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PALAEOGEOGRAPHIC IMPLICATIONS OF BRAID BAR DEPOSITION IN THE TRIASSIC MOLTENO FORMATION OF THE EASTERN KAROO BASIN, SOUTH AFRICA

by

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ABSTRACT

The Triassic Molteno Formation in the main Karoo Basin, South Africa, forms a northerly thinning intracratonic clastic wedge deposited by sandy braided rivers of South Saskatchewan type. Deposition of the sandy facies was dominated by channel floor megaripples producing trough cross-bedded cosets; transverse bars, represented by solitary, large-scale planar sets are not significant.

Departures from this regional pattern of sandstone deposition occur along the northern distal margin of the Molteno basin around Bethlehem in the Orange Free State. Here thickness trends and clast size delineate a deep channel system interpreted as the main braided exit channel from the basin. Because of its depth and constriction by local height differentials the competency and capacity of the flow were able to reproduce features more typical of proximal rather than distal depositional settings. The sandy facies is dominated by fine gravel with lesser amounts of coarse sand. Gravel occurs as longitudinal bars some of which contain low angle foreset stratification whose orientation is consistent with lateral growth and marginal riffle migration. The scale of the bars and simple depositional form imply that they may have been larger than modern equivalents and the flows deeper.

The coarse sand occurs mainly as falling water stage features associated with the gravel bars. Shallow channel-fills, bar edge sand wedges, bar top sheet sands and thicker channel sands have been recognised and compared with similar features in modern and ancient braided stream sediments. When traced to the southeast the deep channel sediments contain few longitudinal gravel bars and more transverse bars; the vertical sequence from longitudinal to transverse bars at this locality points to the increasing distality of the depositional site through time.

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INTRODUCTION

The Molteno Formation in the main Karoo Basin, South Africa (Fig. 1) forms a northerly thinning intracratonic fluvial clastic wedge covering an area of about 25 000 km². The age of the formation is uncertain. It is regarded as Upper Triassic (Carnian) by Anderson and Anderson (1970) and Middle Triassic by Plumstead (1969) and Keyser (1973). Regional facies analysis shows the formation to consist of a number of stacked large-scale fining-upward sequences comprising conglomerate, pebbly sandstones, fine sandstone and siltstone, shale and coal. These sequences are thought to have been deposited by braided streams draining an alluvial plain which may have been built on to the distal slopes of alluvial fan complexes of glacial outwash type. Basinward the sediments intertongue with floodplain-lacustrine sediments of the Elliot Formation (Upper Triassic) and Beaufort Group (Permo-Triassic). Full details of the regional depositional model are given by Turner (1983).

The pebbly sandstones, which dominate the sequence are multilateral and multistoried sheet sandstones from 30–120m thick. Internal sedimentary structures consist predominantly of large-scale trough cross-bedding (average thickness for the entire basin 56 cm), with subordinate large-scale solitary planar sets (up to 1,8 m thick) and some flat-bedding. Palaeocurrent trends based on trough foresets indicate a source area to the south and southeast of the present eroded edge of the formation. Petrographic studies suggest that most of the sediment was derived from metaquartzites of the Cape Supergroup to the south and a granitic fault-block terrain probably located off the present southeast

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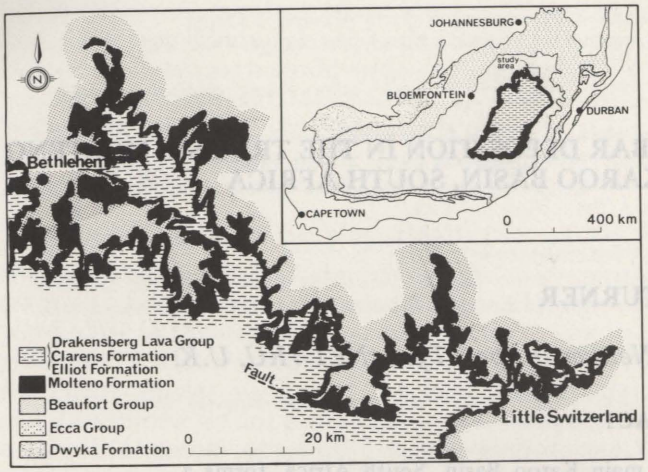


Fig. 1 Generalised stratigraphic and locality map of the study area The inset map shows the Karoo Basin with study area enclosed.

coastline of South Africa (Turner, 1983). Proximal facies equivalents are not preserved because of repeated uplifts and peneplanation of the basin margin from Late Jurassic times onwards (Du Toit, 1954).

The pattern of sedimentation shows a striking resemblance to the South Saskatchewan braided river model of Cant and Walker (1978). Sediments were deposited by perennial, high energy, low sinuosity braided streams. The relative abundance of trough cosets to solitary planar sets implies that sedimentation was dominated by undirectional migration of channel floor megaripples and that channel bars were of little significance. The finer grained sediments were deposited largely from waning bedload to suspension load sedimentation in response to channel shifting, abandonment and overbank flooding. Shales record deposition from vertical accretion following overbank flooding, with standing bodies of water on the alluvial plain providing the locus for plant growth and *in situ* accumulation of coal.

Well rounded to subangular pebbles, cobbles and boulders composed predominantly of quartzite are present, especially in the middle of the formation where they form a thin, but laterally persistent unit at the base of the Indwe Sandstone Member (Turner, 1975). This is the only regionally extensive lithostratigraphic unit in the formation.

The pattern of sandstone deposition outlined above remains the same throughout the Molteno basin except along the northern distal margin of the basin (Fig. 1) where the sandstone (Indwe Sandstone Member) is locally much thicker and longitudinal bars become less common and sandy transverse bars are important depositional features. These bedforms are generally lacking elsewhere in the basin and represent an important local change in depositional conditions requiring modification of the existing regional depositional model. The purpose of the present paper therefore is to describe and interpret the localised change in sedimentation along this northern margin of the basin within the framework of the established regional model.

