CHAPTER 4

PALAEOCURRENTS, DEPOSITIONAL ENVIRONMENTS AND BASIN EVOLUTION

The stratigraphy and sedimentology of the Yeneena Group have been described in detail and briefly interpreted in the foregoing chapters, and the main characteristics of each formation are summarised below. Palaeocurrent data are now added to the descriptions, and detailed discussions of the depositional processes and environments of both the lower and upper Yeneena Group are then given. Finally, the tectonic setting of the basin and the changing styles of sedimentation are discussed.

4.1 SUMMARY OF THE YENEENA GROUP SEDIMENTARY FEATURES

4.1.1 LOWER YENEENA GROUP

Coolbro Sandstone - comprises basal conglomerate and an overlying monotonous sandstone sequence. Conglomerate occurs in several lenses, which are thickest at the base of the unit; clasts are mainly of quartzite, vein quartz and lithologies of the Rudall Metamorphic Complex. Sandstone is dominantly moderately well sorted, fine to coarse grained quartzose and micaceous arenite, some is subarkose; medium to very thick bedded; cross bedding common in some areas, mainly solitary tabular sets, some stacked trough sets; parallel lamination also common, and convolute bedding in places.

Broadhurst Formation - dominated by shale, some of which is very carbonaceous. Minor micaceous sandstone, dolomite and iron formation. Sedimentary structures are rare, but grading occurs in very thin siltstone beds, and slumping and ripple marks have also been recorded.

Choorun Formation - comprises cyclic units of pebble and cobble conglomerate, coarse to medium grained sandstone and shale; and non-cyclic sandstone units. Cross bedding, parting lineation and channelling all occur in sandstones, and long axes of pebbles in conglomerates are aligned apparently perpendicular to palaeocurrent directions. Sandstone varies from quartz-rich arenite to subarkose.

4.1.2 UPPER YENEENA GROUP

The sedimentary characteristics of the three facies of the Isdell Formation and the five facies of the Malu and Telfer Formations are best summarised in tabular form (Table 4.1). The two stratigraphically highest formations in the group can be summarised as follows.

Puntapunta Formation - comprises three lithologies - dolomite, limestone and minor carbonate-rich sandstone or siltstone. Dolomite includes dolarenite and dololutite, which are thin to medium bedded; limestone is dominantly calcarenite, which is mostly thick to very thick bedded. Arenaceous beds are massive, parallel laminated, cross laminated or cross bedded; dololutite beds usually parallel laminated; rare intraformational conglomerate. Carbonate-rich sandstone and siltstone are poorly bedded, and contain cross lamination, parallel lamination and parting lineation.

Wilki Quartzite - monotonous sequence of fine to medium grained quartzite, commonly in laterally extensive units about 5 m thick. Most outcrops are apparently massive, but rare cross lamination occurs.

4.2 PALAEOCURRENTS

The importance of palaeocurrents in the Yeneena Group is twofold; firstly, to assist the recognition of depositional environments, and secondly, to determine the basin palaeoslope. Particular depositional environments have characteristic patterns of palaeocurrent directions (e.g. Selley, 1968); for example, fluvial deposits and turbidites generally have unimodal dispersal patterns, near-shore tidally influenced sediments have bimodal patterns, and shallow shelf deposits commonly have bimodal or polymodal patterns (Pettijohn et al., 1973, p.140). Care must be taken in the determination of palaeoslopes from palaeocurrents, as currents can flow parallel to contours in some marine environments, and some directional structures can be orientated up-slope (on-shore) in tidally dominated environments. Palaeocurrents for fluvial sediments

Facies	Lithology	Bedding	Sedimentary structures
Isdell Formation			
C1	Dololutite, minor dolarenite	Medium to very thin	Dololutite - massive or laminated; dolarenite - massive, parallel- and cross lam., some starved ripples and ripple-drift cross lam., rare grading
C2	Dolarenite, and dolomitic carbon- aceous mudstone	Dolarenite- medium; dol. mudstone - very thin	Cross lam. very common in dolarenite; rare slumping in dol. mudstone
C3	Dolarenite and calcarenite, minor dololutite	Medium to very thick	Cross bedding
Malu and Telfer Formations			
S1	Siltstone and clay- stone, minor very fine grained sandstone	Thin to very thin	Siltstone - sharp bases and tops, cross lam., some grading, rare cross lam.; claystone - massive to faint parallel lam.; sandstone - grading, parallel-, cross-, and convolute lam., sharp slightly erosional bases
S2	Very fine grained well sorted sand- stone, minor interbedded mudstone	Thick to very thick	Mostly massive, some parallel- and cross lam., rare grading, sharp tops and bases, rare low angle cross stratification
S3	Very fine to medium grained sandstone (some bimodal), minor interbedded mud- stone, some calcarenite	Thin to thick	Flute marks, groove marks, load casts, grading, parallel- and cross lam.
S4	Medium grained sand- stone, minor inter- bedded mudstone	Thick to very thick	Sharp tops and bases, rare erosional and load sole marks, massive, very rare cross lam., rare grading.
C4	Dolarenite and dololutite	Mostly thin to medium, some thick	Dolarenite - sharp tops and bases, massive or parallel- and/or cross laminated; cross bedding in thick beds.

Table 4.1 Summary of facies characteristics of the Isdell, Malu and Telfer Formations.

and turbidites are generally regarded as being directed down the palaeoslope, although turbidity currents can turn at the base of a slope to flow parallel to the basin edge, and in some cases directional structures in the upper part of a turbidite can be orientated at right angles to basal erosional features (e.g. Selley, 1968).

4.2.1 METHODS

Cross stratification is the most common type of palaeocurrent indicator in the Yeneena Group, and this is followed in abundance in the Malu and Telfer Formations by flute casts; less common structures of use in this regard are parting lineation, channel orientation and pebble fabrics.

The accuracy with which these directional features could be measured varied between areas. The most readily measured features were the foreset dips of cross bedding in the Coolbro Sandstone (e.g. Plate 2.1B), and the sole marks exposed on vertical bedding planes in the Karakutikati Range (Plate 3.4F). The least readily measured were small sets of cross lamination exposed on flat surfaces in the Isdell Formation (e.g. Plates 3.2C and D) and larger scale cross bedding in the Puntapunta Formation (Plate 3.3F). At many outcrops in these two dolomitic formations the only direction that could be measured was the strike of the bedding, which represented a single vector component of the palaeocurrent direction; such measurements have not been included in the data below.

Three slightly different methods of palaeocurrent measurement were employed. 1) At outcrops beyond the limits of the Telfer Mine measurements were made of the apparent current direction relative to the strike and dip of the bedding, and corrections for the structural dip were then made. 2) At Telfer, orientated samples of cross lamination in the relatively soft weathered sediments exposed in the pits were collected, and were cut by diamond saw parallel to bedding planes, revealing plan views of the foresets. The orientated samples were then "unfolded" (i.e. the bedding plane was rotated about the strike to a horizontal position) and the palaeocurrent directions were measured directly from the foreset plan views. 3) Measurements from deeper levels around Main Dome at Telfer were made on orientated cores (e.g. Plate 3.5C).

In all of these determinations, corrections for structural dip were made on the assumption that folded beds had been tilted simply at right angles to the strike, and that no rotation parallel to the strike had occurred. This is a valid assumption for most areas of the upper Yeneena Group, as folding is dominantly concentric in form (Chapter 5). However, rotation of fault blocks could have occurred in some areas of the lower Yeneena Group; for example, along the southwest margin of the Rudall Metamorphic Complex (Fig. 2.2).

4.2.2 RESULTS

Rose diagrams illustrating the palaeocurrent directions of various stratigraphic units in the Yeneena Group are shown in Figure 4.1, and are discussed below.

Coolbro Sandstone

Relatively few palaeocurrent measurements have been made on this thick formation (Fig. 4.1A), all of which are from cross bedding in the upper part of the unit. Fourteen of the measurements were taken near locality B (e.g. Plate 2.1B) on Figure 2.2, and ranged in direction from 36° to 135°, with a mean of 64°. Four measurements from near locality A were more variable - 23°, 155°, 290° and 310° (Plate 2.1C).

The unimodal array of measurements from locality B would be consistent with a fluvial origin for the sandstones, the source of detritus being from the west-southwest. The varied current directions from near locality A are more difficult to reconcile with a fluvial origin, and may suggest a marine influence (see section 4.3.1).

Broadhurst and Choorun Formations

No directional sedimentary structures were found by the writer

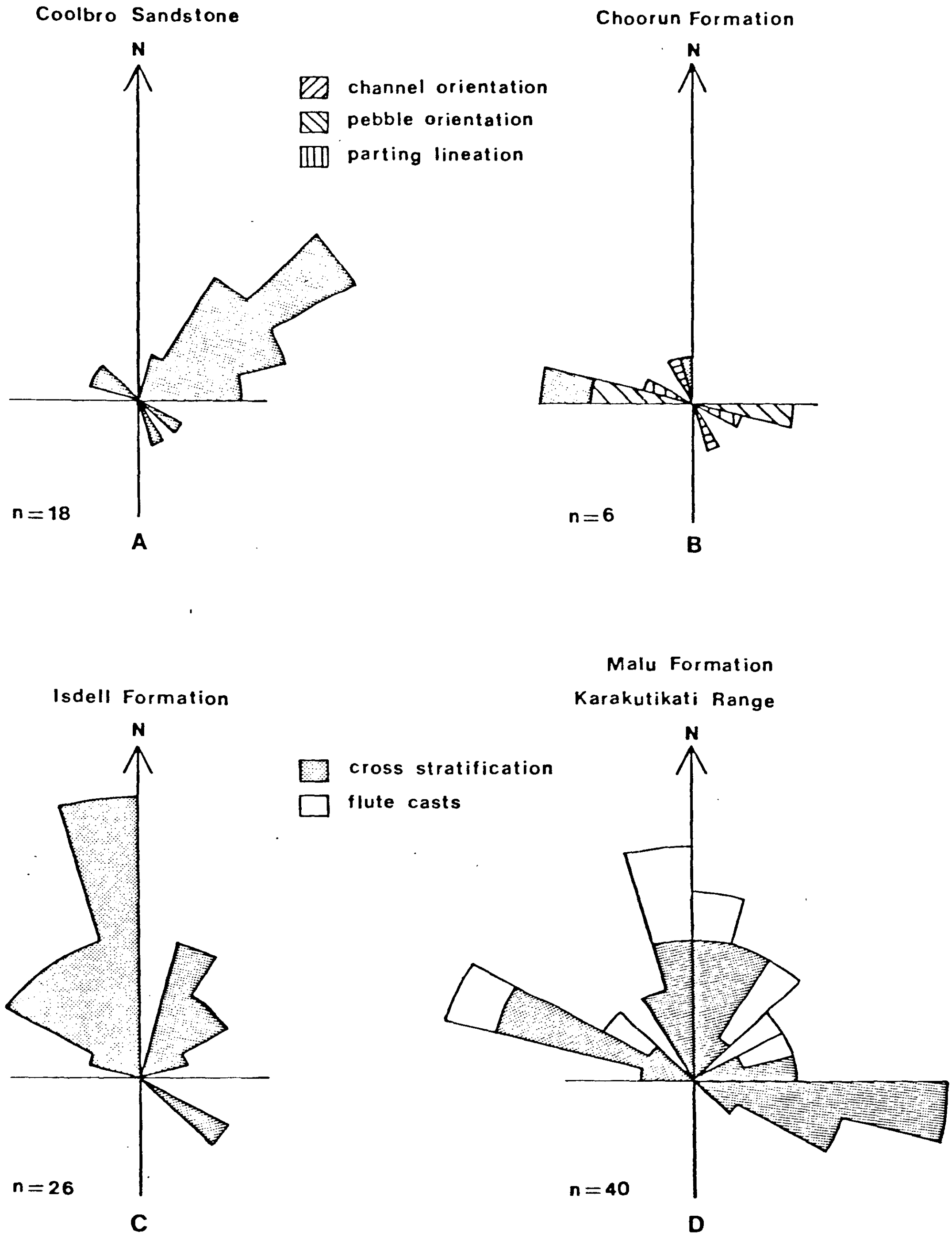


Figure 4.1 - Rose diagrams of palaeocurrent directions from various formations of the Yeneena Group. n indicates number of measurements. Most measurements from cross lamination or sole marks, but cross bedding in A and B. In B alternative directions are shown for channel orientation, pebble short axes and parting lineation.

