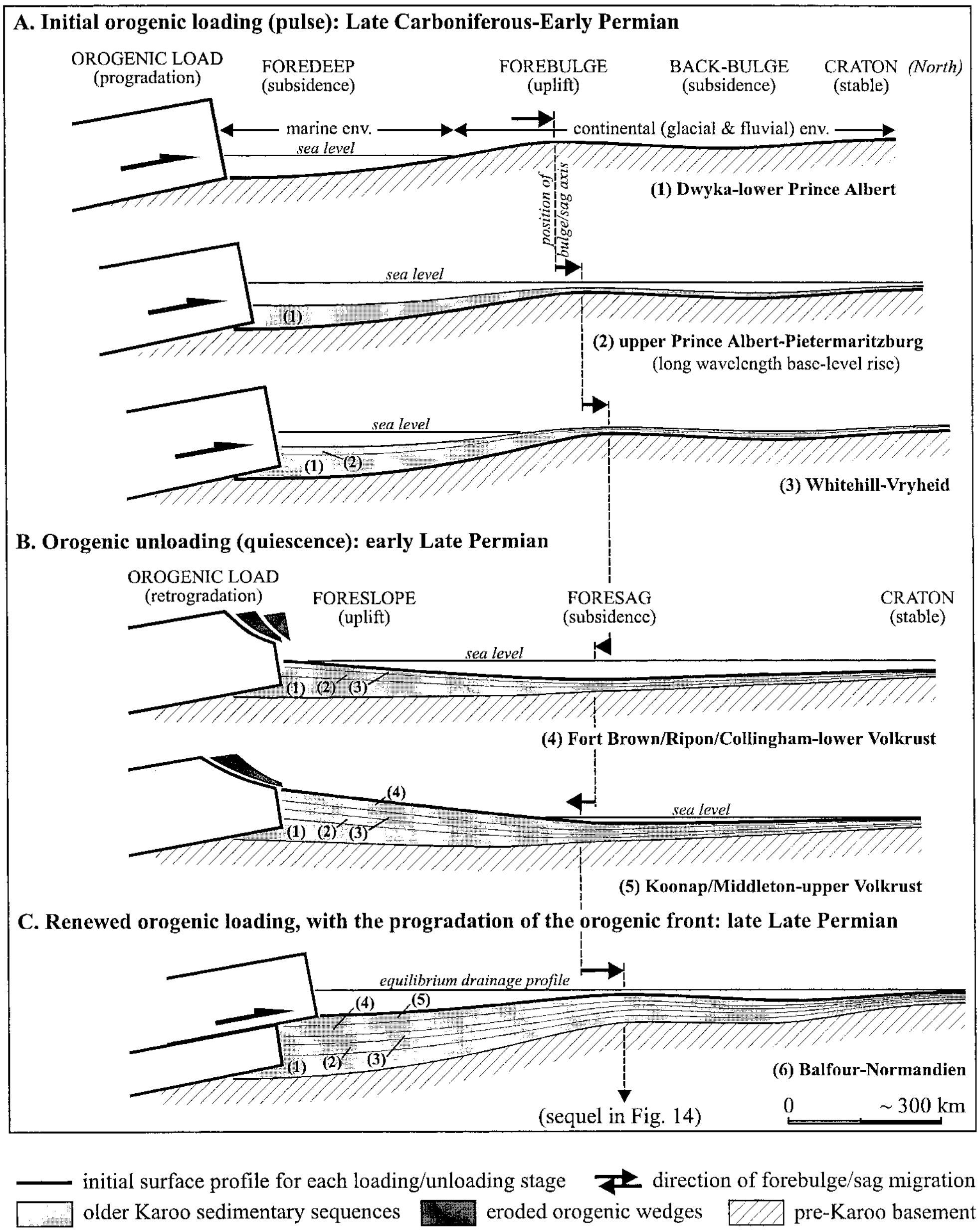

Fig. 13. Schematic model for the evolution of the Karoo sedimentary basin, Late Carboniferous–Permian: overall progradation of the foreland system. Not to vertical scale. The cratonward migration of the peripheral bulge is attributed to the progradation of the orogenic front, e.g. during time-steps (1), (2), (3) and (6). Stages A and C of orogenic loading are separated by stage B of orogenic unloading during which the retrogradation of the orogenic load due to the erosion of the orogenic front may cause an orogenward migration of the foresag. The depocentre of the Karoo basin is alternately the foredeep, during orogenic loading, and the foresag, during orogenic unloading. The six cross-sectional views illustrate the evolution of the surface profile, as well as the distribution of depositional environments (marine vs. nonmarine) during consecutive time-steps. For example, time-step (1) suggests the coeval deposition of marine and nonmarine Dwyka facies, as well as the debut of the Ecca Sea (lower Prince Albert Formation) restricted to the southern part of the basin, which explains the diachroneity of the Dwyka–Ecca contact (Figs 4 and 12). The Ecca–Beaufort contact is also diachronous (Figs 4 and 12), which is explained by the persistence of the Ecca Sea within the foresag area coeval with the debut of fluvial aggradation in the more proximal region (time-step 5). The basin is underfilled during time-steps (1)–(4), with deep marine sedimentation; it reaches a filled phase during time-step (5), with shallow marine–nonmarine sedimentation; and evolves into an overfilled phase during time-step (6), with fully nonmarine sedimentation across the entire basin. Time-step (1): basin configuration during the deposition of the Dwyka and lower Prince Albert sediments (pulses P1, P2 and P3 in Fig. 12). Time-step (2): basin configuration during the deposition of the upper Prince Albert and correlative Pietermaritzburg sediments (pulse P3 in Fig. 12). Time-step (3): basin configuration during the deposition of proximal Whitehill and distal Vryheid sediments (pulse P3 in Fig. 12). Time-step (4): basin configuration during the deposition of proximal Fort Brown/Ripon/Collingham and distal lower Volkrust sediments (quiescence Q3 in Fig. 12). Time-step (5): basin configuration during the deposition of the correlative Koonap/Middleton nonmarine and upper Volkrust marine sediments (quiescence Q3 in Fig. 12). Time-step (6): basin configuration during the deposition of the Balfour and correlative Normandien sediments (pulse P4 in Fig. 12).