SEDIMENTOLOGY

When traced basinwards the regionally extensive Indwe Sandstone Member of the formation thins out from about 60 m to a few metres except locally along the northern distal margin of the outcrop in the vicinity of Bethlehem in the Orange Free State, where a pronounced thickening and increase in clast size record the position of a deep channel system (Fig. 2). Regional clast size and thickness trends argue against a local sediment source area such as the structural Harrismith dome (Fig. 2), a view substantiated by the fact that the dome was not exposed at this time and apart from some thinning and loss of sequence across the structure it had little influence on sedimentation (Turner, 1975). Limited palaeocurrent data around the dome provides further evidence against the idea of a local intrabasinal sediment source area. Reworking of clasts from underlying coarse lower Beaufort Group sediments is also unlikely because of their more varied lithology, angular nature and much smaller size. In the absence of any recognisable local intrabasinal sediment source an alternative explanation must be sought consistent with regional sedimentary trends. One possible explanation is that the channel system and its sediment fill record the convergence of several small channels into one

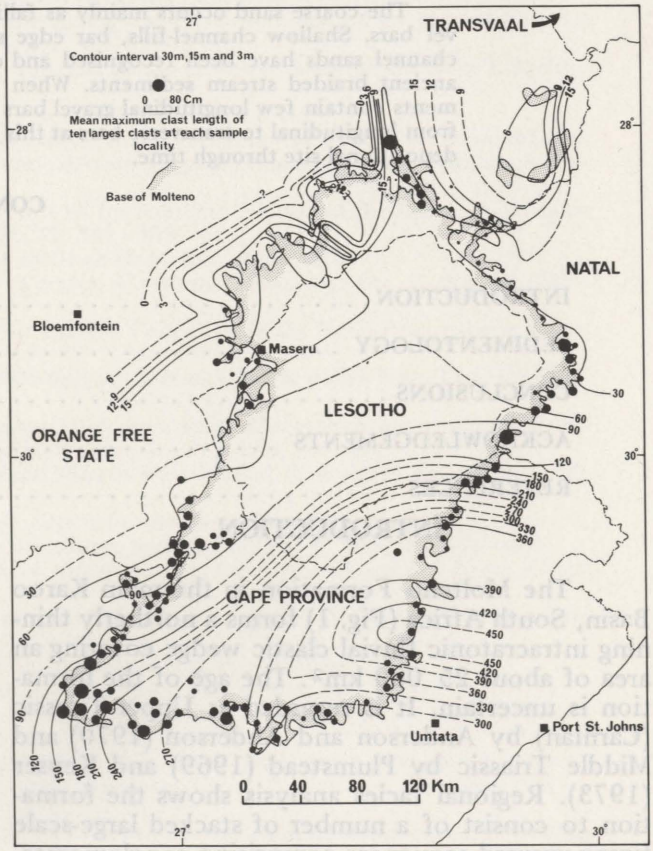


Fig. 2 Clast size and isopach map of total thickness of the Molteno Formation. Note the local increase in clast size and thickness along the northern margin of the outcrop, contrary to regional trends. The dome-shaped feature delineated by the isopach contours in the northeast is known as the Harrismith dome (Turner,1975).

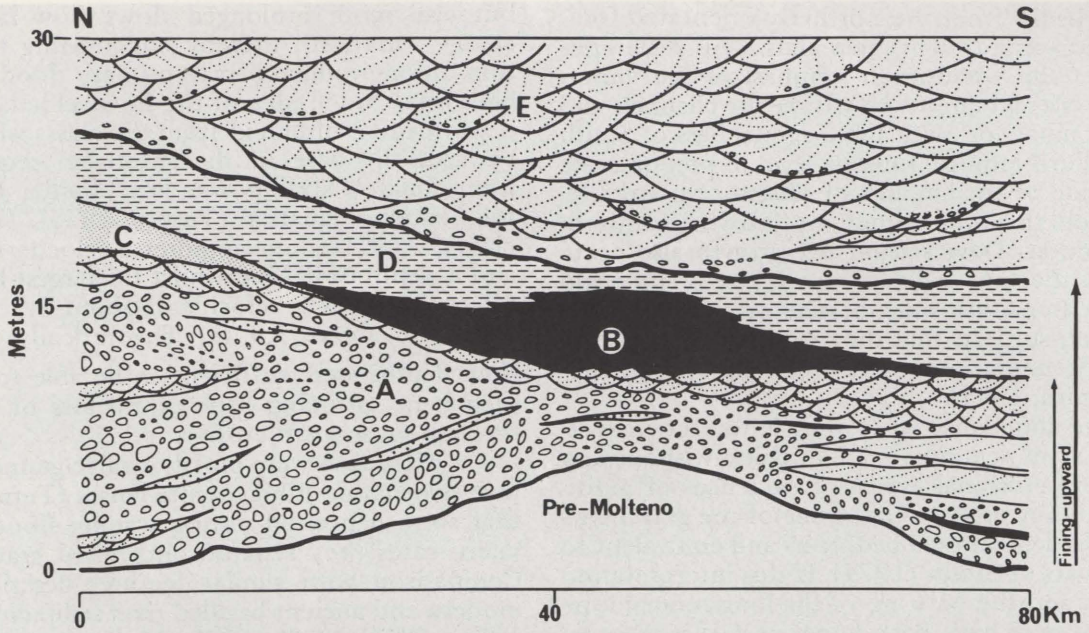


Fig. 3 Generalised section showing characteristic features of the Molteno Formation along the north-central margin of the basin. General transport direction from south to north. (A) The lower part of the formation (Indwe Sandstone Member) consists of pebble and granule conglomerate (gravel) overlain by trough cross-bedded very coarse, pebble and granule rich sandstone. The conglomerate is thicker and generally coarser in the north (distal) becoming thinner and slightly finer to the south (proximal) accompanied by a change in the dominant stratification type from longitudinal gravel bars to more sandy transverse bars and subordinate gravel bars confined to a thin basal zone. The remainder of the succession consists of (B) shale, (C) fine-medium sandstone, (D) silty-shale and (E) coarse granule and pebble rich trough cross-bedded sandstone. The section line extends from Danielsrust to Oliviershoek Pass and corresponds to sections 32, 41 and 42 in Fig. 10.

main braided channel, due perhaps to distal channel entrenchment or constrictions and height differentials composed by local features such as the Harrismith dome (Fig. 2) – a situation analogous to that of the Knik River, Alaska (Bradley et al., 1972). However, the decrease in overall grain size and thickness to the south, and the accompanying changes in the dominant stratification pattern and fining-upward trend (Fig. 3) typical of the remainder of the succession favours channel constriction and convergence rather than localised short-term channel entrenchment. Palaeocurrent data for this northern margin of the basin provides some additional support for this idea of convergence (Fig. 4) and the development of a main braided exit channel at this locality (Fig. 5).