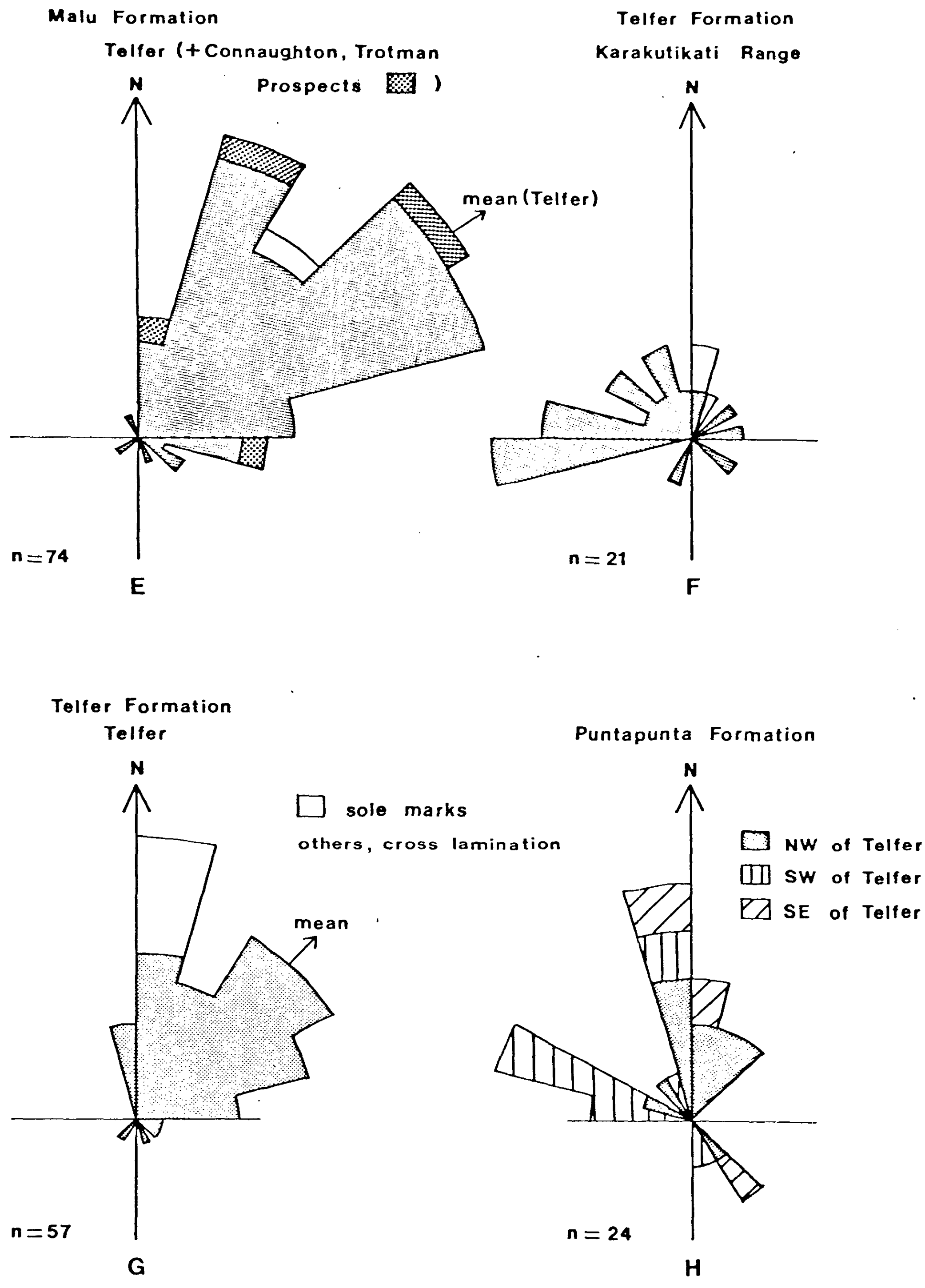


Figure 4.1 - continued.

in the Broadhurst Formation, and only a few were measured in the Choorun Formation (Fig. 4.1B). Four palaeocurrent indicators (including cross bedding, channel orientation and pebble orientation) at locality D (Figs. 2.2 and 2.3) show west-northwest and north-northwest directed currents; parting lineation near locality E (Fig. 2.2) also suggests a palaeocurrent to the west-northwest; and cross bedding at locality E indicates a palaeocurrent to the north. These few data suggest a general source area to the southeast for those fluvial sediments of the Choorun Formation located just west of the Rudall Metamorphic Complex.

Isdell Formation

Cross lamination in the Isdell Formation (Fig. 4.1C) shows that palaeocurrents were dominantly to the north-northwest and the northeast, with a few being to the southeast. Most current directions are confined to a 120° arc on either side of true north, and there are no measurements in the southwestern quadrant.

All of the current directions were derived from dolarenite beds in Facies C1 and C2, and all but three of these measurements were taken on the geological section shown in Figure 3.4. In this section there are distinct differences in orientation of palaeocurrents from Facies C1 and those from Facies C2. In Facies C1 twelve current directions were between 300° and 360°, and two were to 125°; in Facies C2, eleven current directions were between 355° and 70°, with one being to 310°. This is consistent with the interpretations (section 4.3.3) of many dolarenite beds in Facies C1 being contourites, and those in Facies C2 being deposited from traction currents flowing down the palaeoslope (to the northeast).

Malu Formation

The two major areas where directional sedimentary structures were measured in this formation are at the Karakutikati Range and at Telfer (Figs. 4.1D and 4.1E). In the former area there are three major palaeocurrent directions, towards the west-northwest, north-northeast and east-southeast. The diverse directions measured from flute casts, as well as from cross lamination, suggest that there was no single dominant palaeoslope during deposition of the Malu Formation. However, stratigraphic units within the formation at particular localities tend to have more restricted palaeocurrent directions, particularly from flute cast measurements (see Fig. 3.2), indicating that the slope direction varied with time (and locality). These patterns are consistent with the interpretation below of the Malu Formation as mass flow deposits laid down on submarine fans.

At Telfer, there is a unimodal pattern of palaeocurrent directions (Fig. 4.1E), with a mean towards the northeast (omitting the single southwesterly measurement in the calculation of the mean). This is interpreted to represent the dip of the palaeoslope. The few measurements from the Connaughton - Trotman Prospects area are also consistent with this trend (Fig. 4.1E).

Telfer Formation

In this formation at both the Karakutikati Range and at Telfer palaeocurrent measurements were restricted to the Outer Siltstone Member, due to the lack of directional structures in the Camp Sandstone Member. In the former area the currents were dominantly towards the west (Fig. 4.1F), but a few measurements are scattered around the compass. Measurements of two flute casts suggest a northerly palaeoslope, implying that the dominant currents were parallel to the slope.

However, at Telfer there is no great disparity between flute cast measurements and cross lamination measurements (Fig. 4.1G). The palaeocurrent direction pattern is similar to that for the Malu Formation at Telfer, and suggests a palaeoslope towards the northeast.

Puntapunta Formation

Three major distinct palaeocurrent directions have been measured from this dominantly carbonate formation, towards the west-northwest, the north and the southeast (Fig. 4.1H). This diversity is consistent with the interpretation below of a possible shelf origin for the formation.

4.3 DEPOSITIONAL PROCESSES AND ENVIRONMENTS

The interpretation of depositional processes and environments of Precambrian sediments is commonly hampered by the absence of palaeontological criteria, the structural complexity of the sedimentary sequence and the degree of metamorphism. However, there is a voluminous literature on Precambrian sedimentary rocks, and sufficient original characteristics of the sediments are preserved in many Precambrian successions around the world to allow fairly conclusive interpretations to be made. Interpretations of depositional processes are generally more conclusive than those of depositional environments.

Some of the formations of the Yeneena Group are easier to interpret than others due to their greater content of sedimentary structures, and their less pervasive deformation (Chapter 5). For some of the formations the present data are insufficient to enable the distinction between two possible alternative interpretations. The brief explanations of the formations and facies given in Chapters 2 and 3 are expanded and discussed in some detail below.

The sediments of the lower Yeneena Group contrast strongly with those of the upper Yeneena Group. The older sediments are interpreted as being deposited on a stable craton, whereas the younger sediments were probably deposited along the craton margin.

4.3.1 COOLBRO SANDSTONE AND CHOORUN FORMATION

The only previous published interpretations of these two sandstone formations are those of Chin et al. (1980), who considered that both units were of marine shelf origin. However, both formations are interpreted here as dominantly of fluvial origin, although a shallow shelf origin for part of the Coolbro Sandstone cannot be ruled out on the basis of the limited present evidence. In marked contrast with the siliclastic sediments of the Malu and Telfer Formations, these two older formations do not have features which are characteristic of deeper water mass flow deposits (section 4.3.4), and such mechanisms are not considered here. Also, there is no evidence of deltaic deposition, such as large scale coarsening upwards sequences with characteristic facies associations (e.g. Elliot, 1978), in the monotonous sandstones of the Coolbro Sandstone or Choorun Formation. Therefore this major sedimentary environment will not be considered further.

Evidence for fluvial deposition

Recent reviews of fluvial deposits have been given by Collinson (1978) and Walker and Cant (1979), and problems involved in the recognition of Proterozoic sandy fluvial systems have been documented by Long (1978). Three major alluvial environments can be distinguished - alluvial fans, braided rivers and meandering rivers. Each of these environments imposes a characteristic assemblage of features on the depositing sediments.

Alluvial fan deposits are generally coarse grained, with interbedded conglomerate and sandstone lenses, and only minor mudstone. The sediments are immature, and are commonly locally derived. On modern fans, sediment is either water-lain or deposited from debris (or mud) flows (Bull, 1972; Collinson, 1978). The water-lain beds may be deposited from sheet floods or storm surges. Sheet floods are shallow flows that spread over the fan surface and deposit sediment very rapidly in ephemeral channels. The resulting beds can be cross bedded, laminated or massive (Bull, 1972). Thick massive conglomerates can be deposited from storm surges (e.g. Scott and Gravlee, 1968).

The only sediments in the Coolbro Sandstone and Choorun Formation that might have been deposited on an alluvial fan are the basal massive conglomerates and interbedded sandstones of the Coolbro Sandstone. Such an interpretation is supported by the inclusion in the conglomerates of locally derived clasts from the underlying metamorphic rocks. The upward decrease in both clast size and conglomerate bed thickness from the basal unconformity (section 2.3.1), indicates waning of the depositional currents or retreat of the head of the fan. Other examples of Precambrian conglomeratic sediments that have been interpreted as alluvial fan deposits, and that overlie a major unconformity, have been documented by Williams (1969) and Turner and Walker (1973).

Braided streams can deposit thick sequences of pebbly or sandy sediments, and numerous Proterozoic sandstone sequences have been interpreted as possible braided stream deposits (e.g. see Long, 1978, p. 314; Winston, 1978; Eriksson and Vos, 1979). Such Proterozoic deposits are characterised by an abundance of planar and trough cross bedding, with lesser amounts of parallel lamination and ripple cross lamination; fining upward sequences occur, but are comparatively rare compared with Phanerozoic sequences; mudstones and associated features such as mudcracks are rare; and palaeocurrent patterns tend to be unimodal (Long, op.cit.). Detritus finer than sand is very sparse or absent because it can be transported through the high energy braided system without accumulation (Walker and Cant, 1979).

In the Coolbro Sandstone, planar and trough cross bedding (Plates 2.1B and 2.1C) were probably formed by migrating sand waves and dunes respectively (section 2.3.1), and in at least some areas a unimodal palaeocurrent pattern exists (section 4.2.2). These aspects would be consistent with deposition in a braided sandy fluvial environment. In a modern braided sandy fluvial system Walker and Cant (1979) proposed that trough cross bedding resulted from the migration of dunes in deeper channels, whereas planar cross bedding formed from the migration of cross-channel bars and sand waves on shallow sand flats. Parallel lamination can also be formed on sand flats during upper flow regime conditions of deposition. Convolute bedding formed by de-watering due to compaction, could also occur in sandstone deposited in such a fluvial environment (Eriksson and Vos, 1979).

The fining upward sequences of conglomerate, sandstone and purplish red shale in the Choorun Formation (Fig. 2.3) are also interpreted as fluvial sediments. Deposition may have taken place in either a braided or meandering stream environment. The former environment is perhaps more probable because the amount of shale is relatively minor and ripple cross lamination has not been recorded.

Typical meandering fining upward sequences commonly have a pebbly lag deposit at the base (usually with pebble long axes at right angles to the current - Harms et al. 1975), and pass upward through trough cross bedded sands, to ripple cross laminated fine sand (all deposited in a channel), and finally to silt and clay of the floodplain (Allen, 1970; Walker and Cant, 1979). Such meandering stream deposits have been recognised in very few Proterozoic sequences (Long, 1978). This may be a consequence of the absence of vegetation, which helps to stabilise channel banks in modern rivers (Schumm, 1968).

Possibility of shallow marine deposition.

In view of the previous interpretations of the Yeneena Group as marine shelf deposits (Chin and Hickman, 1977; Chin et al., 1980), the products of siliclastic shelf environments are briefly considered below.

There are relatively few documented examples of ancient shallow marine siliclastic sediments, compared with the numerous interpretations of ancient fluvial deposits. However, sedimentary processes on modern shelves have been extensively (but by no means completely) documented, and some ancient siliclastic deposits have been convincingly demonstrated to be of shelf origin. Recent reviews of the sedimentary processes and deposits in such environments have been given by Johnson (1978) and by Walker (1979), and only a brief summary is practical here.

On modern shelves mud and sand are the main deposits, and in ancient shelf sediments sandstone bodies are commonly surrounded by mudstone. The most important sediment transport mechanisms are storm and tidal currents, which can give rise to diverse palaeocurrent patterns; unimodal, bimodal and polymodal patterns are all possible (e.g. Pettijohn et al., 1973, p.140). Walker (op. cit.) recognised three main groups of modern shelf sand bodies. First, shoal retreat massifs (the inner shelf sand bars of Johnson, 1978), which are large sand sheets (tens of km in dimension, and probably 10 - 30 m thick) extending seawards from capes and estuaries, and which probably result from transgression over coastal sands. Second, linear sand ridges, tens of kilometres long, a few kilometres wide and a few tens of metres thick. These can be both storm-controlled and tide-controlled. Upward coarsening sequences can be developed in such ridges (Stubblefield et al., 1975). Third, sand waves, which are commonly much larger in scale than sand waves in fluvial systems, being several metres thick and having wavelengths of a few hundred metres. These are commonly formed by tidal currents, and can result in large scale cross beds, which if present can greatly assist the recognition of a shallow marine environment.

Johnson (1978, p.229) stated that "sedimentological data such as texture, sedimentary structures, lithofacies associations, sand body geometry and palaeocurrent patterns, are usually not in themselves diagnostic" of a shelf environment, but that "marine body fossils, trace fossils, certain minerals and geochemical parameters may be characteristic". Thus, the recognition of shelf environments in Precambrian siliclastic sediments is not straightforward.

However, thick Precambrian sandstone sequences have been interpreted as shelf deposits, the Jura Quartzite of Scotland (Anderton, 1976) perhaps being the best example. In this stratigraphic unit, which is up to 5 km thick, abundant cross bedding (generally 0.1 - 2 m thick, but up to 4.5 m thick) occurs in medium to coarse grained sandstone, which has been interpreted as the result of migrating dunes and sand waves (Anderton, op. cit.). In addition, very fine to medium grained sandstone occurs interbedded on the scale of less than 1 m with siltstone and mudstone. Parallel lamination, cross lamination and grading occur in the sandstone beds, which were probably deposited from storm-induced (turbidity) currents.

Such turbidites have also been recorded by Hamblin and Walker (1979) in Jurassic sediments from Western Canada. These authors also described hummocky cross stratification (Harms et al., 1975) in beds overlying the turbidites, and attributed this structure to storm wave action in a shallowing regressive sequence (the undisturbed turbidites having been deposited below storm wave base).

It is possible that some of these various shelf features could exist in parts of the Coolbro Sandstone which have not been examined in detail. However, the only documented feature which might suggest a shelf environment is the divergence of the sparse palaeocurrent measurements near location A (Fig. 2.2 and section 4.2.2). The association of cross bedding, parallel lamination, parting lineation and convolute bedding at location B (Fig. 2.2 and section 2.3.1) is thought much more likely to be the product of a fluvial depositional environment than that of a shallow shelf.

4.3.2 BROADHURST FORMATION

As stated in Chapter 2, the Broadhurst Formation accumulated in a quiet marine environment, which permitted the deposition of sparse iron formation, minor dolomite and abundant carbonaceous shales.