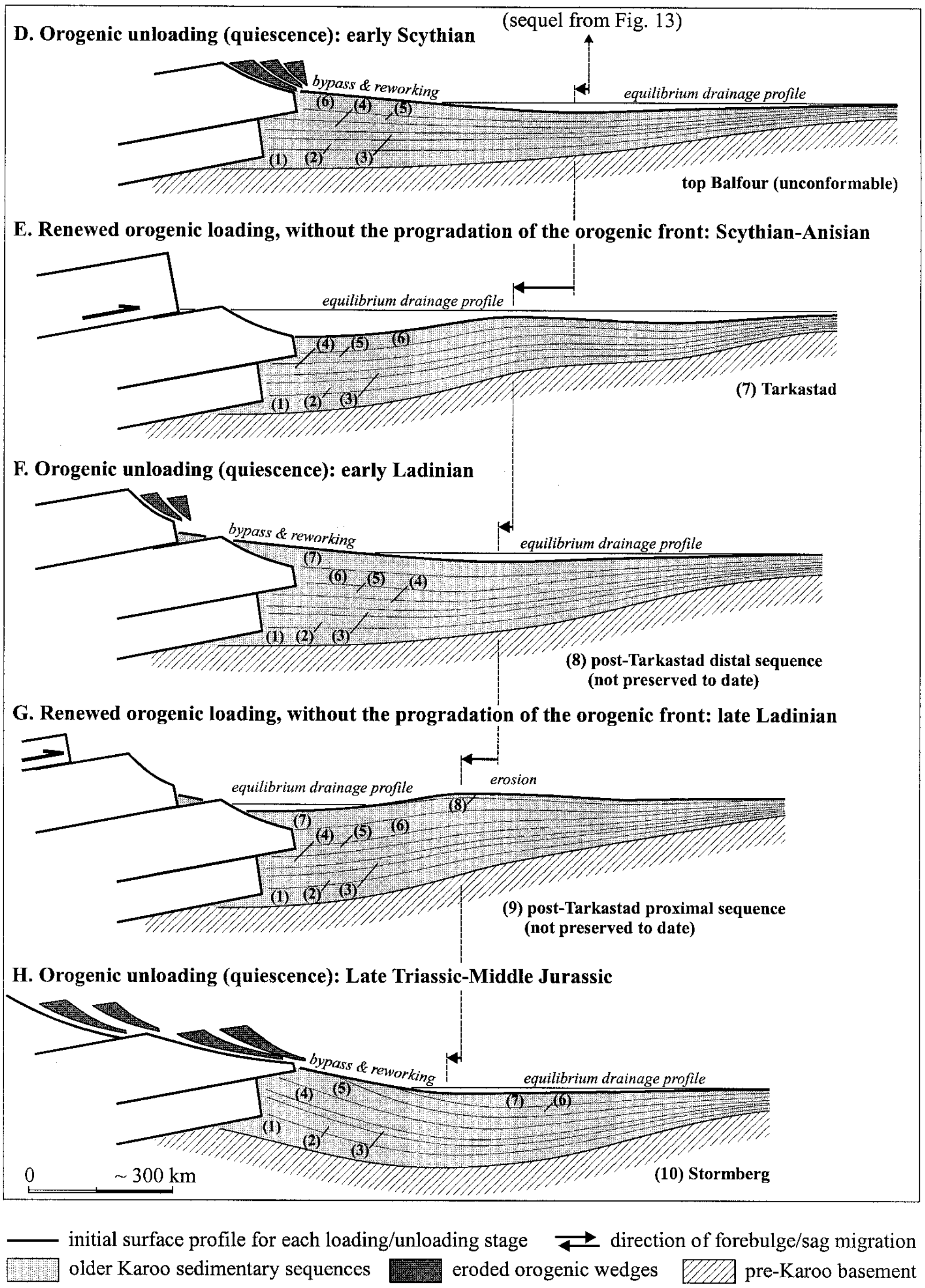

Fig. 14. Schematic model for the evolution of the Karoo sedimentary basin, Triassic–Middle Jurassic: overall retrogradation of the foreland system. Not to vertical scale. The orogenward migration of the peripheral bulge/sag may either be attributed to the retrogradation of the orogenic load due to the erosion of the orogenic front during times of quiescence (e.g. stages D, F and H of orogenic unloading) or to piggyback thrusting accompanied by a retrogradation of the centre of weight within the orogenic belt during orogenic loading (e.g. stages E and G). The depocentre of the Karoo basin is alternately the foredeep, during orogenic loading, and the foresag, during orogenic unloading. The basin is overfilled, with fully nonmarine sedimentation during stages D–H. As a function of the relative position between the equilibrium drainage profile and the flexural profile, areas of sedimentation, bypass or erosion are separated within the basin. Stage D: the distal sedimentation of the Normandien Formation continues while an unconformity develops in the proximal region at the top of the Balfour Formation (quiescence Q4 in Fig. 12). Stage E: basin-wide sedimentation of the Tarkastad Subgroup (pulse P5 in Fig. 12). Stage F: deposition of a distal sequence subsequently eroded during the forebulge uplift (quiescence Q5 in Fig. 12). Stage G: deposition of a proximal sequence subsequently eroded during the foredeep uplift (pulse P6 in Fig. 12). Stage H: foresag deposition of the Stormberg Group, coeval with the erosion of the older foredeep sequences (quiescence stages Q6, Q7 and Q8 in Fig. 12).