At the same time channel constriction improves its competence and transport capacity (Boothroyd and Ashley, 1975) reproducing features that might be expected in more proximal situations. Within this environment gravel and sand were deposited contemporaneously. Matrix poor, small pebble and granule gravel is very common and dominates the lower part of the sequence (Fig. 3). It accumulated as sheet-like longitudinal gravel bars elongate parallel to the transport direction. Individual bars are up to 1,7 m thick, 10 m wide and several tens of metres in length. Internally the bars are massive or show low angle foresets which are commonly convex upwards and contain rare sandstone clasts (Fig. 6). Foreset dip directions are orientated towards the west and northwest, and de-

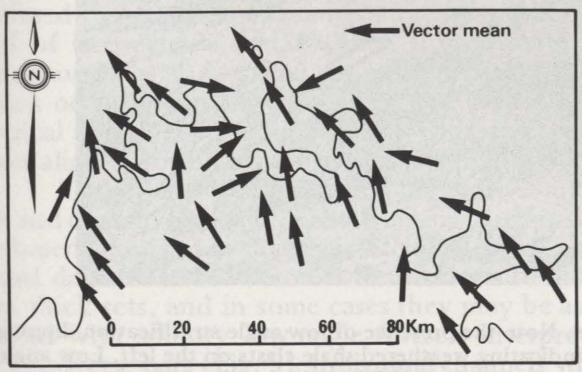


Fig. 4 Palaeocurrent map of the northern distal margin of the Molteno Basin. Vector means based on 1018 measurements of the axes of trough cross-bedded sets.

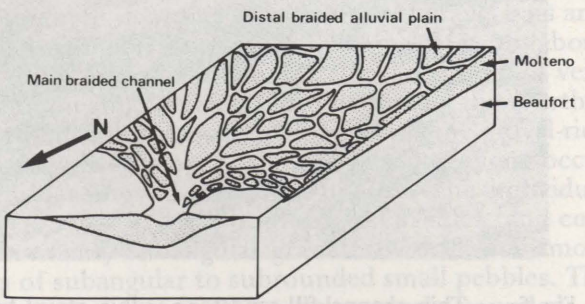


Fig. 5 Generalised depositional model for the Molteno Formation along the north-central distal margin of the basin.

viate markedly from the northerly orientated foresets of adjacent trough cross-beds. This difference in foreset dip orientation emphasizes the importance of lateral bar growth within the channels and invites comparison with the diagonal bars of Smith (1978). Furthermore, foresets tend to predominate on one side of the bar and are absent or poorly developed on the other, implying growth mainly in one direction. Downstream bar growth also occurred, but the low angle of the foresets and rapid upstream transition into massive or crudely stratified gravel, suggests that they were formed by marginal riffle migration rather than slip face avalanching (Smith, 1974).

The abundance of low angle foreset cross-stratification in sets up to 1,7 m thick without any intervening erosional events or evidences of multiple deposition suggests that some of the gravel bars are simple depositional bedforms and equivalent to the unit bars of Smith (1974). If this interpretation is correct and the bars are of the longitudinal type then they may have been larger and of greater relief than in modern systems and the flows responsible for their formation deeper (Rust, 1978), particularly where humid climates prevail and rainfall was high as in the Molteno (Turner, 1983). Such conditions would not only promote deeper flows but also more prolonged flows (low rate of discharge variation), thereby encouraging the formation of longitudinal bars as primary flood generated bedforms. Some idea of the flow velocity and flow depth can be obtained from the clasts which occur towards the base of the gravel bar sequence, by using Malde's sixth-power law (Malde, 1968) and the Manning equation (Turner, 1975). The average clast size of 16,5 cm requires a velocity of 2,0 m/s and a depth of 0.73 m whilst the largest boulder of 42 cm requires a velocity of about 3,3 m/s and a depth of about 1,8 m. Theoretically therefore, flow depths were sufficient to be able to generate such bars, assuming little or no loss of bar relief through subsequent erosion.

The sands were mostly coarse-grained, immature feldspathic lithic arenite lenses (Turner, 1975) that were deposited during waning floods, and in many cases they separate individual gravel layers. Comparison with similar features described from modern and ancient braided river sediments (Smith, 1971, 1974; Miall, 1977; McGowen and Groat, 1975; Boothroyd and Ashley, 1975) suggests that the arenites occur: (1) within the gravel sequence as channel-fills (Fig. 6); (2) as bar edge and slip face sand wedges (Fig. 7); (3) as bar top sheet sands (Fig. 8); and (4) as thicker channel sands, late-