The origin of iron formation is problematical (Eugster and Chou, 1973; Pettijohn, 1975, p.420; Button, 1976; Dimroth, 1979), and the thin but laterally extensive examples in the Broadhurst Formation are insufficiently well known to suggest a definite mode of origin. Most genetic models for iron formations suggest deposition in a shelf or barred lagoon environment (e.g. James, 1954), but deposition in playa lakes has also been advocated (Eugster and Chou, 1973), and Dimroth outlined a pelagic origin for Algoma-type (volcanic-associated, cherty) iron formations.

The iron formation of the Broadhurst Formation is basically iron-rich shale; the iron minerals are thought to have been chemically precipitated during restricted circulation in the basin. The association of iron formation with dolomite (section 2.3.2) supports such an origin (c.f. Button, 1976). No evidence of volcanism has been recorded in the Broadhurst Formation that might suggest a volcanic origin for the iron-rich sediments.

Carbonaceous shales were probably deposited under anoxic sea floor conditions in a restricted basin (e.g. Hallam and Bradshaw, 1979), and the thin graded siltstone beds within "normal" shale from the formation (section 2.3.2) indicate periodic influxes of dilute turbidity currents during periods of less restricted circulation. The association of very fine grained detrital and chemical sediments suggests deposition in a tectonically stable tranquil marine setting.

4.3.3 ISDELL FORMATION

Interpretation of the Isdell Formation presents a number of problems, which include, a) the origin of the dololutite in Facies C1, b) the origin of the dolomitic mudstone in Facies C2, c) the origin of dolomitic sand and the nature of the currents which introduced this sand into the depositional basin, d) the high content of albite in some dolarenite samples, and e) the timing of dolomitisation. Some of these considerations are also pertinent to the Puntapunta Formation (section 4.3.5).

Dololutite

This widespread lithology of the Isdell Formation resembles pelagic carbonate. However, true pelagic carbonate sediments only occur in sequences of Upper Silurian and younger ages, as they are derived from the comminuted skeletons of pelagic calcareous micro-fossils, which do not occur in older rocks (e.g. Tucker, 1974).

Carbonate mud can also be produced on shallow carbonate platforms. On modern platforms lime mud can be formed by mechanical or biological abrasion of larger carbonate particles, or biogenically precipitated by calcareous algae; it is uncertain whether or not it can also be formed by direct inorganic precipitation from sea water (Blatt et al., 1972, p.424). Carbonate mud so produced on a platform may be deposited there, or may be swept off the platform to be deposited on an adjacent slope to form evenly-bedded carbonate mud. Such carbonate mud has been termed "peri-platform ooze" by Schlager and James (1978). According to McIlreath and James (1979) most Precambrian pelagic slope carbonates may well have been almost wholly peri-platform ooze.

The dololutite of Facies C1 is interpreted as such carbonate, deposited on a gentle slope in quiet relatively deep water. The slope is considered to have been gentle due to the absence in Facies C1 of carbonate breccias or slump folds, which are characteristic of some carbonate slope deposits elsewhere (Cook and Taylor, 1977; McIlreath and James, 1979). However, frequent periodic influxes of carbonate and siliclastic sand occurred, producing the interbedded dolarenite layers (see below), which suggests accumulation on a slight slope. The depositional environment must have been tranquil to allow the carbonate mud to settle from suspension. The water depth is uncertain, but it must have exceeded storm wave base; modern pelagic carbonate deposition occurs in water depths of less than 100 m to greater than 4500 m (McIlreath and James, op. cit.).

Relatively deep water deposition for the entire Isdell Formation is also suggested by the complete absence of stromatolites, which are abundant in many Proterozoic carbonate sequences, and indicate deposition of carbonate in shallow water (in the photic zone) by algal colonies (e.g. Truswell and Eriksson, 1972; Button, 1973; Hoffman, 1974; Young, 1974).

Dolomitic mudstone

The dark grey dolomitic mudstone of Facies C2 (Plate 3.2G) differs from the dololutite of Facies C1 by having a higher content of terrigenous clay and a higher content of organic carbon. However, the environment of deposition is suggested to be a similar carbonate slope, but with an increased supply of terrigenous detritus. The soft sediment folds shown in Plate 3.3B strongly suggest deposition on such a slope. Relatively rapid deposition was probable, which prevented the oxidation of carbonaceous debris, which was probably derived from marine algae.

Dolarenite

The thin dolarenite beds of Facies C1 represent periodic influxes of carbonate, quartz and feldspar sand onto the carbonate slope. Some of the sand was introduced by turbidity currents. For example, the parallel lamination overlain by climbing ripples in a dolarenite bed shown in Plate 3.2D represent the Bouma B and C divisions of a turbidite (see section 4.3.4 for a general description of turbidites). However, many of the very thin dolarenite beds show no sequences of sedimentary structures. These sands were transported by traction currents, which resulted in parallel- or cross lamination within the beds (e.g. Plates 3.2B and 3.2C). The thinness of these beds, the sharp tops and bases, the evidence for traction current emplacement, the absence of definite turbidite features and the diverse palaeocurrent directions (section 4.1.2) suggest that the dolarenite beds are contourites (see section 4.3.4 for a general description of contourites).

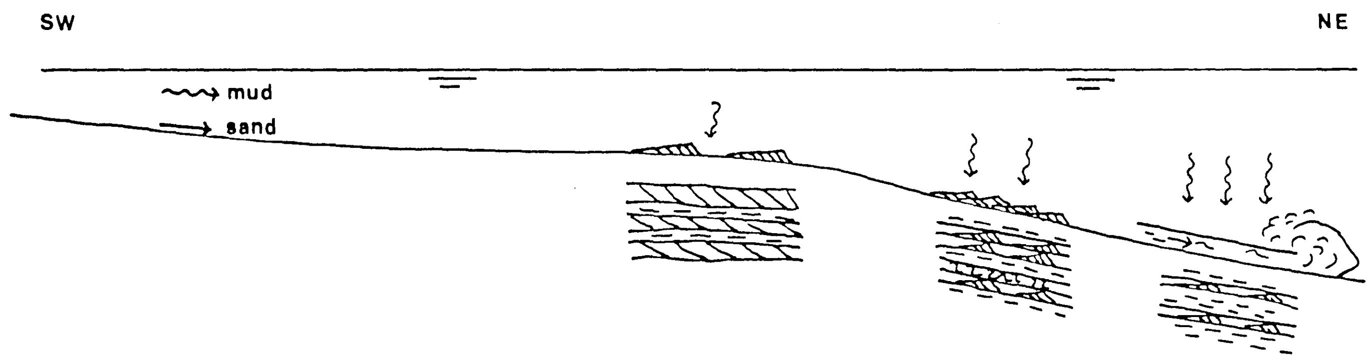
The beds of cross laminated dolarenite in Facies C2 indicate more continuous deposition of carbonate sand by traction currents. In at least some areas the palaeocurrents are at variance with those from Facies C1, and are thought to have been directed down the palaeoslope.

The thick-bedded, cross stratified dolarenite of Facies C3 was also deposited from traction currents. The depositional environment is suggested to be the outer part of a carbonate shelf, where cross bedding (Plate 3.3F) was caused by the migration of bed forms such as sand waves. A schematic summary of the postulated depositional environments for the three facies is shown in Figure 4.2.

The origin of carbonate sand particles within the dolarenites is uncertain, but the most likely origin is in an inner shelf or coastal environment. The particles could have formed by the breakdown of algal mats, or possibly by the aggregation of carbonate mud into fecal pellets and grapestone lumps by soft-bodied organisms (e.g. Wilson, 1975, p.5). It must be re-emphasised, however, that no stromatolites occur within the Isdell Formation, and any such inner shelf or coastal environment must have been far to the south or southwest of the present Isdell Formation outcrops.

Albite in dolarenite

Dolarenite beds commonly include over 25% detrital albitic plagioclase, and in some samples the content exceeds 40% (Table 3.3).



Inner shelf	Outer shelf	Carbonate slope	
Formation of carbonate sands and muds. Transport to deeper water by traction currents and in suspension.	Deposition of Facies C3; dolarenite deposited by traction currents, minor carbonate mud from suspension.	Localised deposition of Facies C2; dolarenite deposited from traction currents, rapid deposition of carbonaceous mud.	Deposition of Facies C1; abundant carbonate mud from suspension, thin dolarenite beds deposited from turbidity currents and contour currents.

Figure 4.2 - Hypothetical depositional environments for Facies C1, C2 and C3 of the Isdell Formation.

The amount of detrital quartz is often much less than that of feldspar. Such a disparity is unusual in all but the most arkosic of sands, and it is suggested here that much of the plagioclase could be of tuffaceous origin. Abundant plagioclase occurs in andesitic and dacitic crystal tuffs, which if eroded and redeposited could result in feldspathic sands. Subsequent alteration during low grade regional metamorphism converted the plagioclase to albite.

The formation of dolomite

In ancient carbonate rocks it is commonly uncertain whether dolomite was formed as a primary mineral by direct precipitation from sea water, or whether it is a replacement product of calcium carbonate (Blatt et al., 1972, p.479). This is also a problem for the abundant dolomite in the Isdell (and also the Puntapunta) Formation. A full discussion is beyond the scope of the present work, but the following tentative suggestions can be made.

It is probable that dolomite is not a late-stage alteration product, as no obvious replacement textures (such as dolomite rhombs penetrating calcite crystals) have been recorded. However, the grain size is so small in many of the rocks that such replacement relationships, if present, are unlikely to be obseved. It is thought probable that dolomite is either primary, or was formed by the metasomatic alteration of calcium carbonate during very early diagenesis, by reaction of the carbonate crystals with sea water at or near the sediment - water interface. Such a metasomatic origin has been postulated for rare examples of modern dolomitic sediments, such as in Deep Spring Lake in the western U.S.A. (Peterson et al., 1966), and primary dolomite formation has been demonstrated in the lagoonal setting of the Coorong in South Australia (von der Borch, 1976).

4.3.4 MALU AND TELFER FORMATIONS

Deposition of these two formations was regarded by Chin and Hickman (1977) and Capill (1977) as having taken place in a shelf environment. However, due to the presence of mass flow deposits (see below) and the general absence of shelf characteristics (in particular the absence of large-scale cross stratification - see section 4.2.1) the five facies of these two formations are interpreted as relatively deep marine deposits. No absolute depositional depths can be placed on the facies, because, in the absence of palaeontological depth indicators, the terms "shallow" and "deep" can only refer to basins which are regularly current- or wave-agitated, or normally quiet, respectively (e.g. Turner and Walker, 1973).

Characteristics of mass flow deposits

Mass flows (or sediment gravity flow) are relatively well known processes that can transport and deposit very large quantities of terrigenous detritus of all grain sizes. The following discussion is mainly concerned with the deposition of sand, there being no coarser detritus in the Malu and Telfer Formations. Four types of mass flows and their resulting deposits have been described by Middleton and Hampton (1976) and Rupke (1978). These are, turbidity currents, fluidised flows, grain flows and debris flows, each of which can deposit a bed with a characteristic assemblage of sedimentary structures.

Turbidity currents are flows in which the entrained sediment is supported by fluid turbulence. The mechanics of sediment transport in such flows has been described by Middleton (1966a and b) and Middleton and Hampton (op. cit.). The resulting deposits of such flows are turbidites. "Classic" turbidites (Walker, 1978) can be recognised by a suite of sedimentary structures, the most important of which are; a) erosional sole marks on the sharp bases of sandstone beds; b) grading in the sandstone bed; c) the common presence of an entire or partial sequence of internal structures (the Bouma sequence - Bouma, 1962), from massive or graded at the base (division A), upward through parallel lamination (division B), cross lamination (division C) and further faint flat lamination (division D) to mudstone (division E); and d) a bedding regularity, where individual beds can be traced for long distances without appreciable thickness changes.

A distinction can be made between proximal turbidites, and distal or thin-bedded turbidites (Walker, 1967 and 1978). The former are deposited in the initial area of deposition, and tend to be thicker bedded, coarser grained and begin with Bouma division A, whereas the latter deposits are thinner bedded, finer grained and begin with higher Bouma divisions.

In grain flows the entrained sediment is supported by grain to grain interactions. Grain flow deposits differ from classic turbidites by being generally ungraded and having few internal sedimentary structures. However, diffuse parallel lamination can occur at any level, and dish structures (Stauffer, 1967) may also be present. The bases of beds are sharp and are either flat or have odd load casts and scours, and the tops of beds are also sharp and flat. Mudstone clasts can occur at any level in the bed (Stauffer, op. cit.).

Fluidised flows are rather similar to grain flows, but sediment is supported in the flow by escaping pore fluids (Middleton and Hampton, 1976). The resulting deposits may be poorly graded, have sharp bases with load-induced and erosional sole marks, contain dish structures, water escape pipes and convolute lamination, and have sharp flat tops.

In debris flows, the larger grains are supported by a matrix of fine sediment and interstital fluid. The resulting deposits are coarse grained (commonly of boulder grade), poorly sorted, massive to poorly graded, and have sharp but irregular tops and bases (Middleton and Hampton, op. cit.).

The types of mass flows outlined above are conceptual end members of a variety of flow types, and in nature there are gradations between types of flows (Middleton and Hampton, op. cit.). However, sandstone sequences, and indeed individual sandstone beds, can commonly be assigned to one or other of these flow types, which can have palaeogeographic implications, as discussed in the section on depositional environments below.

Characteristics of contourites

In modern deep ocean basins thermohaline currents flowing parallel to sea-bed contours along continental margins are capable of transporting and redpositing mud, silt and sand; the resulting deposits are termed contourites (e.g. Bouma and Holliser, 1973). These deposits occur in the same depositional environments as many sediment gravity flow deposits, and therefore the possibility that some of the fine grained siliclastic sediments of the Malu and Telfer Formations are contourites must be considered.

Stow and Lovell (1979) have given an extensive review of contourites, and have compared and contrasted these deposits with turbidites. Thin-bedded turbidites are not always readily distinguishable from contourites, but the characteristics of the latter can be summarised as follows; a) relatively thin-bedded, normally less than 5 cm thick; b) sharp bases and tops; c) ubiquitously laminated, with heavy mineral concentrations along parallel- and cross laminae; d) relatively fine grained, normally of silt or clay grade; e) well to very well sorted, with less than 5% matrix; f) normal or reverse graded; and g) with a dispersal pattern parallel to the basin margin (Rupke, 1978, p.389; Stow and Lovell, 1979).

Deposition of Facies S1 - S4

The sediments of Facies S1 - S4 include undoubted turbidites and probable grain flow deposits. Fluidised flow deposits have not been recognised, and debris flow deposits do not occur. In addition, the presence of contourites is doubtful.