unconformities separating the three coarsening-upward second-order sequences of the Stormberg succession (Fig. 12).

As the orogenic quiescence (stage Q8) continued after the deposition of the Clarens Formation, the flood basalts of the Drakensberg Group filled up the most depressed zone of the contemporaneous topographic profile, which was the subsiding foresag. This explains the occurrence and great thickness (up to 1.8 km) of the Drakensberg Group in the distal region of the Karoo foreland system.

Migration trends of the Karoo foreland system

The pattern of migration of the stratigraphic hinge line, which separates contrasting proximal and distal stratigraphies, is illustrated in Figs 11 and 12. The hinge line shifted towards the craton during the Late Carboniferous–Permian interval, and back towards the orogen during the Triassic–Middle Jurassic. As contrasting stratigraphies develop in response to out of phase base-level changes between the proximal and distal regions of the foreland system, the pattern of hinge line migration reflects a corresponding migration of the flexural foredeep and forebulge. Flexural and stratigraphic responses within foreland systems have been the object of numerous studies (e.g. Beaumont, 1981; Jordan, 1981; Stockmal & Beaumont, 1987; Jordan & Flemings, 1991; Sinclair *et al.*, 1991; Sinclair & Allen, 1992; Waschbusch & Royden, 1992; Watts, 1992; Beaumont *et al.*, 1993; Crampton & Allen, 1995; DeCelles & Giles, 1996), which indicate that migration of the forebulge and foreland basin stratigraphy are sensitive to a number of parameters including thrust front advance rate, orogenic wedge geometry and surface processes.