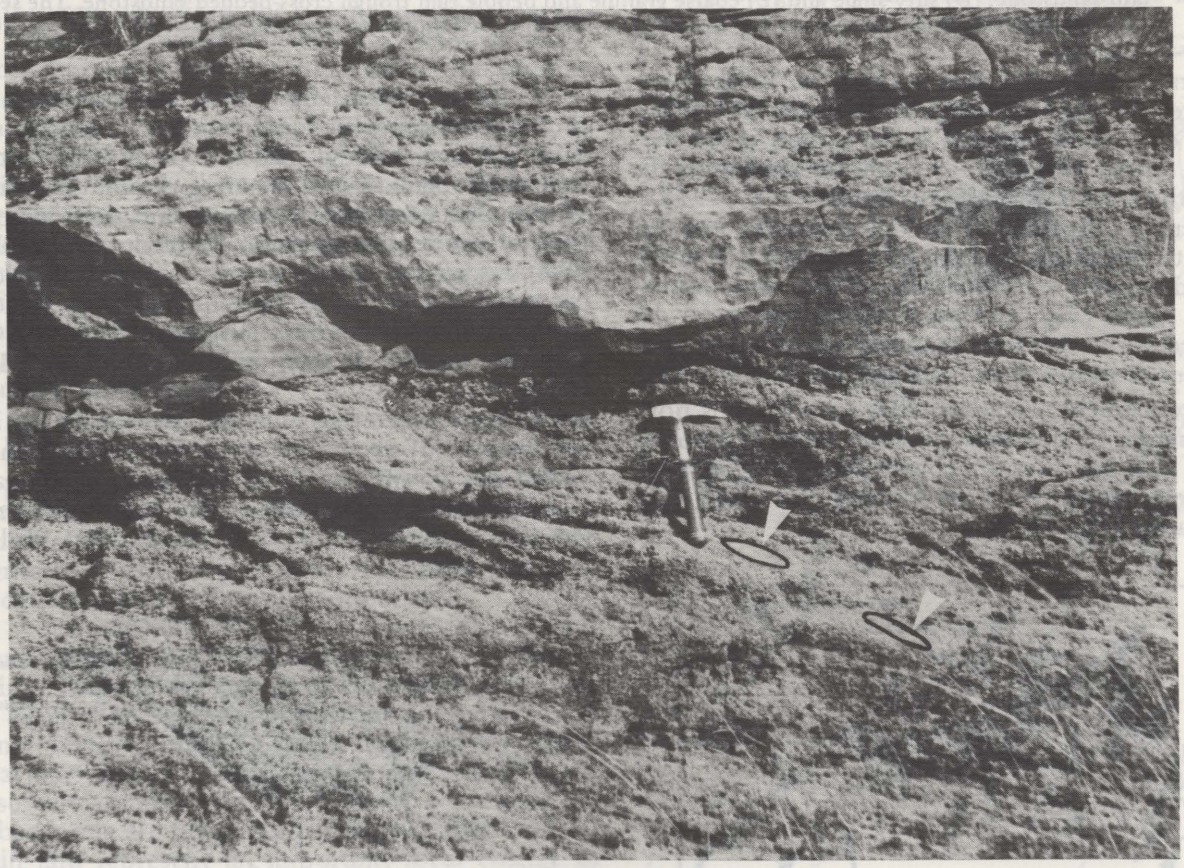


Fig. 6 Thin channel-fill sandstone within gravel bar sequence. Note the presence of low angle stratification dipping into the channel on the right, and small pock marks indicating weathered shale clasts on the left. Low angle gravel bar foresets dipping to the right can be seen beneath the channel-fill sandstone. Some foresets contain small, discoidal sandstone intraclasts (outlined in black and arrowed in white) with their long axes preferentially aligned down the foreset slope. General transport direction towards the observer.

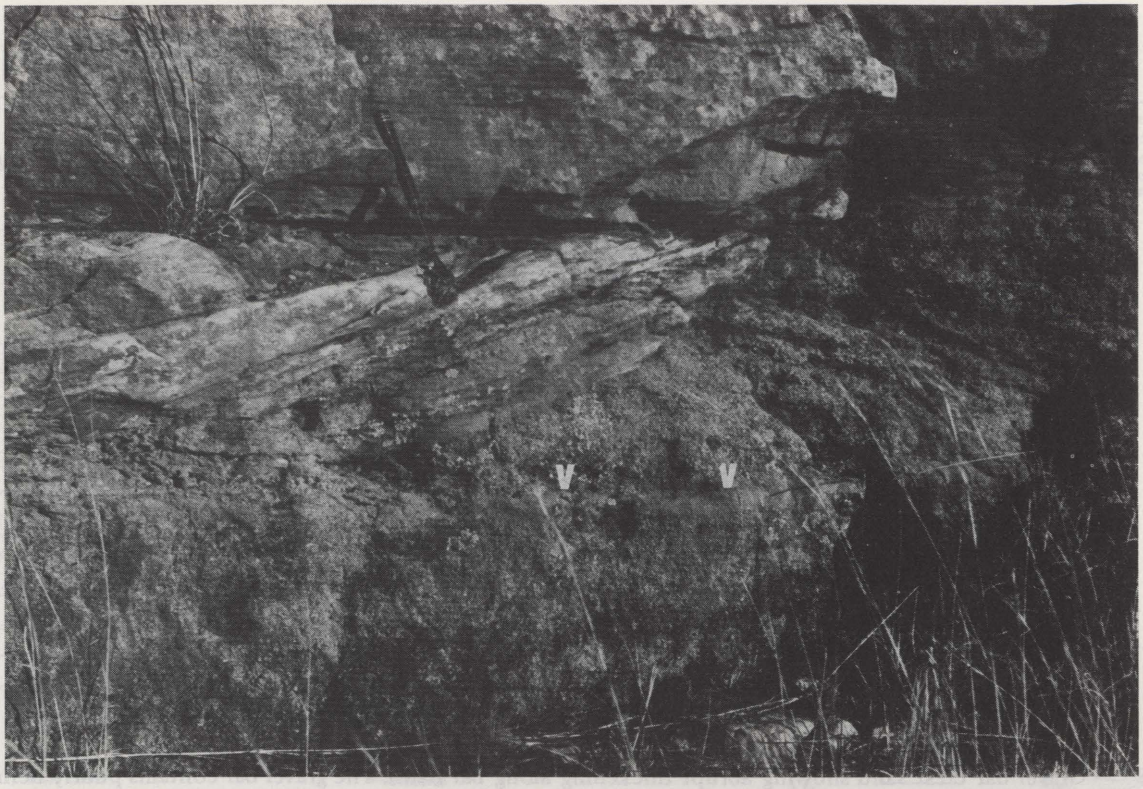


Fig. 7 Bar edge sand wedge with low angle foresets. A thin bar top sand occurs below the wedge (arrowed) which is overlain by a thicker, erosively based sandstone representing part of a sand-filled channel cut into the gravel bar top. General transport direction towards observer.

ral to and interbedded with the gravel. The shallow channel-fills are up to 2,5 m wide and 0,5 m deep, and locally contain shale intraclasts. Internally they are massive or contain traces of low angle stratification dipping into the channel indicative of side-filling (Fig. 6). Trough cross-bedding, as recorded for many modern gravel bar surface channels (Boothroyd and Ashley, 1975) is not developed. The channel-fills record deposition within small channels dissecting the bar surface during falling water stage, under conditions which favoured the rapid dumping of poorly sorted sediment and inhibited the development and preservation of bedforms.

Bar crest sands are generally less than 45 cm thick, but they may thicken abruptly along the bar flank as shown in Fig. 8. Where individual bars are defined by the bar crest sands, they have a bed relief of between 25 and 170 cm. Intraformational sandstone clasts up to 5-6 cm in length and 2-3 cm thick occur on gravel bar foresets. The clasts have a typical flattened discoidal shape with their long axes aligned down the foreset slope (Fig. 6).