The sandstone beds of Facies S3 are the most distinctive turbidites in the sequence, due to the presence of such features as flute casts (Plates 3.4F and 3.4G), tool marks (Plate 3.4H), graded bedding, parallel lamination (Plate 3.5A) and cross lamination (Plate 3.5B). Flute casts were produced by erosional eddying at the base of the turbidity current (see Allen, 1971), and tool marks were probably formed by small mudstone clasts (which were eroded from the underlying substrate) being dragged along the depositional surface. Graded bedding and Bouma sequences of sedimentary structures were formed during deposition in decelerating flows; the massive to graded division A represents deposition of a "quick" bed, division B represents traction current deposition in the upper flow regime, and division C represents deposition in the lower flow regime (Walker, 1965).

The generally massive fine to very fine grained sandstones of Facies S2 are possibly also turbidites, although the most diagnostic features such as sole marks and sequences of internal sedimentary structures have not been recorded. The massive nature of these beds might suggest a grain flow or fluidised flow origin, but the presence of cross lamination within the lower and middle parts of beds indicates traction current deposition for at least some of the sand.

The rare low angle cross stratification and the features resembling ripple marks which occur in Facies S2 sandstone (section 3.3.2) are difficult to interpret. It is possible that these features represent reworking of mass flow sands, perhaps by storm wave action. If this is so, then deposition can be inferred to have taken place in relatively shallow water (c.f. the hummocky cross stratification in sandstone overlying inferred shelf turbidites, of Hamblin and Walker (1979) - section 4.3.1). However, there is no substantive evidence that all Facies S2 sandstones were deposited in a shallow shelf environment (see below).

The medium grained sandstones of Facies S4 are interpreted as grain flow deposits. They possess many of the characteristic features, such as sharp slightly erosional bases (Plate 3.5F) and sharp tops, massive to faintly laminated interiors and enclosed mudstone flakes. However, no dish structures have been observed.

It is uncertain whether the siltstone beds within Facies S1 are thin-bedded turbidites or contourites (section 3.3.2). The bedding characteristics could fit into either category of deposit, and the lack of directional sedimentary structures within the beds renders it impossible to determine the direction of transport relative to the depositional slope, indicated by sole marks beneath sandstone beds (section 4.2.2). However, the association of Facies S1 with sandy turbidites suggests that the siltstone beds are more likely to be thin-bedded turbidites than contourites.

Deposition of Facies C4

The parallel-sided graded dolarenite beds of this facies in the central part of the Outer Siltstone Member at the Karakutikati Range (Plates 3.6C and 3.6D) are interpreted as turbidites (section 3.3.2), which were derived from the east-southeast (i.e. the palaeocurrents were directed towards the west-northwest - see Figs. 4.1F and 3.2). The thin interbeds of dololutite represent deposition of carbonate mud in a quiet marine environment. The thicker cross bedded units which occur at intervals within the turbidites (Plate 3.6E) represent deposition on sand waves by traction currents, suggesting a carbonate shelf environment of deposition (see below).

Depositional environments of the Malu and Telfer Formations

In the Malu Formation, and in the lowest part of the Telfer Formation, the four siliclastic facies can be grouped into a number of sequences, which in most cases coarsen upward. This is best exemplified at the Karakutikati Range, where eight coarsening upward cycles and one fining upward cycle can be recognised (Fig. 3.2). A typical coarsening upward cycle occurs at the base of the formation, where mudstone of Facies S1 passes up into a unit of interbedded turbidites and mudstone (Facies S3 and S1), and then into a unit of turbidites and grain flow deposits (Facies S3 and S4). The thick-bedded medium grained sandstones of Facies S4 are abruptly overlain by mudstone of Facies S1, and this boundary defines the base of another coarsening upward unit (Fig. 3.2). Such cyclic sedimentation of mass flow deposits can best be interpreted as due to deposition on prograding submarine fans, as outlined below. The dominant direction of progradation was to the north or northeast, as indicated by both palaeocurrent directions (section 4.2.2), and by the northwards thickening of the Malu and Telfer Formations (sections 3.2.2 and 3.2.3).

Submarine fan models have been proposed in recent years, which relate the distribution of mass flow types to fan morphology (e.g.

Walker, 1978). Deposits characteristic of upper, middle and lower fans can be distinguished. The upper fan consists of feeder channels, which are probably the sites of conglomerate deposition, and inter-channel areas, where mudstones and thin-bedded turbidites accumulate. The middle fan areas are sites of suprafan lobes, where deposition of massive and pebbly sandstone probably occurs in braided shallow channels on upper parts of the lobes, and more typical turbidites are deposited on the lower parts of the lobes. On the lower fan and basin plain, thin-bedded turbidites and mudstones accumulate. Two basic types of vertical sequences result from deposition on submarine fans (Rupke, 1978). Thickening and coarsening up sequences reflect localised deposition by prograding depositional lobes, and thinning and fining up sequences are characteristic of the lateral migration or gradual abandonment of a channel.

Several of the coarsening upward sequences in the Malu and lowermost Telfer Formations are interpreted as the result of deposition on prograding fan lobes. Thick sequences of Facies S1 could represent lower fan or basin plain deposits, but thin units of this facies interbedded with coarser sediments (particularly Facies S4) may represent deposition on elevated parts of the middle fan. Facies S2 and S3 were probably deposited on the lower parts of suprafan lobes, and sandstones of Facies S4 could have been deposited in shallow channels on the upper parts of these lobes. No deposits indicative of an upper fan environment occur in the sequence.

The dimension and number of co-existing fan lobes present during deposition of the Malu and Lower Telfer Formations is uncertain; very detailed mapping of this sequence is needed to determine the lateral extent of individual coarsening upwards units. However, the lateral change from Facies S4 to S3 in the upper Footwall Sandstone Member at the Karakutikati Range (described under the heading of Facies S4 in section 3.3.2), may represent the down-fan change from deposition of massive grain flow sandstones in migrating shallow channels to finer grained turbidites lower on the suprafan lobe (Fig. 4.3).

Some of the coarser grained sedimentary units are of great lateral extent; for example the Rim Sandstone Member extends as a unit of fairly constant thickness for over 25km along the length of the Karakutikati Range. This suggests that at times there may have been a tectonic, as well as a sedimentological, control on the distribution of facies. The slope on which the sediments were deposited may have been fault-controlled, and if so an episode of active faulting could conceivably have caused an increase in the degree of the slope. This may have allowed periodic transport and deposition of medium grained sand by mass flows initiated at numerous points along the length of the slope, resulting in a laterally extensive unit comprising individual sandstone beds of more restricted extent (up to a maximum of perhaps a few kilometres wide). Reversion to periods of less active faulting may have allowed the resumption of mass flow deposition on large fans.

Following deposition of the final coarsening upward cycle in the lowest part of the Telfer Formation (Fig. 3.2), the depositional environment changed, and carbonate sedimentation became important. The environment at the Karakutikati Range is interpreted as the outer part of a carbonate shelf, where carbonate muds and carbonate turbidites could be deposited below storm wave base, but where occasionally persistent traction currents deposited cross bedded carbonate sands. Siliclastic sediments were still periodically deposited, but the coarsest detritus was the very fine to fine grained sands of Facies S2 (e.g. the Camp Sandstone Member). Some of these sands are probably turbidites, but some may have been deposited (or reworked) by shelf currents, which resulted in the rare examples of low angle cross bedding in sandstone of this facies (see above, and section 3.3.2). North of the Karakutikati Range siliclastic sediments are dominant in the Telfer Formation, suggesting deeper water sedimentation than for the carbonate sediments of this range.

The Telfer Formation essentially represents a transitional period of sedimentation, between the presumably deep-water deposition of siliclastic sediments on submarine slopes and fans (Malu and lowermost Telfer Formations) to deposition of carbonate sediments in a shelf environment (Puntapunta Formation - see below).

4.3.5 PUNTAPUNTA FORMATION

An outer shelf environment is suggested for both the carbonates and the carbonate-rich sandstones of the Puntapunta Formation. Some of the parallel-sided dolarenite beds interbedded with dololutite (Plate 3.6F) may be turbidites, and if so, deposition must have taken place beneath wave base. However, cross bedding in carbonate arenites (Plate 3.7A), and intraformational conglomerate (Plate 3.7B), indicate intermittent strong traction currents. Also, the wavy bedding in some of these carbonate arenites (Plate 3.6G) could be the result of storm wave action, as the structure resembles the hummocky cross stratification described by Harms et al. (1975, p.87) and Hamblin and Walker (1979), which has been interpreted as due to strong current surges of varied direction caused by storm waves in shallow water.

In common with the Isdell Formation, no stromatolites occur within the Puntapunta Formation, and thus a very shallow or coastal environment can be rejected. However, such an environment may have existed farther south or southwest during deposition of the Puntapunta Formation, where carbonate sand and mud may have been produced (see section 4.3.3).

4.3.6 WILKI QUARTZITE

The environment of deposition of the Wilki Quartzite is uncertain, due to the lack of recognisable sedimentary characteristics in this extensively recrystallised formation, but tentative conclusions can be drawn from the gross characteristics of the formation.

Unlike the Malu and Telfer Formations interbedded mudstone does not occur in the Wilki Quartzite, the unit being dominated by quartz arenite. This suggests that the sediments were deposited in a relatively high energy environment, and a deep water mass flow origin can be rejected, particularly in view of the absence of any definite turbidite features.

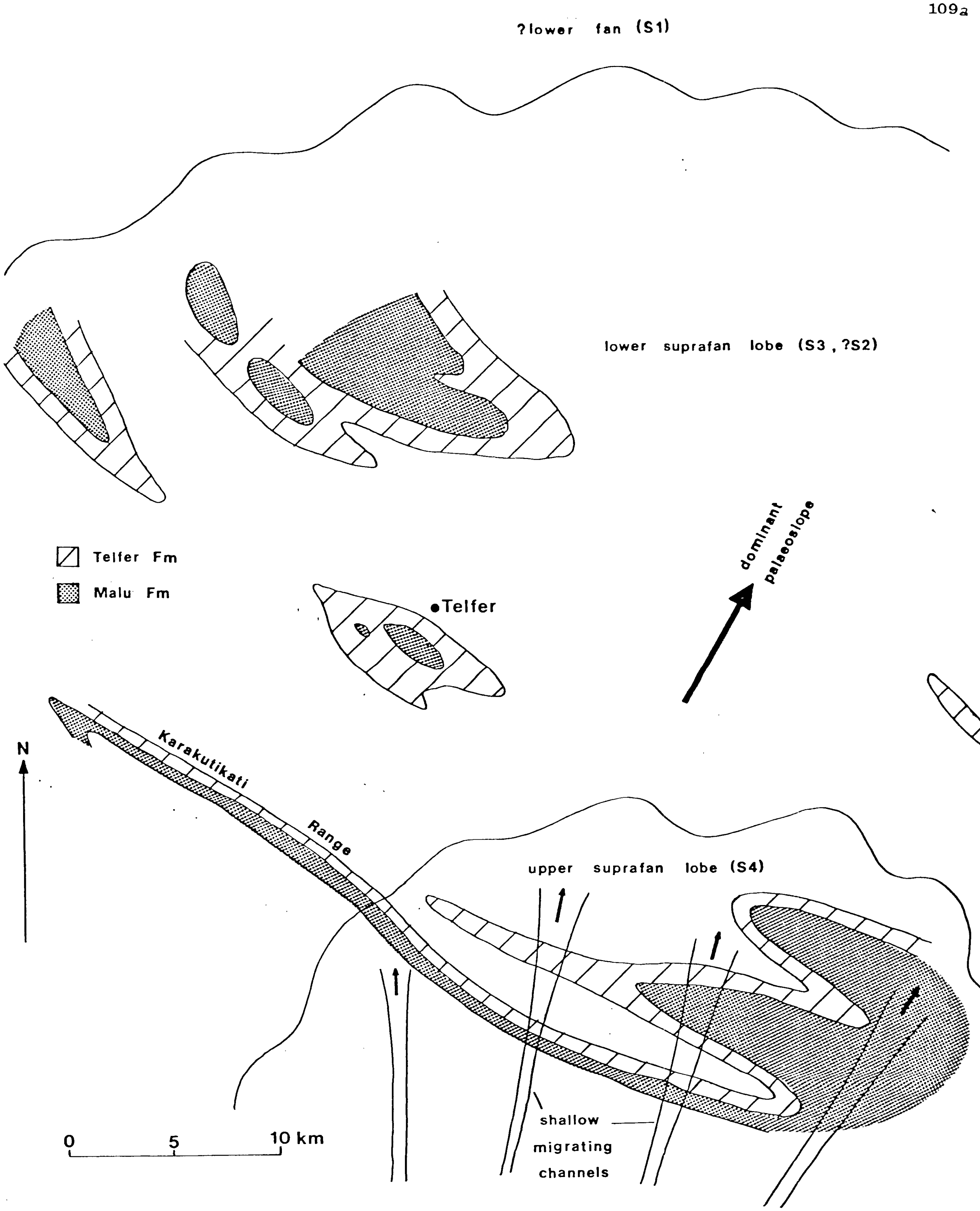


Figure 4.3 - Hypothetical palaeogeography (submarine fan) at the time of deposition of the upper Footwall Sandstone Member at the Karakutikati Range, where the southeastwards transition from Facies S3 to Facies S4 can be seen (see text). Geology from Figure 3.1.

One feature of the formation that is remarkable is the laterally extensive nature of sedimentation units about 5 m thick (Plate 3.7C). Such extensive sheet sandstone bodies are probably more likely to have been formed in a marine environment than in a continental setting, where lenticular bedding and channelling might be expected. The entire formation is therefore suggested to represent siliclastic shelf deposits.

4.4 TECTONIC SETTING AND BASIN EVOLUTION

The entire Yeneena Group could be described as a miogeosyncline (as defined, for example, by Aubouin, 1965), for the following reasons. The group is a very thick sequence of folded sedimentary rocks which have suffered low grade regional metamorphism and have later been intruded by granite (Chapter 5); apart from rare dolerite dykes (Chapter 1) no volcanic rocks are known within the group. However, the use of this descriptive term does not imply a unique plate tectonic setting for the group.

Two plausible tectonic settings can be considered for this part of Western Australia during deposition of the Yeneena Group. The first is that of a passive cratonic margin and the second is that of an intracratonic basin. Both possibilities would have been the result of an overall extensional tectonic regime.