For the particular case of the Karoo Basin, our data suggest that the cratonward migration of the foreland system (Late Carboniferous–Permian) occurred during a time dominated by orogenic loading in the CFB (pulses P1–P4, Fig. 12). Although the first-order orogenic loading lasted until the end of the Middle Triassic, the foreland system recorded a high-rate shift towards the orogen during the Early Middle Triassic, which was continued with a low-rate shift in the same direction during the Late Triassic–Middle Jurassic stage of first-order orogenic unloading (Fig. 12). Figures 13 and 14 attempt to explain these migration trends using as a main controlling factor the redistribution of orogenic load within the CFB. The progradation of the foreland system (Fig. 13) is attributed to the progradation of the orogenic front during stages of orogenic loading. For the Late Carboniferous–Permian interval, when the progradation of the Karoo foreland system took place, evidence for the gradual advance of the CFB against the basin is supplied by the Dwyka, Ecca and lower Beaufort strata that are incorporated within the orogenic structures in the proximity of the orogenic front. The geology of the CFB indicates that no further progradation of the orogenic front occurred after the end of the Permian, although the first-order orogenic loading continued until the end of the Middle Triassic. We therefore interpret the Early Middle Triassic orogenic loading as a result of piggyback thrusting during pulses P5 and P6 (Figs 12 and 14E,G), which led to the retrogradation of the centre of weight in the orogenic belt and consequently to a high-rate retrogradation of the foreland system. The subsequent low-rate retrogradation of the foreland system during the Late Triassic–Middle Jurassic stage of first-order orogenic unloading is attributed to the erosion of the orogenic front leading to a slow retrogradation of the centre of weight in the orogenic belt (Fig. 14H). Such stages of low-rate retrogradation of the foreland system due to the erosion of the orogenic front probably took place during all quiescence stages of the CFB first-order cycle (e.g. Figs 13B and 14D,F).

It is not clear whether the basement tectonics exerted any control on the position of the hinge line at any of the analysed stratigraphic levels. We show in Fig. 11 the hinge line position at di erent time-slices together with the migration trends, as well as the alignments of the most important sutures in the Precambrian basement that underlies the Karoo Basin. At first sight, no immediate relationship can be established between the basement tectonics and the hinge line position, as the tectonic and stratigraphic trends intersect each other and follow di erent patterns. A correlation might exist between the overall foredeep–forebulge limit in the region west of Aliwal North and the demarcation line that separates the Kaapvaal Craton from the Namaqua–Natal Belt (Fig. 11), although no apparent influence of the basement tectonics on the patterns of hinge line migration can be established. A similar situation has been encountered in the Western Interior Basin where the foreland basin hinge line follows the contact between two Archean blocks of the underlying basement (Catuneanu *et al.*, 1997b).

Comparison with the North American Western Interior Foreland Basin

The analysis of the Late Carboniferous–Middle Jurassic south-Gondwanian foreland basin, part of which the Karoo Basin currently covers, to an area of about 600 000 km^2 within southern Africa, reveals important similarities with the Western Interior Foreland Basin of North America. Excepting for the fact that the south-Gondwanian basin is older and was fragmented as a result of Gondwana break-up, the similarity between the actual 3-D architecture of the two basins (along-dip and along-strike distances, thickness of preserved sequences) is worthy of note. For comparison, both basins extended over more than 6000 km along strike in front of their adjacent orogens (Leckie & Smith, 1992; fig. 5 in de Wit & Ransome, 1992), and host sedimentary successions with thicknesses exceeding 6 km in the foredeep (Cole, 1992; Leckie & Smith, 1992; Rubidge, 1995). In addition, the width of the preserved sedimentary fill is also similar, reaching 865 km on the Canadian side of the Western