The thicker channel sands are up to 2 m thick and contain trough cross-beds arranged in erosively-based cosets. Locally they show an abrupt upward decrease in scale from 1 m thick sets to 20-30 cm thick sets, and in some cases they may be associated with solitary low angle foresets interpreted as transverse bars (Fig. 9). Individual channels within these thicker sands are difficult to detect probably because of rapid channel shifting, and the unconsolidated nature of the sediments. Adjacent to the main braided channel, coarse to medium grained trough cross-bedded sandstone was deposited (Fig. 10). Individual sets are up to 35 cm thick, and the sandstones are noticeably lacking in pebble and granule size clasts in comparison with the main braided channel sediments. The environment was essentially one of relatively minor shallow braided streams, where depths of flow and scour were reduced, and sand rather than gravel was deposited. However, discrete interconnected channels are not preserved presumably on account of channel shifting, the poorly consolidated nature of the sediments and reworking of any interchannel deposits.

When traced some 80 km to the southeast, to Little Switzerland along the edge of the Drakensberg escarpment (Fig. 1), the deep channel sediments change character. Gravel is less common and mainly confined to the lower 1,5 m of the succession where it occurs as longitudinal gravel bars and small granule gravel-filled channels up to about m thick. Clasts of pebble size or larger are not very common and overall tend to be smaller than further north in the palaeoflow direction. Fine gravel-rich layers alternating with gravel-poor layers occur within some sandstones (Fig. 11). The individual layers range from about 3 to 15 cm thick, and contain mainly subangular granules with minor amounts of subangular to subrounded small pebbles. The layers do not appear to be erosively-bounded, or form part of discrete small-scale fining-upward sequences In view of this the grain size variation may represent a response to discharge fluctuations and incremental bar growth, with each coarse and fine

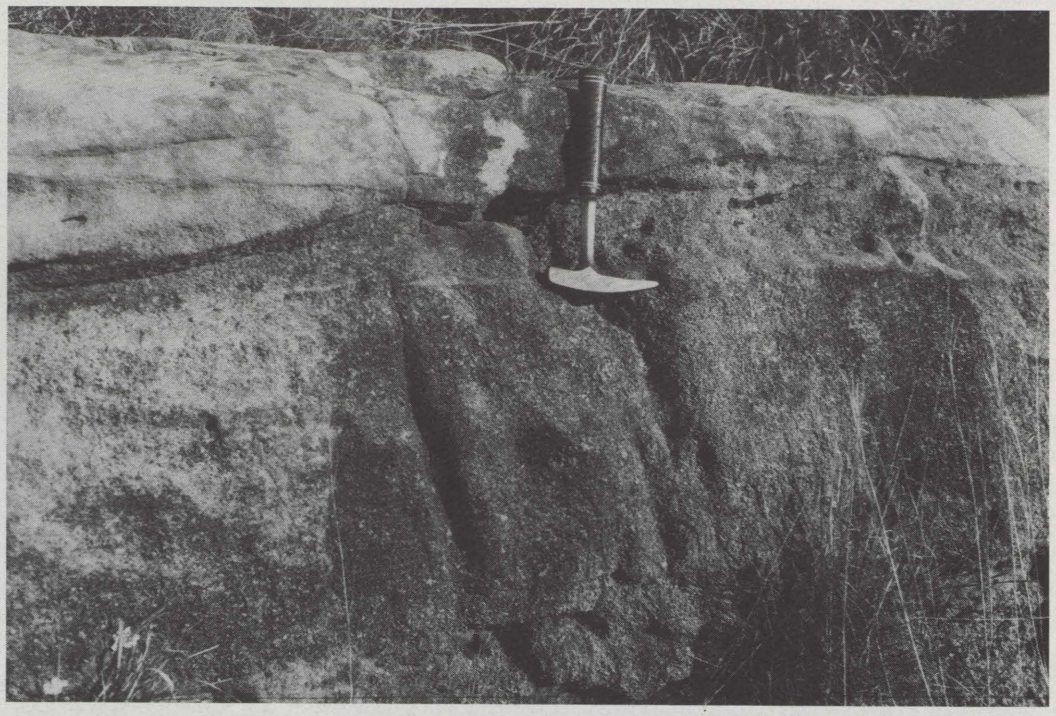


Fig. 8 Gravel bar crest sand showing abrupt thickening along bar flank. The gravel bar contains poorly defined low angle foresets dipping from the bar crest towards the flank. Other much thinner bar top sands occur at the level of the hammer head. General transport direction towards observer.

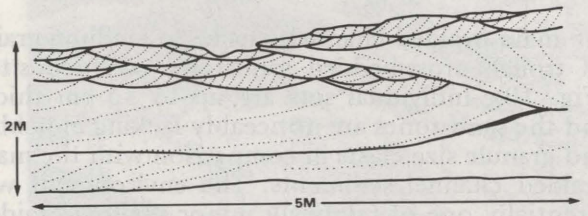


Fig. 9 Stratification sequence in braided channel-fill sandstone. Trough cross-bedding (sinuous crested dunes) dominates and shows an abrupt upward decrease in scale. A foreset cross-bedded unit of transverse bar origin occurs at the bottom left. Drawn from a photograph.

layer representing a depositional couplet recording high and low discharge. Such alternations are particularly well developed where deposition is dominated by lateral accretion of low slope surfaces such as riffles (Smith, 1974).

Gravel bars and channel-fills are overlain by solitary sets of large-scale planar cross-bedding interpreted to be coarse sandy transverse bars. These now become important depositional features in this area with individual bars up to 80 cm thick and 6 m in length, the foresets are generally straight, tabular and strongly discordant with the lower bounding surface. Internally the bars show the following features; (1) deformed foresets (Fig. 12); (2) curved convex upper surface tapering off upcurrent (Fig. 13); and (3) finer grained small trough cross-bedded bar top sands (Fig. 13), a feature that appears to be lacking from transverse bars elsewhere in the basin possibly due to the rapid rate of discharge fall and disequilibrium between the flow and the bedform. Intervening between the gravel bar and overlying transverse bar shown in Fig. 12 is a thin wedge up to 20 cm thick of better sorted, quartz-rich arenite, interpreted as a bar top sand. This could represent reworking of the bar top during falling water stage, or possibly subaerial exposure of the bar surface which was then subjected to wind action, winnowing out the fines and leaving behind a coarse structureless aeolian sand. Transverse bar foreset directions often diverge strongly from the northerly regional palaeocurrent trend based on trough axes. Another feature is the presence within the gravel sequence of an erosively-bounded, black, carbonaceous, finely laminated shale lens up to about 1,2 m thick, floored by a 15-20 cm thick matrix supported conglomerate containing large quartzite pebbles and cobbles with an average clast size of 6 cm. The erosive floor and conglomerate, and the abrupt lithological change to shale, can best be explained in terms of an abandoned braided channel-fill. The presence within the shales of well preserved delicate fossil plant material assigned to the Triassic *Dicroidium* flora supports this view and emphasises the very quiet local conditions under which sedimentation occurred, remote from any influence of the active braided channels. The observed vertical relationships between bar types suggests increasing distality of the braided system through time (Smith 1971), an interpretation consistent with its stratigraphic position beneath fluvio-lacustrine sediments (Turner 1983). A generalised interpretative depositional model for the deep channel sediments in the lower part of the Molteno Formation is shown in Fig. 14.