In the first case deposition would have taken place along the northeastern margin of the Archaean to Lower Proterozoic Western Australian shield, from which the detritus forming the clastic units of the Yeneena Group would have been derived. Deposition in terrestrial, shallow marine, slope and deeper marine environments could have taken place in such a tectonic setting. These environments may have co-existed as lateral equivalents, in which case the Yeneena Group could have formed, at least in part, by large-scale slow migrations of the sedimentary environments. However, the relatively abrupt upward change from one major unit of the Yeneena Group to another, and the apparent layer-cake relationship between at least the units of the upper part of the group, suggest that the major causes of the vertical facies variations were tectonic and climatic changes that affected the entire basin.

The second possibility, that of an intracratonic basin setting for the Yeneena Group, is therefore considered the most feasible. In this case deposition may have taken place in the southwestern part of a failed rift (or aulacogen - Burke, 1977; Mitchell and Reading, 1978). Despite the uncertainty of the stratigraphic relationship between the lower and upper Yeneena Group (Chapter 3) the succession shown in Figure 2.1 is thought broadly to represent a time-sequence which can be interpreted in terms of an evolving aulacogen.

Three major phases of basin evolution can be recognised. The earliest phase was terrestrial sedimentation (Coolbro Sandstone and Choorun Formation) and ?shallow restricted marine deposition (Broadhurst Formation). This was followed by a period of deeper water sedimentation, firstly of carbonates, probably in a slope environment (Isdell Formation), and secondly by siliclastic deposition on submarine fans (Malu Formation). The third phase of basin evolution is recorded by the transition from deep water to shelf environments (Telfer Formation), deposition of a thick carbonate sequence on the outer part of a shelf (Puntapunta Formation) and finally deposition of ?shallow marine siliclastic sediments (Wilki Quartzite).

These three phases are comparable with the major episodes of basin development in some younger intracratonic basins, such as the North Sea (Mitchell and Reading, 1978; Ziegler 1975). The first phase represented the establishment of a broad subsiding basin where terrestrial and shallow marine deposition took place. The second phase represented the main rifting stage when deepening of the basin occurred accompanied by mass flow sedimentation. The Malu and lower Telfer Formations are the major units of the Yeneena Group which could have formed at this time. Down-to-the-northeast faulting south of the Karakutikati Range may possibly have provided the main locus for the initiation of mass flows, which spread northeastwards into the basin.

Faulting within the basin at this time is suggested by thickness variations in some sedimentary units at Telfer (see section 5.2.3). Due to the abundance of feldspar in parts of the Isdell Formation it is possible that the earlier part of the basin-deepening phase was accompanied elsewhere in the basin by intermediate to acidic volcanism. The final phase of basin evolution can be explained by the cessation of active rifting and the continued show subsidence of the entire basin. As the basin was infilled shallow-water deposition became prevalent once more.

The two major carbonate units (the Isdell and Puntapunta Formations) may represent periods of climatic change (probably increased aridity), when relatively small amounts of siliclastic detritus were being supplied to the basin.

CHAPTER 5

DEFORMATION, METAMORPHISM, GRANITE INTRUSION AND PHANEROZOIC GEOLOGY

5.1 INTRODUCTION

Following deposition of the sediments of the Yeneena Group the Precambrian history of the Paterson Province proceeded through a period of extensive folding, faulting and metamorphism, which was followed by the intrusion of two granite bodies in the Telfer area. These facets of the geology are obviously important to the understanding of the entire geological history of the region, knowledge of which is necessary before the Telfer gold deposits can be placed in their correct perspective. The writer's field work and petrographic examination of the Yeneena Group have added new data on the structural geology and metamorphism of the area to the details already determined by the Geological Survey of Western Australia and by Newmont. These aspects are described below, and for completeness, a short section on the geological evolution of the area during Phanerozoic times is also given.

5.2 DEFORMATION

5.2.1 REGIONAL ASPECTS

In the Rudall Metamorphic Complex to the south of Telfer (Fig. 2.2) two periods of pre-Yeneena Group deformation (D1 and D2) have been recognised by Chin et al. (1980). The major structural trend of the Paterson Province however, was caused by a third deformation (D3) which post-dated the Yeneena Group. As a result of this deformation both the Yeneena Group and the Rudall Metamorphic Complex are folded and faulted along northwest-southeast to north-northwest-south-southeast trends (Figs. 2.2, 3.1). The Upper Proterozoic Bangemall Group was deposited after the D3 deformation, and was then affected by a further gentle deformation (D4 of Chin et al., 1980) with northeast trending fold axes. No clear evidence of this D4 deformation exists in the Yeneena Group north of the Rudall Metamorphic Complex.

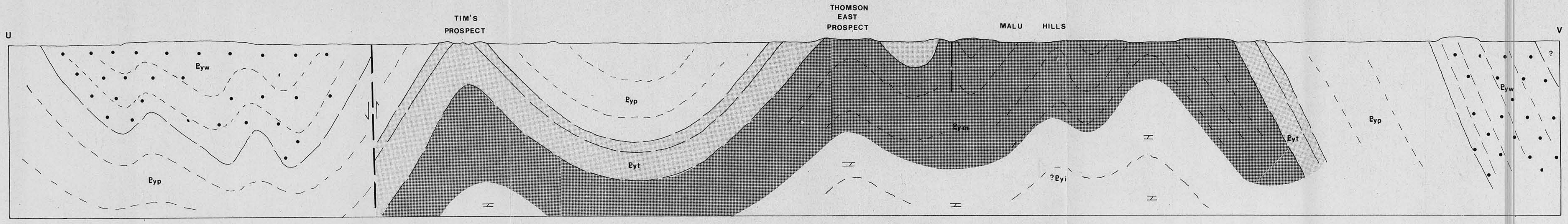
The D3 deformation of the Coolbro Sandstone in the Rudall River area resulted in open folds with the widespread formation of shear planes which obscure the bedding over wide tracts of the outcrop. These shears represent axial plane cleavage, and the folding is therefore of similar style. This can be demonstrated at a locality along the track about 3 km south of Coolbro Creek (Fig. 1.1), where small scale folds with horizontal axes and amplitudes of about 50 cm are associated with numerous vertical axial shear planes.

On the northeast side of the Coolbro Sandstone outcrop, tight folding occurs in the Broadhurst Range area (Fig. 2.2). Here, axial plane cleavage is very well developed in the shaly sediments of the Broadhurst Formation, indicating that the deformation is also of similar fold style.

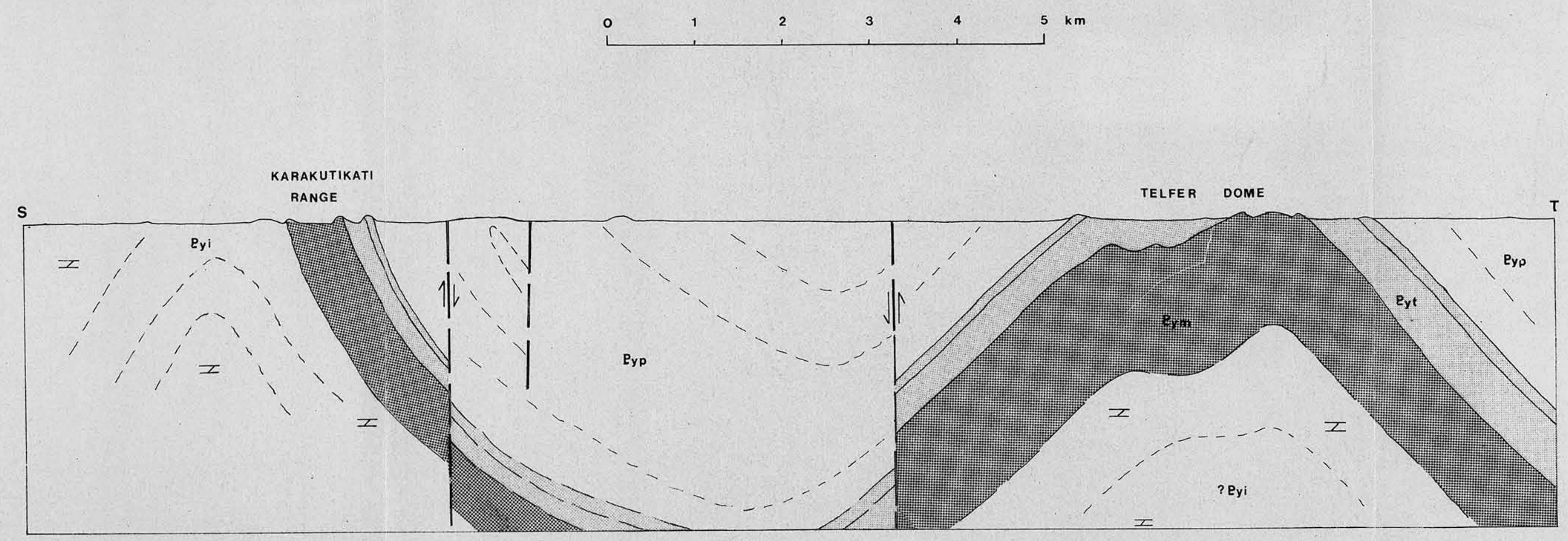
On the southwest side of the Coolbro Sandstone outcrop, and also along the southwestern margin of the Rudall Metamorphic Complex (Fig. 2.2) numerous thrust faults occur (Chin et al., 1980), the northeast side of the fault zone being the upthrust side. Chin et al. stated that this fault zone corresponds to the steep gravity gradient on the southwest margin of the "Warri Gravity Ridge" (i.e. the Anketell Regional Gravity Ridge of Fraser (1976) - see Fig. 1.4), and suggested that crustal upthrusting along this line of faults has caused the gravity high (see section 1.5).

To the north of the Rudall River area the structural style of the upper Yeneena Group differs somewhat from that of the lower Yeneena Group. Folding is essentially concentric in form, which contrasts with the similar fold style of the lower Yeneena Group. Axial plane cleavage is seldom seen, except in a few specimens of unweathered core at Telfer (section 5.3.1). Two clastic sequences (the combined Malu and Telfer Formations, and the Wilki Quartzite) and two carbonate sequences (the Isdell and Puntapunta Formations) are interlayered, and the greater competency of the former sequences compared with that of the latter has resulted in the structural pattern shown on Map 1. Homoclinal ridges and discrete folds of the clastic formations are surrounded by extensive areas of less regularly deformed carbonate. The structure of this area is described in more detail below.

FIGURE 5.1 – STRUCTURAL CROSS SECTIONS OF THE TELFER REGION (MAP 1)



KEY AS ON MAP 1



5.2.2 THE TELFER REGION (MAP 1)

In addition to structural differences between this region and the Rudall River area to the south, there is variation of structural style within the Telfer region. Three structural zones are recognised, a southwest zone, a central zone and a northeast zone. The southwest zone includes the area of the Trotman Hills and the Karakutikati Range, the central zone extends from this range to the Thomson East Prospect, and the northeast zone includes the more complexly deformed area of the Malu Hills and the Wilki Range (Fig. 3.1).

The southwest structural zone comprises tight and isoclinal folds (using the terminology of Fleuty, 1964) which have horizontal fold axes and steeply dipping fold limbs. Bedding dip is commonly vertical or slightly overturned. A major anticline occurs in the Isdell Formation to the southwest of the Karakutikati Range (Fig. 3.1), the northeastern limb of which is formed by this range. Numerous small scale faults cut the Karakutikati Range (Map 1) and form a conjugate set orientated at high angles to the trend of the major fold axis. Two large scale faults cut the southwestern limb of this anticline and truncate the outcrop of Malu Formation at the Trotman Hills (Fig. 3.1). Along one of these faults to the east of the Trotman Hills occurs a large outcrop of auriferous quartz and silicifed dolomite, forming part of Newmont's "Grace Prospect" (see Chapter 8).

In the central structural zone the folds generally have shallower limb dips than in the southwestern zone; they are either open or gentle (Fleuty, 1964) in cross section, and characteristically have plunging axes. The most gently plunging fold occurs just north of the Karakutikati Range, where the Telfer Formation outcrops in the long northwest plunging nose of the Connaughton Prospect fold (Fig. 3.1). Farther north and east fold plunges are steeper; for example, at the western end of the Telfer Dome the plunge is 41°. Both northwest and southeast plunging fold axes occur, and structures such as Telfer Dome and Thomson East Prospect (Fig. 3.1) are elongate domes or periclines.

Several faults occur in the central structural zone (Map 1), the largest trending northwest-southeast along the northeastern side of the Wilki Quartzite outcrop south of Tim's Prospect (Fig. 3.1). This fault has a downthrow to the southwest (Fig. 5.1) and has a maximum throw of at least 1500 m at the northwestern end. The throw is less souheastwards, and the fault apparently dies out near the O'Callaghan Prospect (Fig. 3.1) where en echelon lenses of quartz at the surface (Chapter 8) may be the result of left lateral strike slip movement along several small faults which fan off the end of the major fault (Map 1). The fault type is uncertain as the attitude of the fault plane is unknown, but its presence in an area of strong compressional deformation suggests a reverse or thrust fault (Fig. 5.1).

Another fault in the central structural zone whose sense of movement can be deduced is the north-south trending fault to the east of Fallow's Field Prospect (Fig. 3.1). Left lateral (sinistral) strike-slip movement is suggested by the displacement of ridges of similar lithology in the Puntapunta Formation. More severe compression has occurred on the west side of the fault than on the east, as is evident by the tight folds at Fallow's Field Prospect (Map 1). Such strike-slip movement is consistent with a principle compressive stress directed northeast-southwest, orthogonal to the fold axes.

In the northeast structural zone the fold pattern in the Malu Hills area is more complex than in the central and southwest zones. The folds are more closely spaced and less elongate than farther south, and two synclinal cross folds occur, with axes at right angles to the major northwest-southeast structural trend (Map 1). To the northwest and southeast of the Malu Hills there are quartzite ridges with strike trends at varying angles to the major strike direction. The annular structure which includes the Wilki Range (Fig. 3.1) is composed of Wilki Quartzite. On the northwest, southwest and southeast sides of this structure the quartzite dips at high angles towards the centre of the ring, and on the northeast side a steeply dipping ridge of the same formation faces northeast. In the centre of the structure Chin and Hickman (1977) indicated two small outcrops of granite (Map 1), and Newmont geologists found isolated blocks of garnet-biotite-feldspar-quartz gneiss in the same area (D.S. Tyrwhitt, pers. comm. 1980). A circular granite pluton is thus thought to occupy the centre of the structure, the forceful intrusion of which is suggested to have caused the anomalous strike trends (see section 5.3.2). The arcuate arrangement of quartzite ridges to the northwest of the Malu Hills (Map 1) may have been caused by a similar mechanism, the forceful intrusion of the Mt. Crofton Granite.