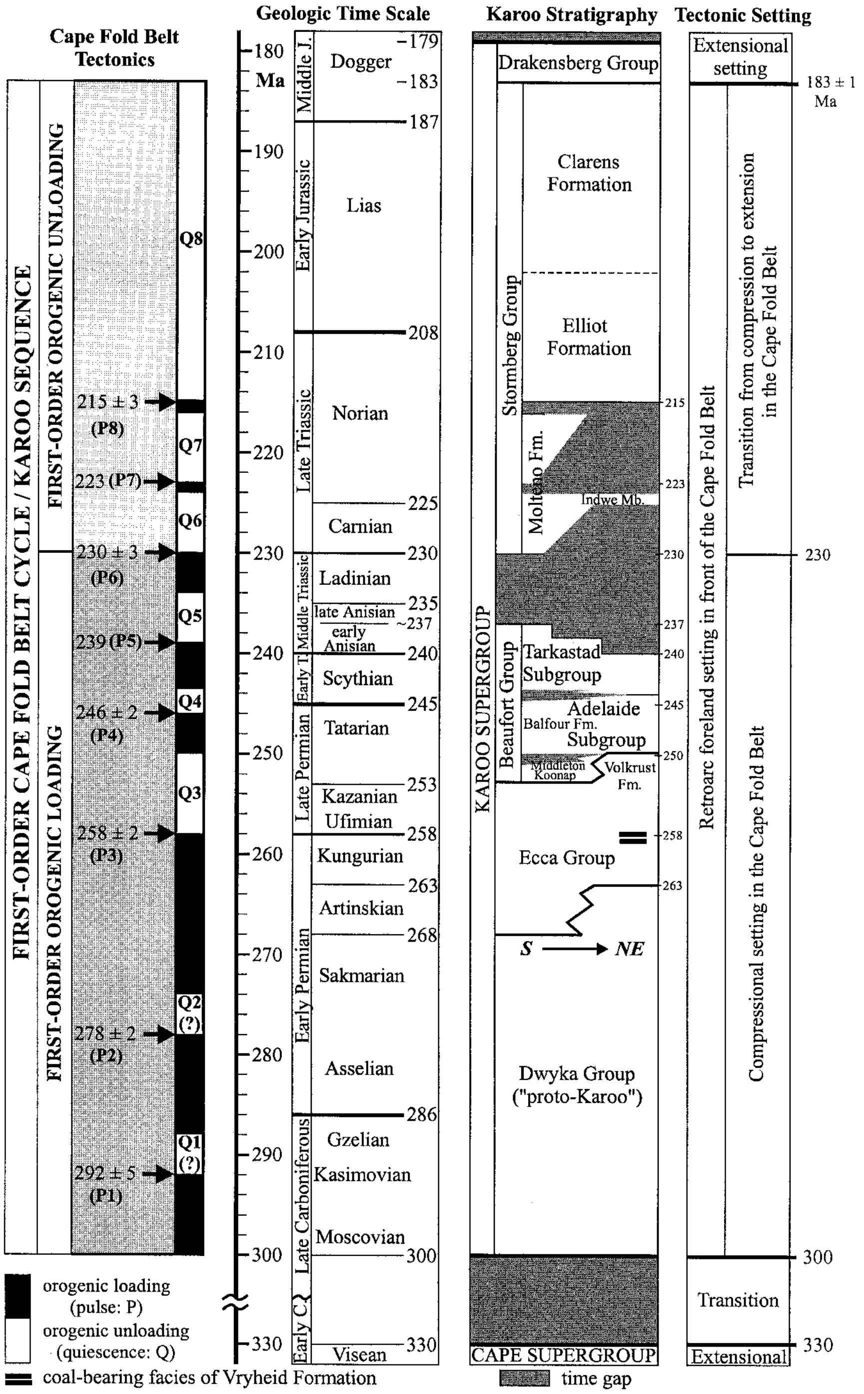

Fig. 15. Tectonic regimes during the evolution of the Karoo foreland basin.

Interior and about 850 km in the Karoo Basin. These figures indicate comparable areas covered by the two basins, and comparable volumes of sediments hosted by them. In turn, this suggests a strong parallel between their geodynamic evolutions, supracrustal loads, flexural profiles, rates of generation of accommodation space and sediment influx. Probably the most important consequence is the fact that within the Karoo, like in the Western Interior, sediment accumulation and preservation took place in both proximal and distal sectors of the basin, the distal stratigraphy being represented by stacked sedimentary sequences rather than a forebulge unconformity.

The pattern of stratigraphic hinge line migration for consecutive time-slices is also similar between the two basins, with initial cratonward and subsequent orogenward shifts. The maximum distance reached by the Western Interior hinge line relative to the orogenic front is 350 km, and 415 km in the case of the Karoo Basin. The influence of basement tectonics on the foreland basin flexural profile and hinge line position is di cult to assess. However, a suggestion is made here that as in the case of the Western Interior Basin, where the hinge line is generally superimposed on the limit between two Archean provinces of the Precambrian basement (Catuneanu *et al.*, 1997b), the Karoo hinge line west of

Aliwal North roughly follows the southern limit of the Kaapvaal Craton of the Precambrian basement (Fig. 11).