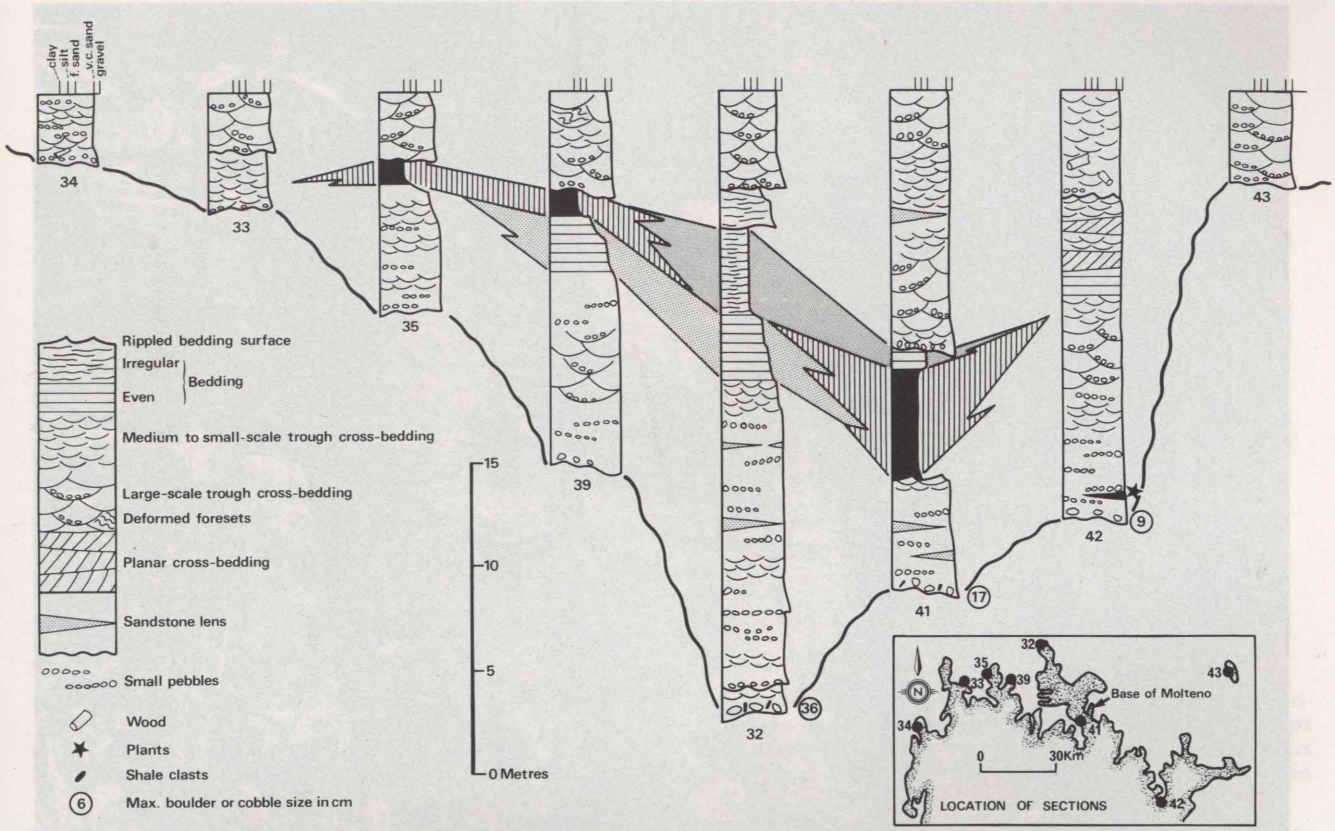


Fig. 10 Correlation of sections along the northern margin of the Molteno Basin showing lateral variations in thickness and stratification.

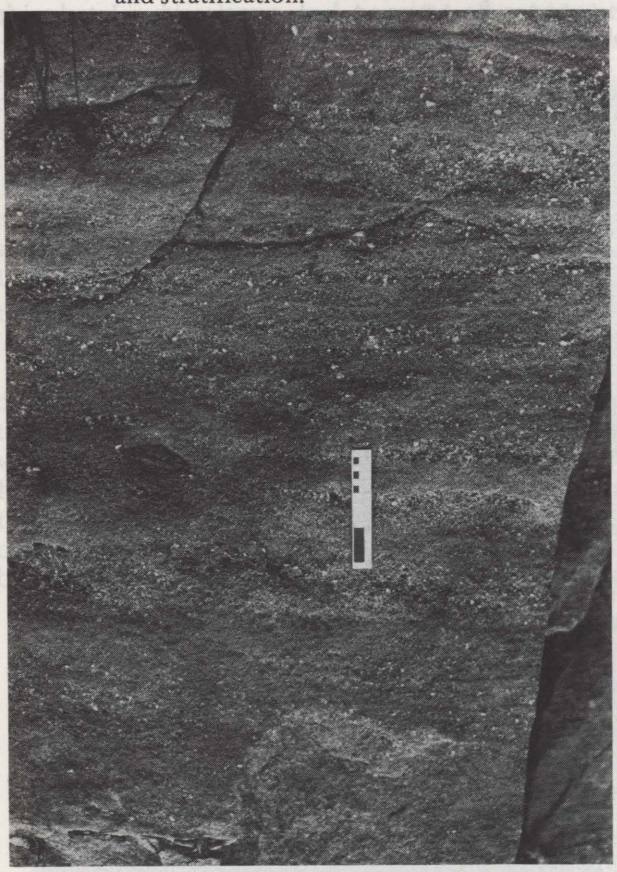


Fig. 11 Internal structure of longitudinal bar showing systematic variation in grain size, and depositional couplets of gravel-rich and gravel-poor very coarse sandstone.