5.2.3 STRUCTURE OF THE TELFER DOME

The Telfer Dome has been studied by the writer in considerably more detail than other structural units throughout the region. Apart from the obvious importance of this structure to the study of the Telfer gold deposits, it provides data on the type of deformation that typifies the central structural zone of the Telfer region,

Folds

The outline of the dome is defined by the outcrop of the Camp Sandstone Member (Map 1 and Fig. 5.2), within the confines of which are two subsidiary domes named Main Dome and West Dome (Plate 5.1). Main Dome is a relatively simple asymmetrically dipping structure, outlined by the Rim Sandstone Member. In plan it is a fairly flat oval, 3 km long and 800 m wide, with more pronounced curvature on the northeast side than on the southwest. Bedding dip on the northeastern flank of Main Dome steepens from 30° at the surface to about 55° at a depth of 700 m. On the southwest side of the dome the dip is a more uniform 25°. The axial plane dips steeply to the southwest.

The West Dome structure is more complex than Main Dome, having a central core of Rim Sandstone flanked by fault zones (Fig. 5.2), and a second anticlinal axis on the southwest. Limb dips are generally steeper than at Main Dome, with dips of about 35° on the northeast and about 50° on the southwest of the central part of the structure.

The fold style is broadly concentric in form (Fig. 5.2B), with movement along some bedding planes having taken place. In some places

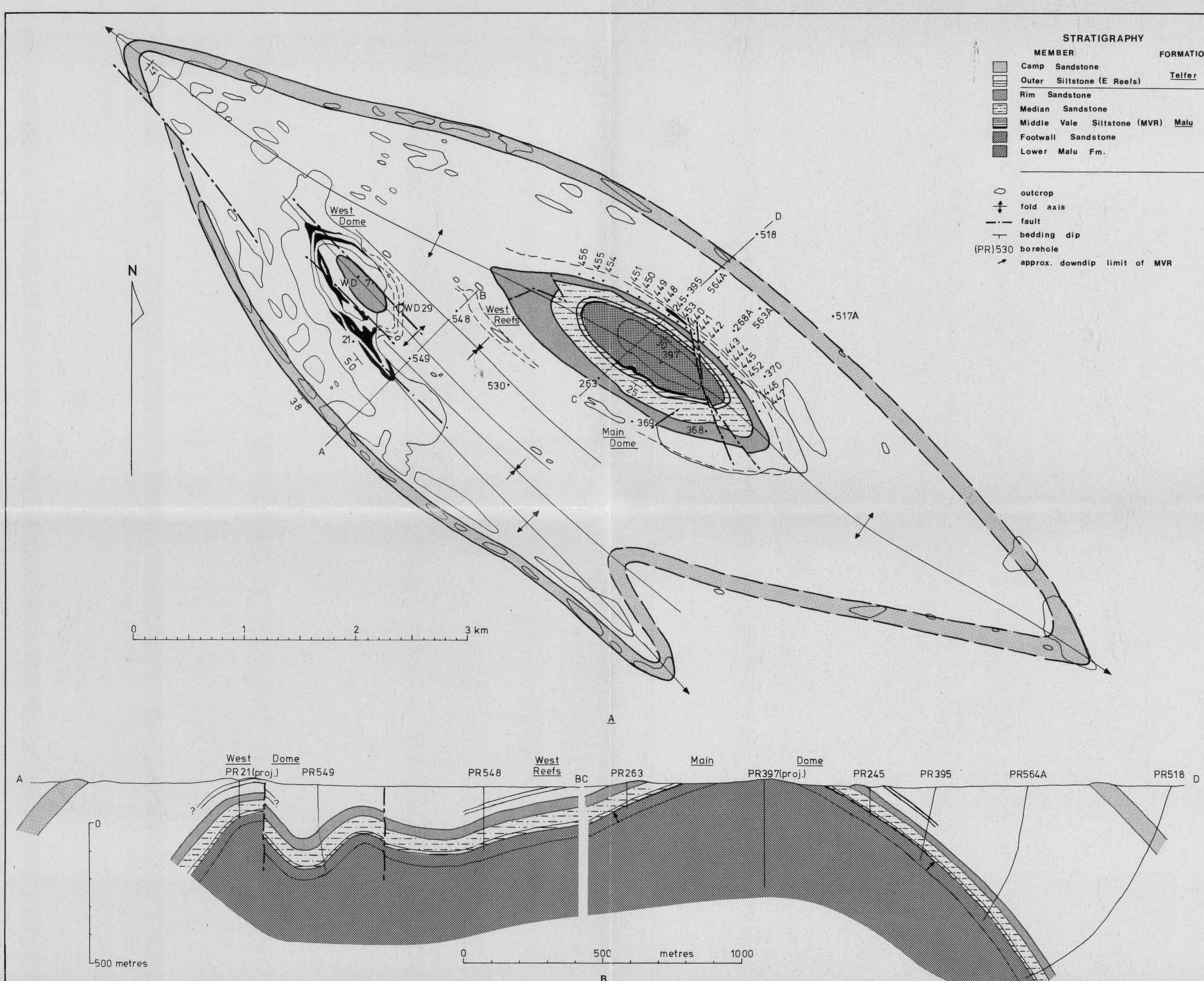


Figure 5.2 – A. Geological map, B. Cross section of the Telfer Dome.

this can be demonstrated where well developed slickensides occur on thin quartz veins which lie between bedding planes (as was also noted by Capill, 1977). This incidentally also demonstrates that the quartz veins pre-date folding. Some movement occurred at a low angle to the bedding (see below), indicating that the folds are not purely concentric in form, but it is important to note that the structures are not of similar fold style, where much of the movement is accommodated along axial plane cleavage. A further departure from purely concentric fold style is the fact that thickness variations of some sedimentary units occur across the Telfer Dome (see below). However, these variations are not entirely symmetrical with the folding, and they may be a sedimentation or compaction feature, rather than being caused by thickening and thinning during folding. This is discussed in more detail below.

Cleavage

Cleavage is not a prominent feature of the deformation at Telfer, but in outcrops of weathered mudstone around the dome shaly partings parallel to the local bedding dip occur. Cleavage is best seen in some thin sections of unweathered mudstone, where the dominant cleavage orientation is parallel, or at an angle of less than about 15°, to the bedding (see Plate 7.4D). The cleavage is caused by the alignment of sericite laths, and morphologically it can be classified as fine continuous cleavage (Powell, 1979). It is best developed in claystone beds directly beneath sandstone beds. The alignment of the cleavage at low angles to the bedding indicates that it formed in response to shear stress parallel, or almost parallel, to the bedding, and therefore formed during folding. De Sitter (1964, p.169) described such cleavage as concentric cleavage, which is found in folds with a basically concentric form.

Axial plane cleavage is rarely found at Telfer, and when present it is subordinate to the concentric cleavage. In a few thin sections of unweathered mudstone a spaced crenulation cleavage has been developed, the concentric cleavage being crenulated by axial plane cleavage. This indicates that folding was initially concentric in form, but that as deformation progressed incipient movement parallel to the axial plane occurred, the fold style becoming more similar in form.

An important structural feature which is associated with the concentric cleavage in some mineralised rocks at Telfer is the development of pressure fringes (Spry, 1979, p.240) on either side of pyrite crystals. This is described further in Chapter 7, but briefly, the pressure fringes are aligned parallel to the cleavage, indicating that the pyrite pre-dates the deformation (Plate 7.4D).

Faults

Several faults cut the Telfer Dome (Fig. 5.2A) and were probably formed at the same time as folding. At the northwestern nose of Main Dome reverse faults occur, which downthrow the crest of the structure (Fig. 5.2A and Plate 3.7E). This is typical of concentric folds where a room problem occurs at depth in crestal positions; that is, deformation cannot be entirely accommodated by flexural slip along bedding planes, and hence faulting results. Towards the southeastern end of Main Dome a narrow graben cuts the anticline at an angle of about 50° to the fold axis. This has been well exposed in Pit 1, where the Rim Sandstone Member is clearly downfaulted (Plate 3.7D). the main orebodies on the southeastern flank of the fold are also downfaulted in this graben (see Chapter 6). This transverse tensional faulting is consistent with the doubly plunging character of the Main Dome fold, where tensional stresses directed along the fold axis would be expected. It is therefore thought that this graben developed synchronously with the folding.

At West Dome three high angle fault zones occur (Fig. 5.2). These trend parallel to the fold axes and all are downthrown to the northeast.

Thickness variations across the Telfer Dome

During the course of the writer's core logging at Telfer (Chapter 6) it became apparent that some of the sedimentary units vary in thickness across the dome. Sufficient boreholes exist, which penetrate the succession from the top of the Rim Sandstone to the base of the Middle Vale Siltstone, to allow isopach maps of the Rim Sandstone and the Median Sandstone plus Middle Vale Siltstone to be constructed (Figs. 5.3A and B). True thicknesses were used in these constructions, based on downhole thicknesses and angle to bedding measurements. In many cases sedimentary unit boundaries were determined by the examination of drill core, but in some cases drill cores of the complete sequence do not exist, and reliance had to be placed on logged depths from drill cuttings. The reliability of the data is shown in the two figures.

For both sedimentary units a similar isopach pattern exists (Fig. 5.3) with the thickest sequences being found between Main Dome and West Dome, and the thinnest sequences occurring in the structurally highest part of the northeast flank of Main Dome. Although data on the southwest side of Main Dome are sparse, the isopachs are interpreted to swing across the dominant structural trend to form a broad trough which "plunges" to the northwest.

The thickness variations are considerable. In the case of the Middle Vale Siltstone and Median Sandstone Members the thickest sequences are about 50% greater than the thinnest sequences, and for the Rim Sandstone Member the thickest intersections are 100% greater than the thinnest. For both stratigraphic sequences similar thicknesses occur in areas to the southwest of West Dome and to the northeast of Main Dome. It is of interest, and possibly very significant, that the most extensive development of ores occurs on the northeast flank of Main Dome (Figs. 6.2 and 6.6), more or less coincident with the thinnest stratigraphic sequences (Chapter 10).

Interpretation of these thickness variations is not straightforward, as several factors could have played a part in producing the phenomenon. It is suggested, however, that thickening and thinning during folding are not the major cause of the variations. It has been demonstrated above that most features of the folding, faulting and cleavage formation are consistent with essentially concentric folding of the Telfer Dome, with shear stresses directed parallel, or at low angles, to the bedding. It would therefore be unlikely that such

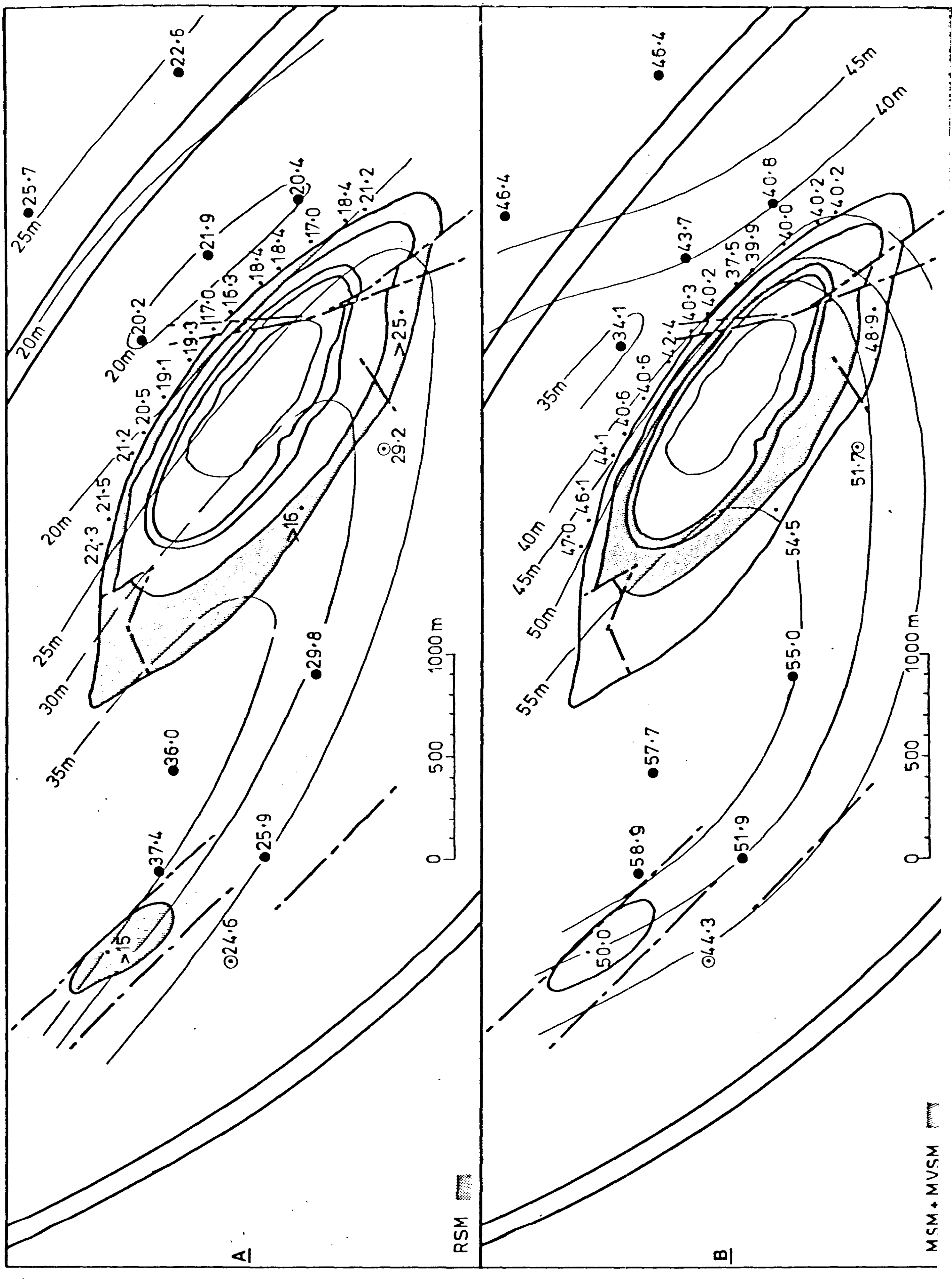


Figure 5.3 - Isopach maps of A. the Rim Sandstone Member, and B. the Median Sandstone plus Middle Vale Siltstone Members at Telfer. Large dots indicate bore holes where all boundaries of these units are cored, dots and circles indicate holes where the top of the RSM is not cored, and small dots indicate holes where neither the top nor bottom of the RSM is cored.

pronounced thickness variations would be caused by this style of deformation. In addition, the isopach patterns are not symmetrical with the folds (thickening in crests and troughs would be expected if the folding was of similar style). However, the thickest sequences do occur near the synclinal axis between Main Dome and West Dome, and in addition, axial plane cleavage is seen in some samples of the Rim Sandstone from one of the drill cores in this area (PR548 - Fig. 5.2). These features suggest that a minor portion of the thickening in the syncline may be due to stretching parallel to the axial plane during folding.