CONCLUSIONS

1 In contrast to previous opinions considering the Karoo as a unitary subsiding basin, we suggest that simultaneous base-level rise (subsidence) and base-level fall (uplift) took place within the basin at any given time due to the reciprocal flexural behaviour of the foredeep and forebulge areas in response to stages of orogenic loading and unloading (Fig. 3A). As a result, the depocentre of the foreland system alternated between the depositional foredeep, during orogenic loading, and the depositional foresag during orogenic unloading (Fig. 3B).

2 The Karoo sedimentary fill is identified as a first-order sequence (Fig. 15), with its internal stratigraphic architecture closely controlled by the orogenic tectonism manifested within the CFB. Eight orogenic cycles of loading and unloading in the CFB have corresponding second-order Karoo sequences (Fig. 12). The timing of the second-order sequence boundaries correlate to the end of orogenic unloading stages in the proximal region of the basin, and to the end of orogenic loading stages in the distal region.

3 The first-order CFB orogenic cycle includes a Late Carboniferous–Middle Triassic first-order orogenic loading stage followed by a Late Triassic–Middle Jurassic first-order orogenic unloading stage (Fig. 15). The first-order loading stage is characterized by the accumulation of thick foredeep sequences with much thinner forebulge correlatives (i.e. the Dwyka, Ecca and Beaufort groups), whereas the first-order unloading stage is defined by the accumulation of foresag deposits (i.e. the Stormberg Group) (Fig. 12).

4 The progradation of coarse clastic facies into the foreland system is alternately related to stages of orogenic tectonism, for the proximal region (depositional foredeep), and to stages of orogenic quiescence for the distal region (depositional foresag).

5 The Karoo foreland system migrated towards the craton during the Late Carboniferous–Permian, and back towards the orogen during the Triassic–Middle Jurassic. The cratonward migration of the foreland system is controlled by the progradation of the orogenic front during orogenic loading. The orogenward shift is associated with the retrogradation of the centre of weight in the orogenic belt as a result of piggyback thrusting (orogenic loading without the progradation of the orogenic front: Early Middle Triassic) or with the retrogradation of orogenic load through the erosion of the orogenic front (Late Triassic–Middle Jurassic; Table 1).

6 A remarkable similarity has been found between the Karoo Basin of southern Africa (part of the more extensive south-Gondwanian foreland basin) and the Western Interior Foreland Basin of North America, in terms of basin width, thickness of sedimentary fill and distance of the stratigraphic hinge line from the orogenic belt (with an average of 300–350 km in both cases). This indicates similar geodynamic evolutions, lithospheric flexural properties, supracrustal loads and sediment influx.

Table 1. Major stages in the evolution of the Karoo foreland system. (1) Late Carboniferous–Permian: orogenic loading with the progradation of the orogenic front, leading to the progradation of the foreland system. (2) Early Middle Triassic: orogenic loading without the progradation of the orogenic front (piggyback thrusting with the retrogradation of the centre of weight in the orogenic belt), leading to the high-rate retrogradation of the foreland system. (3) Late Triassic–Middle Jurassic: orogenic unloading with the retrogradation of the centre of weight in the orogenic belt due to the erosion of the orogenic front, leading to the low-rate retrogradation of the foreland system.

<table>
<tr><th colspan="2">Orogenic processes</th><th>Migration of the foreland system</th></tr>
<tr><td>Middle Jurassic
(B) First-order
orogenic unloading:
foresag depocenter

Late Triassic</td><td>(3) Retrogradation of the
orogenic load due to the
erosion of the orogenic front</td><td>Low-rate retrogradation</td></tr>
<tr><td rowspan="2">Middle Triassic

(A) First-order
orogenic loading:
foredeep depocenter

Late Carboniferous</td><td>(2) Loading without the
progradation of the orogenic
front (piggyback thrusting):
retrogradation of the centre
of weight in the orogenic belt
Triassic</td><td>High-rate retrogradation</td></tr>
<tr><td>Permian
(1) Loading with the
progradation of the orogenic
front</td><td>Progradation</td></tr>
</table>

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