CONCLUSIONS

The Triassic Molteno Formation in the main Karoo Basin, South Africa, forms a northerly thinning intracratonic clastic wedge deposited by coarse sandy braided streams. Regional facies analysis and the pattern of sandstone deposition suggest that the streams were probably of the South Saskatchewan type and that deposition of the sandy facies was dominated by curved discontinuously-crested channel floor megaripples producing trough cross-bedded cosets. Channel-bars of transverse type, represented by large-scale solitary planar sets are not significant within the sandy facies, which fines up into siltstone, shale and coal deposited mainly from channel sifting, overbank floods and within peat swamps on the alluvial plain during periods of tectonic quiescence (Turner, 1983).

Departures from this regional model occur along the northern distal margin of the Molteno outcrop where the sandy facies shows a pronounced thickening and increase in clast size inconsistent with regional trends and its distal position. In the absence of any local intrabasinal sediment source these marked local differences are interpreted in terms of a deep channel system, thought to represent the main braided exit channel from the basin. Because of its greater depth and channel construction, possibly due to height differentials imposed by features such as the nearby Harrismith dome, competency and capacity of the flow was considerably increased thereby reproducing more proxi-

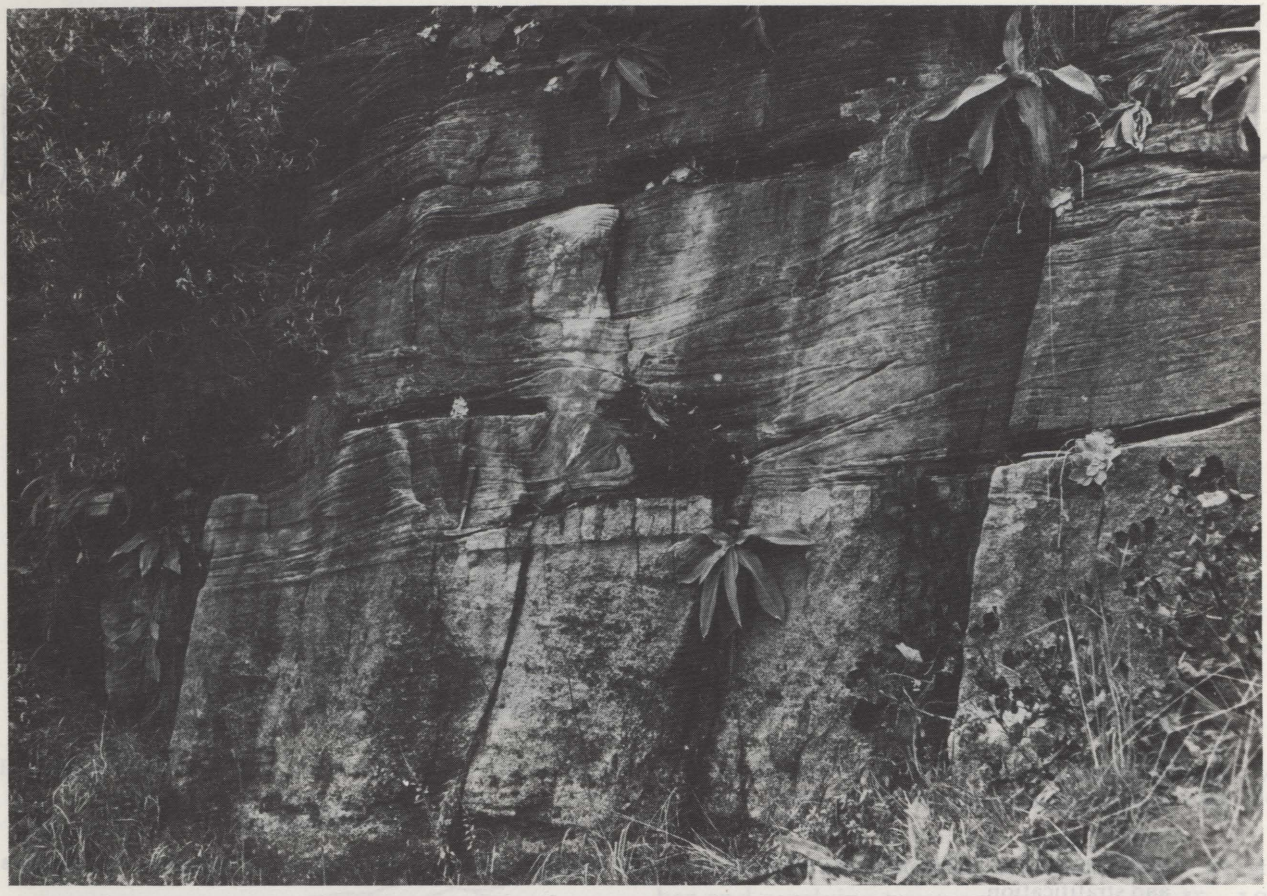


Fig. 12 Longitudinal gravel bar with low angle foresets overlain by locally deformed sandy transverse bar. Intervening between the two is a thin lens of winnowed (wind-blown?) bar top sand.