Evidence of a pre-folding variation in the thickness of the Rim Sandstone Member can be derived from a comparison of cored sequences of this unit across the structure. It would appear that most variation occurs in the upper half of the stratigraphic unit (Fig. 5.4). Individual sandstone beds are thinner, and the sand:shale ratio is lower, on the northeast flank of Main Dome compared with the area between West and Main Domes. This would suggest that more sand was initially deposited in the latter area than in the former, and that the thickness variations may therefore have a syn-depositional cause. Two possible syn-depositional causes must be considered, a purely sedimentological control and a tectono-sedimentary control.

A purely sedimentological control would appear unlikely. It has been demonstrated in Chapters 3 and 4 that the sequences under discussion are composed of sedimentary facies that were deposited by various types of sediment mass flows. These flows were directed to the north or northeast, at right angles to the trend of sediment thickening. Any localised thickness variations in such mass flow deposits might be expected to result from channelling or the growth of small sediment lobes on a depositional fan. In both cases the elongation of a locally thick sequence would be expected to parallel the palaeocurrent direction rather than be orthogonal to it.

However, it is possible that faulting during deposition of the sediments caused differential subsidence at the sedimentation surface, with the greatest subsidence occurring in the region of the syncline between Main and West Dome. If this were the case then a narrow

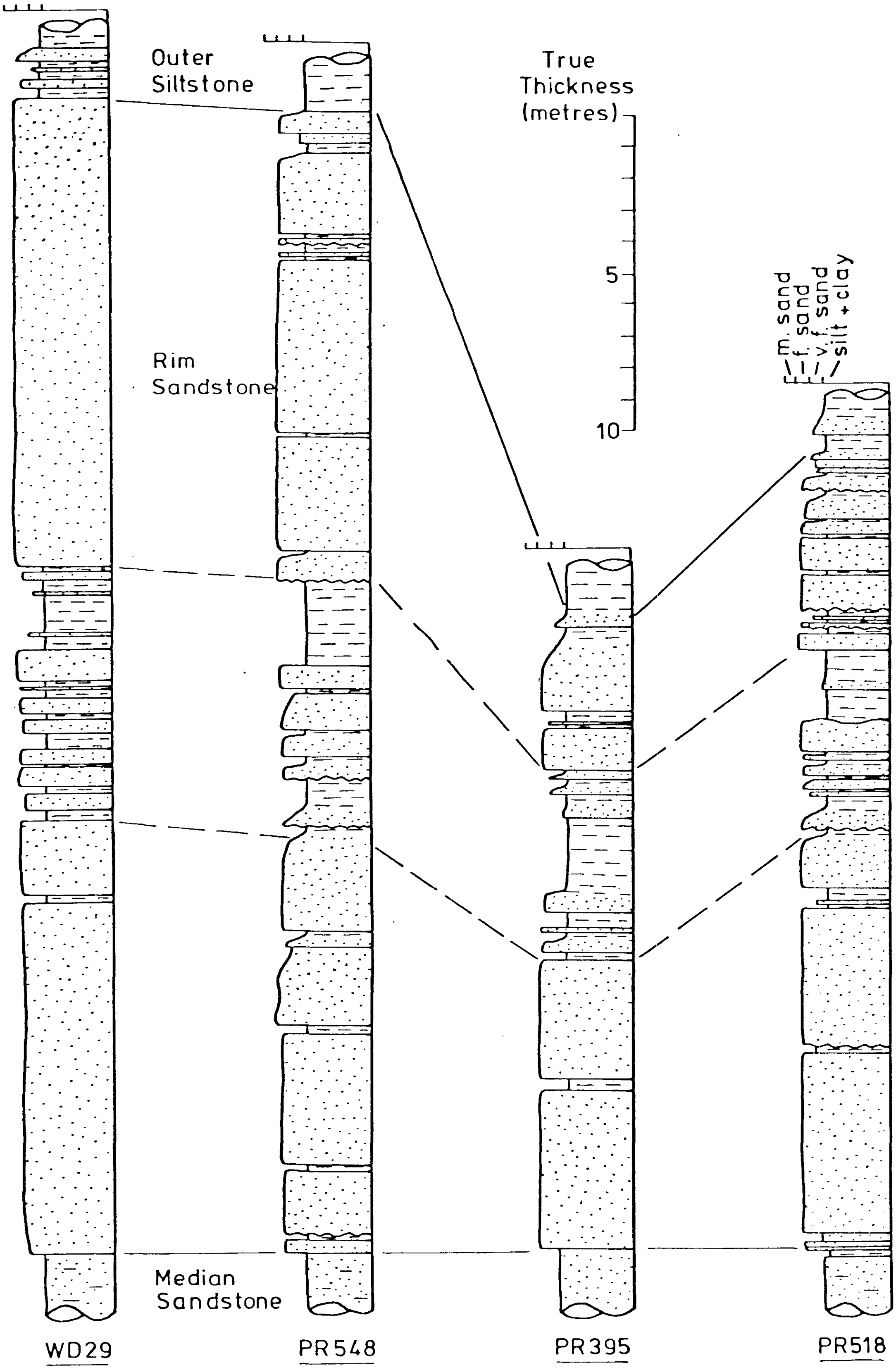


Figure 5.4 - Variations of thickness and stratigraphy of the Rim Sandstone Member in cores from across the Telfer Dome. See Figure 5.2 for bore hole locations.

graben can be postulated at depth, which trended northwest-southeast, and was active during sedimentation as shown schematically in Figure 5.5.

A locally thick pre-folding sequence may have acted as a focus for fold formation, the depositional basin being inverted during deformation to an antiform. That is, the Telfer Dome may be a consequence of a locally thick sedimentary sequence, rather than the thickness variations being a consequence of the folding. That such a process can operate, at least on a larger scale, has been demonstrated by Kingma (1958), who used the term "eversion" to describe uplift of a sedimentary basin in New Zealand.

5.3 METAMORPHISM AND GRANITE INTRUSION

Two metamorphic events can be recognised in the Yeneena Group of the Telfer region, a period of low grade regional metamorphism, and a later localised thermal metamorphic event caused by the intrusion of granite.

5.3.1 REGIONAL METAMORPHISM

The regional metamorphic grade of both the lower and upper Yeneena Group can be classified as lowest grade greenschist facies (as defined by Turner and Verhoogen, 1960, p.531). As pointed out by Chin and Hickman (1977) the metamorphism has been caused more by directional pressure than by elevated temperature, the most extensive recrystallisation occurring in the most deformed areas.

Few specimens of the lower Yeneena Group have been examined petrographically by the writer (Chapter 2), and therefore little comment can be made about the metamorphism of these rocks. However, the Coolbro Sandstone has suffered extensive recrystallisation, with quartz, muscovite and feldspar being the typical mineral assemblage. The Choorun Formation is less deformed and less recrystallised than the Coolbro Sandstone, and its mineral assemblage of quartz, muscovite, K-feldspar and plagioclase is essentially of detrital origin. The shale of the Broadhurst Formation typically contains

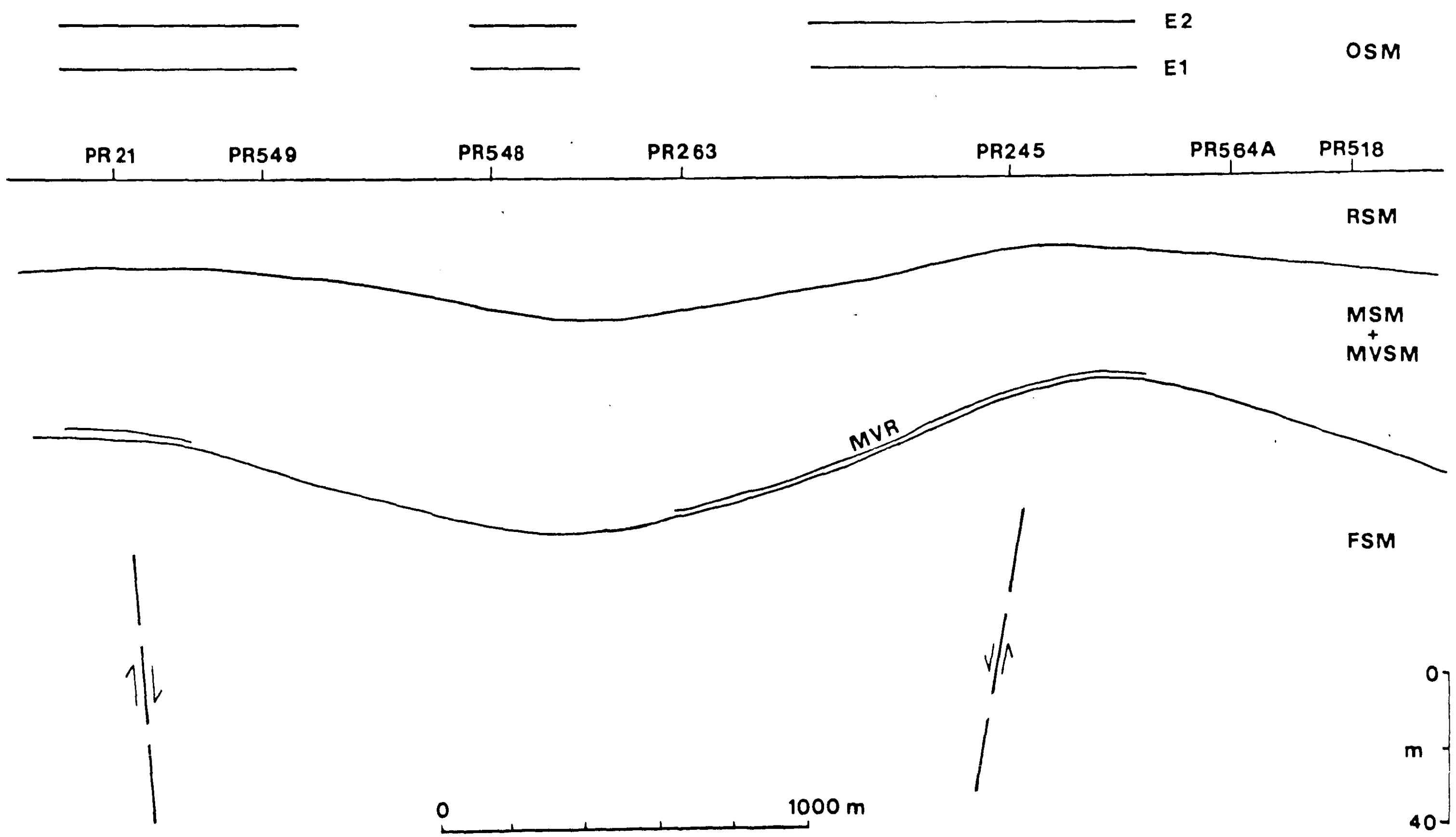


Figure 5.5 - Palinspastic cross section of the Telfer Dome (same line as in Fig. 5.2B) at the time of deposition of the lower Outer Siltstone Member. Thickness variations (Fig. 5.3) in the Rim Sandstone Member (RSM) and Median Sandstone plus Middle Vale Siltstone Members (MSM+MVSM), and possibly also the localisation of syngenetic mineralisation (Chapter 10), may have been caused by faulting at depth. Vertical exaggeration x 10.

quartz, sericite, feldspar, carbonate and kaolinite, with chlorite also occurring in some specimens.

The mineralogy of typical unweathered and partially recrystallised siliclastic rocks at Telfer is quartz, sericite, albite, carbonate and minor K-feldspar. Chlorite is occasionally present, and biotite also occurs in some of the deeper rocks on the northeast flank of Main Dome (see below). Larger quartz grains have generally not been recrystallised, although grain boundaries are commonly ragged due to incipient dissolution. However, the matrix of sandstones and the bulk of finer grained sediments have been reconstituted. Sericite has formed from clay minerals, authigenic crystals of carbonate have grown, and feldspar has commonly been albitised, implying sodium metasomatism. Calcium released from albitised plagioclase probably facilitated the growth of carbonate.

At Thomson and Thomson East Prospects northwest of Telfer (Fig. 3.1) the metamorphic grade is higher. At Thomson Prospect carbonate-rich siliclastic rocks have a pronounced schistosity, with a typical mineral assemblage of quartz, calcite, biotite and plagioclase. At Thomson East Prospect the mineral assemblage is similar, although with less carbonate, but the foliation is slight, with biotite in particular being unorientated. The biotite is probably a result of thermal metamorphism (see below) which was superimposed on the regional metamorphism. In no area is there clear evidence of the biotite subfacies of greenschist regional metamorphism.

Carbonate rocks of the upper Yeneena Group are commonly extremely fine grained and generally non-recrystallised. However, carbonate beds within the Telfer and Malu Formations at Telfer are coarser grained (equivalent to very fine sand grade) and commonly have a pronounced foliation parallel to the bedding. This is probably due to locally greater confining pressure acting on carbonate beds within the more competent siliclastic rocks than occurred in thick sequences of less competent carbonates. In the Isdell Formation south of Telfer recrystallisation of the carbonate rocks has occurred locally, where mineralised veins have intruded the sequence, and sulphide minerals and tourmaline occur in addition to carbonates (see Chapter 8). Epidote also occurs as a metamorphic mineral in some specimens of the Isdell Formation.

5.3.2 GRANITE INTRUSION AND THERMAL METAMORPHISM

In the Telfer region two granite plutons have been intruded into the upper Yeneena Group (Fig. 3.1). In addition, several granite bodies occur northeast of this area (Chin and Hickman, 1977). Exposure of the eastern pluton is minimal (Map 1), but structural effects caused by its emplacement (section 5.2.2) and recrystallisation in a metamorphic aureole (section 5.3.2) are readily apparent. The western pluton, termed the Mt. Crofton Granite by Chin and Hickman (1977), is well exposed as several prominent hills in the area between Mt. Crofton and the Lamil Hills (Map 1). A description of this pluton, based mainly on previous studies, is given below, and it is further considered in relation to the source of the mineralisation at Telfer in Chapter 10.

Mt. Crofton Granite

The Mt. Crofton Granite was described as coarse biotite granite by Trendall (1974), who gave a detailed petrographic description of material used for age dating. Quartz composes about 30% of the rock, feldspar (predominantly in alkali variety) makes up about 60%, and biotite forms most of the remainder. Very minor amounts of sphene, green amphibole, apatite and opaque minerals are commonly associated with the biotite. The description of material collected and examined by Chin and Hickman (1977) differs slightly from that of Trendall, as these authors recorded medium to coarse grained equigranular even-textured biotite adamellite, with some granite. The proportion of biotite was noted to be variable, and is only a minor accessory mineral in some areas, other accessory minerals including epidote, muscovite, magnetite, sphene, zircon, apatite, chlorite and fluorite. Chin and Hickman also presented a single chemical analysis of granite, which is reproduced as Table 5.1. Chemically the rock is very similar to the biotite alkali granite of Nockolds (1954, p.1012).

	%	ppm	
SiO_2	76.6	Ba	255
Al_2O_3	13.3	Li	19
TiO_2	0.10	U	10
Fe_2O_3	0.4	Zr	85
FeO	0.77		
MnO	0.03		
MgO	0.0		
CaO	0.86		
Na_2O	3.30		
K_2O	5.2		
P_2O_5	0.08		
CO_2	0.12		
H_2O^+	0.36		
H_2O^-	0.11		
Total	101.2		

Table 5.1. Chemical analysis of a sample of the Mount Crofton granite (location 21°33'05"S, 121°57'45"E). From Chin and Hickman (1977).

Intrusive contacts of granite against quartzite are well exposed at the edges of outcrops of the Wilki Quartzite south of the Telfer access road (Map 1). Where examined by the writer the contact is steeply dipping, and in detail is irregular due to protrubences of granite into the vertically dipping quartzite. Xenoliths of quartzite occur in places in the granite, and veins of aplite extend for short distances into the quartzite. A zone about 2 m wide of fine grained granite at the contact marks the chilled margin of the pluton. A few quartz veins occur in the quartzite, but no mineralised veins have been recorded in the granite itself.

The age of the Mt. Crofton Granite was determined by Trendall (1974), using Rb-Sr methods. Seven analyses of the coarse biotite granite and one of a finer granite vein defined an isochron of 614 ± 2 m.y. When an additional analysis of an aplite vein is included the isochron becomes 594 ± 2 m.y. Biotites from four of these samples were analysed for Rb and Sr by de Laeter et al. (1977), who determined a mean age of 580 m.y. The Rb and Sr chemistry also indicated that the granite cannot have been derived anatectically from an underlying extension of the Archaean Pilbara Block (Trendall, 1974; de Laeter et al., 1977). This is consistent with the deduction in Chapter 1 that the Telfer region is not underlain by such an extension of the Pilbara Block. Trendall further suggested that the most likely genetic model for the granite magma is differentiation from a mantle source no more than about 20 m.y. before its emplacement. Chin and Hickman (1977) pointed out that the strongly fractionated chemical composition of the granite suggests its emplacement at a relatively high crustal level, which is consistent with the low grade of regional metamorphism of the Yeneena Group.

Thermal metamorphism

The following observations of the thermal metamorphic effects due to the intrusion of granite result from the writer's field work and petrographic examination of the Yeneena Group.

At the Wilki Range east of Telfer an increase in the degree of recrystallisation of the Wilki Quartzite occurs northeastwards across the 1.5 km width of outcrop. On the southwest side of the range, the basal units of the formation are only partially recrystallised and resemble sandstone of the Malu Formation at Telfer. Thin sections of specimens collected across the range reveal a gradual increase in recrystallisation to a point where the sedimentary fabric has been completely destroyed, and a meta-quartzite texture with interlocking irregular shaped quartz grains occurs. Similar coarsely recrystallised meta-quartzite occurs adjacent to the Mt. Crofton Granite northwest of Telfer. At an outcrop northwest of the McKay Prospect (Map 1) an increase in metamorphic grade is again apparent on the granite side of the quartzite, where muscovite crystals occur between interlocking quartz grains, and very minor biotite and chlorite also occur.

The biotite that occurs extensively in rocks at Thomson and Thomson East Prospects is also thought to be due to thermal metamorphic effects of granite emplacement. At the latter prospect the biotite is unorientated, suggesting little directional stress during its formation. Aggregates of biotite in some samples have produced a spotted hornfels texture. At Thomson Prospect, however, the biotite is aligned parallel to a strong foliation. this can be explained by mimetic recrystallisation of orientated muscovite that had formed during regional metamorphism. These areas are between 8 km and 10 km from the margin of the Mt. Crofton Granite, beyond the probable limits of the aureole of this relatively small pluton. It is therefore probable that granite underlies the area of the two prospects, and it is possible that the two granite plutons are connected at depth.

At Telfer a restricted thermal metamorphic effect is also apparent. Biotite occurs in carbonate-bearing siliclastic sediments in boreholes PR517 and PR518 on the northeast flank of the Telfer Dome (Fig. 5.2), composing up to about 10% of the rocks. The mineral is not uniformly distributed through the sediments, however, being in part dependent on bed composition, but it occurs intermittently at depths below at least 340 m. The biotite grains are unorientated, even in samples with a pronounced cleavage, and are commonly roughly equidimensional with ragged edges. The grain size varies with that of the sediment, with the biotite grains generally being slightly larger than the siliclastic particles. The biotite is deduced to be a thermal metamorphic mineral which post dates regional metamorphism and cleavage formation. Chlorite commonly accompanies biotite, but in most specimens the chlorite is aligned parallel to the cleavage and is closely associated with sericite, indicating that it is a regional metamorphic mineral. However, in a few specimens the chlorite is coarser grained and more similar in form to the biotite, suggesting that the mineral was in part reconstituted during thermal metamorphism.

Biotite does not occur in other deep boreholes at Telfer, such as PR397, PR548, PR549 and PR530 (Fig. 5.2), whereas chlorite does occur elsewhere (e.g. Table 3.4). It is therefore likely that a granitic body occurs at depth to the northeast of the Telfer Dome, which adds further credence to the suggestion of a connection at depth of the two granite plutons. There is no evidence to suggest that a granitic body directly underlies the Telfer Dome, or that the intrusion of granite has affected the structure of the dome.

5.4 PHANEROZOIC GEOLOGY

A major unconformity occurs between the folded Yeneena Group and the Permian Paterson Formation in the Telfer region. There are no lower Palaeozoic sedimentary units in the area, but silcrete, which forms a capping on many Precambrian outcrops, may have formed during this time interval (see below). Yeates and Towner (1978) suggested that a long period of erosion and peneplanation followed the deformation of the latest Precambrian sediments, and the intrusion of granite, in this area of Western Australia. They further suggested that the initial transgression of the southern Canning Basin, to the east of Telfer, occurred in the Early Ordovician. The Canning Basin succession includes Ordovician, Silurian and Devonian strata, and it would seem probable to the writer that at times during this interval deposition may have onlapped as far westwards as Telfer, which is about 50 km from the present margin of the Canning Basin. There is no tangible evidence for this, however, as any lower Palaeozoic sediments that may have been deposited were removed by further extensive erosion prior to the deposition of the Paterson Formation.

5.4.1 PATERSON FORMATION

This flat-lying formation is of glacial and fluvioglacial origin. It occurs as prominent scattered remnants of a once continuous sheet of sandstone and conglomerate that mantled the peneplained Yeneena Group. The age of the formation in the Yarrie map sheet area to the northwest of Telfer (Fig. 1.3) has been deduced from spores and pollen to be Early Permian (Hickman and Chin, 1977). The Formation has been described by these authors and only a brief account is given here.

The basal contact of the formation is exposed in places (Plate 3.7G), where the lowest unit, consisting of very poorly sorted unstratified rubbly conglomerate, rests on various units of the Yeneena Group. At one locality 16 km west-northwest of Telfer, glacial striations were recorded by the writer on the unconformity surface (Plate 3.7F). Tillite, which includes glacially striated and faceted boulders, was recorded by Chin and Hickman (1977) within the lower part of the formation. These authors noted that clasts in both the tillite and conglomerate ranged in size from small pebbles to boulders 3 m in diameter, and included quartzite, granitoid rocks and, rarely, basalt, chert, gabbro and dolomite. In addition, clasts of silcrete, seemingly identical to underlying silcrete developed on the Yeneena Group, were recorded by Turner and McKelvey (1981).

Overlying the basal conglomeratic unit are medium to coarse grained quartzose sandstones, which are in places cross bedded and are probably of fluvioglacial origin (Chin and Hickman, op. cit.). Locally these sandstones have moderate dips (e.g. 35°-60° at a small outcrop 8 km west of Telfer), suggesting slumping of unstable sediments caused by the melting of ice. A similar origin was suggested by Chin and Hickman for stratabound zones of folded and faulted sandstones in the northern part of the Paterson Range.

The direction of Permian ice movement in this area of Western Australia was generally to the north. The striations shown in Plate 3.7G indicate ice movement towards 020°, and other striations in the Nullagine map area west of Telfer also indicate northerly ice movement

(Hickman, 1978, p.15). In addition, Chin et al. (1980) suggested northwesterly ice movement from the orientation of U-shaped valleys in the Rudall map area.

5.4.2 SILCRETE

A capping of silcrete has been developed in the Telfer region on many outcrops of dolomite and dolomitic sandstones of the Puntapunta Formation. The silcrete is pale grey and chert-like in appearance, consisting of micro-crystalline silica which formed by a combination of carbonate leaching and silica replacement of the Precambrian strata. These surface deposits effectively mask the nature of the underlying sediments, but the original attitude of the bedding is commonly preserved. The silcrete has been discussed in some detail by Turner and McKelvey (1981), who suggested that silicification took place during the Palaeozoic. This is based on the facts that the Permian Paterson Formation overlies silcrete in some areas (Plate 3.7G), and that the basal Permian sediments contain some clasts of silcrete, as noted above. This conclusion is at variance with the commonly held belief that the silcrete is of Tertiary age (e.g. Chin and Hickman, 1977).

The probability of a pre-Permian origin of the silcrete implies an extensive period of weathering of a peneplain in the Telfer region in Palaeozoic times. The present land surface around Telfer is in part probably an exhumed remnant of the Palaeozoic peneplain. For example, the crests of both Main and West Domes are relatively flat, and are at about the same elevation above the surrounding sand plain (approximately 30 m). The extensive outcrops of the Puntapunta Formation immediately north of the northwestern nose of the Telfer Dome are capped by silcrete, and are also at about the same elevation as Main and West Domes. The tops of all these hills are suggested here to approximate the exhumed pre-Permian peneplain. The significance of this with regard to the Telfer gold deposits is that extensive weathering and possible supergene enrichment of the ores (see Chapter 7) may have occurred during both the Palaeozoic and the Tertiary.

PLATE 5.1 - TELFER DOME

Vertical air photographs by Main Dome and West Dome, Telfer, prior to mining. See Figure 5.2 for geology. Discontinuous linear features are sand dunes; dark areas are rock outcrops. Photos by Kevron Ltd., Perth.

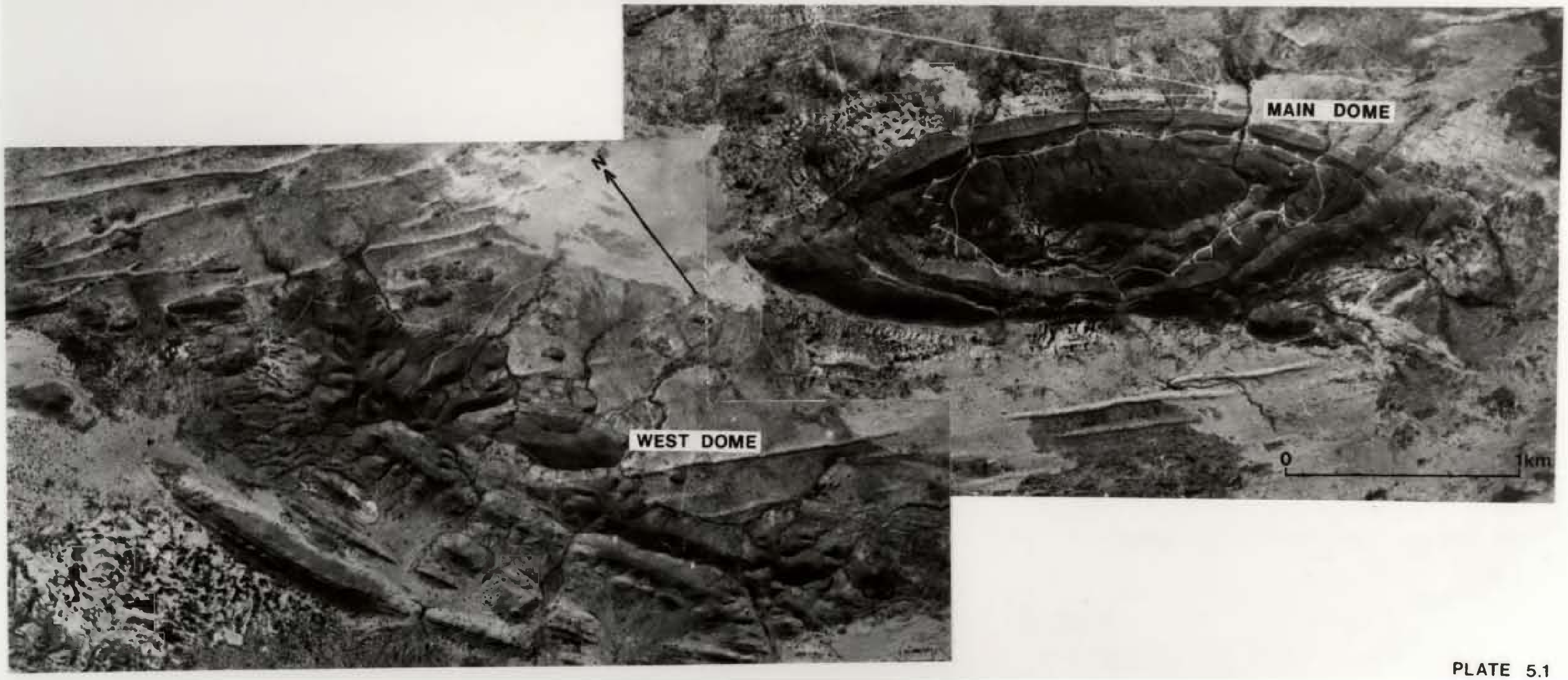


PLATE 5.1