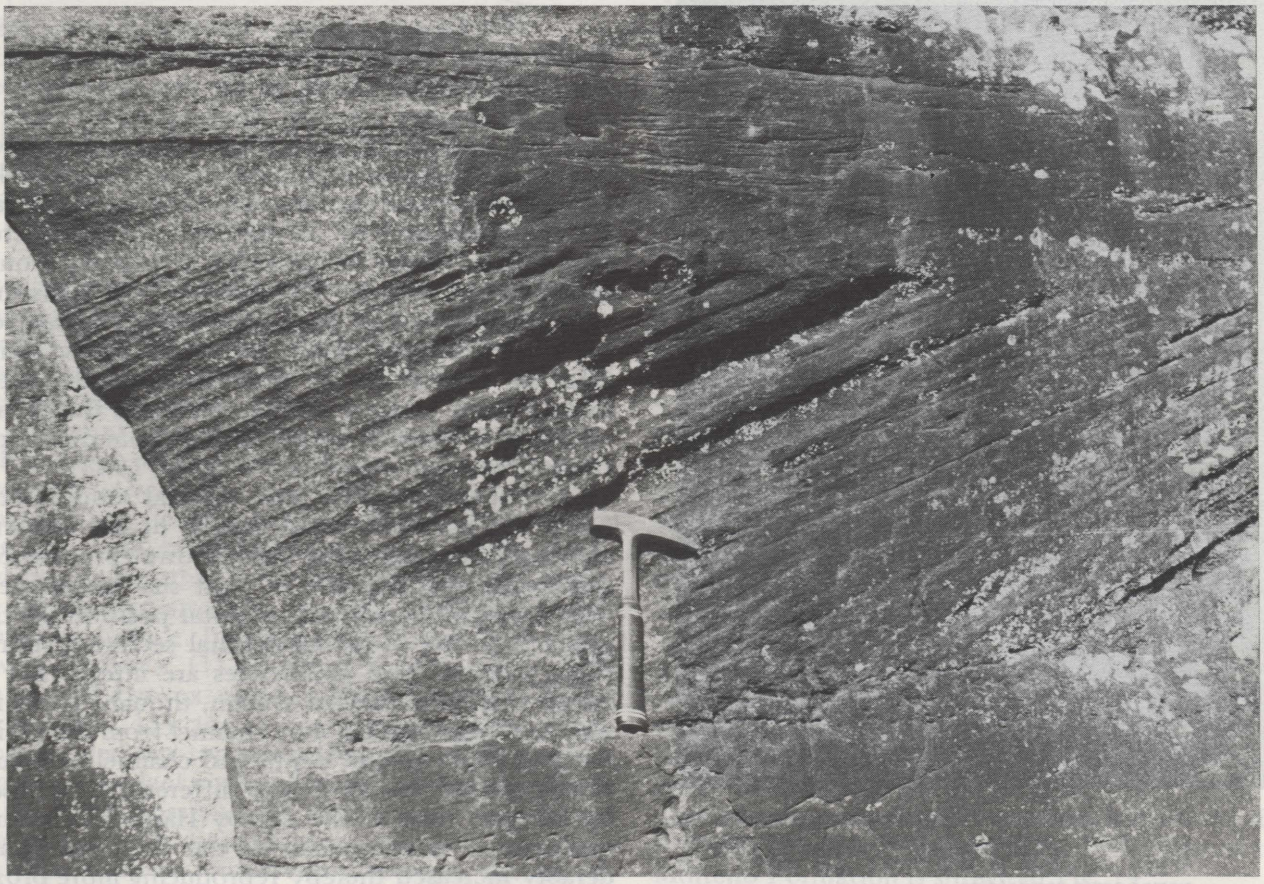


Fig. 13 High angle simple planar set of transverse bar origin overlain by a separate sedimentation unit of small trough cross-lamination. Note the curvature of the bar surface and its tapering-off towards the bar tail to the right.

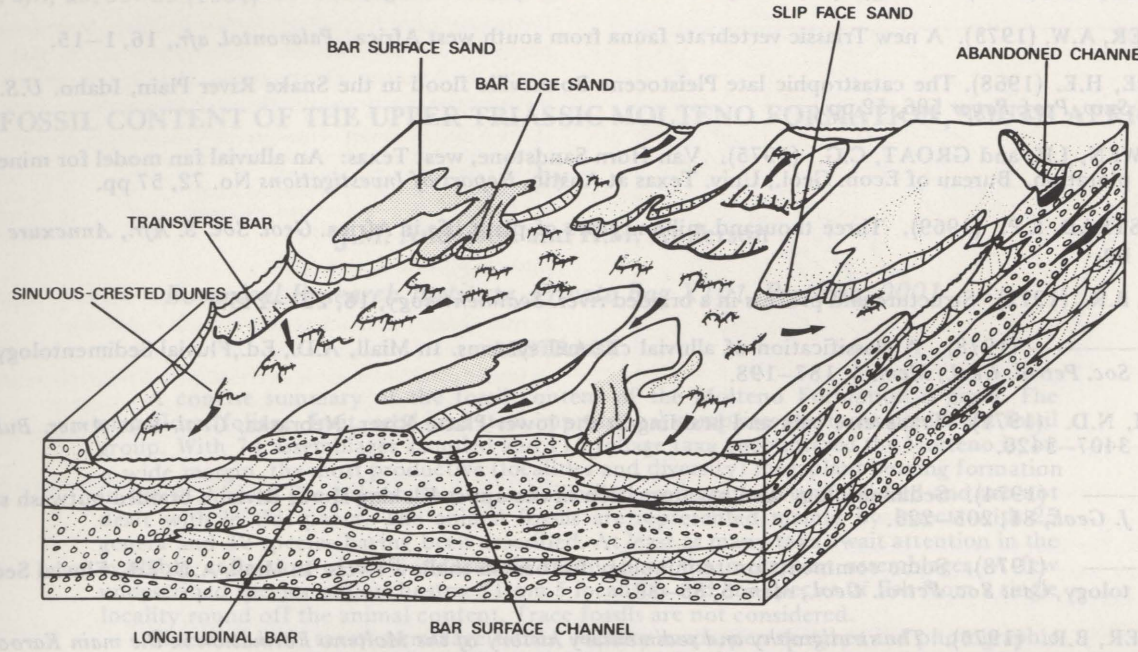


Fig. 14 Block diagram showing generalised interpretive depositional model for the deep channel sediments in the lower part of the Molteno Formation. Coarse stipple indicates gravel and fine stipple sand. Spatial relationships between major bedforms indicates a dominance of longitudinal gravel bars in the north and transverse bars further south. Characteristic sedimentary structures and sequences produced by the model are illustrated in Figs. 6, 7 and 8.

mal characteristics within a distal setting.

The sandy facies is generally coarser than elsewhere in the basin and consists of contemporaneously deposited granule and small pebble size gravel and coarse sand. The gravel was deposited in the form of longitudinal bars characterised by low angle foreset stratification. Foreset dip directions imply lateral growth within the channel, produced largely as a result of marginal riffle migration rather than slip face avalanching. Some of the bars show features consistent with the simple depositional bedforms of Smith (1978) but in terms of their size they must have been larger than equivalent modern bars and the flows deeper.

Comparison with modern and ancient braided sediments suggests that the sandstones associated with the gravels were deposited in a variety of ways during falling flood stage as shallow channel-fill deposits, as gravel bar edge sands, as gravel bar top sheet sands and as thicker internally complex channel sands, lateral to the gravel bars and interbedded with them. When traced some 80 km to the southeast the deep channel sediments change character. Gravel bars are less common and mainly confined to the lowermost part of the succession. Transverse bars become more important bedforms and show a variety of features including deformed foresets, curved, convex-up bar surfaces, tapering off upcurrent towards the bar tail and preserved bar top bedforms. Such bedforms have not been recognised elsewhere in the basin, and the vertical sequence of bar types at this locality points to the increasing distality of the depositional site through time.

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