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DEPARTMENT OF PRIMARY INDUSTRIES AND ENERGY

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Abstract

Detailed geological mapping of the North Shaw and Tambourah 1:100 000 map sheet areas of the Archaean Pilbara Craton in 1994-1996 has identified five unconformity-bound supracrustal assemblages and three generations of granitoid intrusions that range in age from the ca. 3515 Ma Coonterunnah succession to the 2760 Ma Fortescue Group. In the area southeast of the Strelley Granite, a previously unknown angular unconformity was observed to separate the base of the 3240 Ma Strelley succession and conformably overlying clastic rocks of the Gorge Creek Group from older mafic volcanic rocks that probably belong to the ca. 3460 Ma Warrawoona Group.

Three generations of structures were identified across the map areas. The first set, at ca. 3460 Ma, include structures related to doming of the older Coonterunnah succession and Warrawoona Group by the syn-volcanic emplacement of granitoid intrusions in the Carlindi Batholith, North Pole Dome, Shaw Batholith and possibly the Yule Batholith. A migmatitic gneissosity developed in these older plutonic rocks and rare folds in the Coonterunnah succession may be related to this event.

The second set of structures is related to the diapiric emplacement of a second main generation of granitoid rocks at ca. 3300-3240 Ma, which were intruded into, and amplified the pre-existing granitoid domes. Associated structures include tight to isoclinal folds of supracrustal rocks shed off the rising domes, zones of ultramafic tectonic schist and megabreccia in extensional detachment surfaces between diapirically-reactivated ductile middle crust underlain primarily by granitoid rocks, and brittle upper crust dominated by supracrustal rocks. Syn-kinematic amphibolite-facies metamorphic mineral assemblages formed in a kilometre-wide contact aureole around the Yule Batholith.

Superimposed on these structures is a set of discrete shear zones, faults, and tight, predominantly NE-trending folds within the N-S striking Central Pilbara Structural Corridor (CPSC). This third set of structures formed under greenschist-facies metamorphic conditions at ca. 2950 Ma, as a result of predominantly sinistral transpressional deformation., and occurred distinctly after doming. The western boundary of the CPSC is a curviplanar sinistral fault, whereas the eastern margin of the CPSC is formed by a set of en-echelon dextral fault strands. A set of tight, NE-trending folds in the central part of the CPSC formed in response to sinistral wrenching. Shearing caused the originally flat-lying Strelley granite laccolith and associated supracrustal rocks to be displaced laterally to the west and north, and to become tilted onto their side within the CPSC through ramping up the curved eastern boundary of the Yule Batholith. Coarse clastic rocks of the De Grey Group in the Lalla Rookh Basin are interpreted herein as having been deposited in a small foreland-style basin in front of the northerly travelling Strelley Granite, rather than in a sinistral pull-apart basin, as previously hypothesised.

Introduction

Field geological mapping was conducted in the North Shaw and Tambourah 1:100 000 sheet areas of the Pilbara Craton between June 20 and August 31, 1996 while on contract with AGSO (Fig. 1). This work followed on from previous mapping in this same area in 1994 and 1995 while at the University of Newcastle as an Australian Research Council (ARC) Postdoctoral Fellow, through support by ARC small grants, and the logistical field support of SIPA Resources (1994), and CRA Exploration Pty. Ltd. (1995). The principal areas mapped, and the year, are shown on Figure 2.

Mapping has confirmed the presence of *five* supracrustal successions, and *three* main episodes of granitoid magmatism coeval with the middle three supracrustal successions (Table 1). A brief description of the supracrustal successions and their regional extent will be presented, followed by a description of granitoid magmatic events and a more detailed explanation of the structural style associated with each of the three main events.

Table 1. Supracrustal successions and granitoid magmatic episodes in the North Shaw and Tambourah map sheets, Pilbara Craton.

Supracrustal successions	Granitoid magmatic events	Tectonic style
2760 Ma: Fortescue Gp.[1]	(present in other map sheets)	Rifting[10]
2950 Ma: De Grey Gp.[2]	2950 Ma: granitoids[6]	Regional sinistral shear[2]
3325 Ma Wyman & Kelly Fms. in E.[3], & 3240 Ma Strelley succession & Gorge Creek Gp. in W.[4]	3300 Ma domes in E.[7] & 3240 Ma Strelley Granite laccolith[4] & granitoids in the Yule Batholith in W.[8]	Granitoid diapirism and coeval basin formation during extension[11]
3460 Ma: Warrawoona Gp.[3]	3460 Ma: TTG granitoids[5,7-9]	Arc or mantle plume??
3515 Ma: Coonterunnah[5] succession	?	?

1. Arndt et al. (1991). 2. Van Kranendonk and Collins (in press). 3. Thorpe et al. (1992), McNaughton et al. (1993). 4. Unpublished U-Pb SHRIMP data, University of Western Australia. 5. Buick et al. (1995). 6. Bickle et al. (1989). 7. Williams & Collins (1990). 8. Unpublished U-Pb SHRIMP data, this report. 9. Bickle et al. (1983, 1993). 10. Blake (1993). 11. Van Kranendonk (1995).

In this report, stratigraphic terms are after Hickman (1983, 1990), Morant (1995; Strelley succession), and Buick et al. (1995; Coonterunnah succession). Lithostratigraphic and structural names are shown on Figure 1 and include three main subdivisions; the Western Domain, the Central Pilbara Structural Corridor, and the Eastern Domain (Van Kranendonk and Collins, in press). The Western Domain is comprised of little deformed rocks of the Yule and Carlindi batholiths, the Pilgangoora Belt, Pilgangoora Syncline, Strelley Belt, and Strelley Block (terms from Hickman, 1983; Wilhelmij, 1986; and Krapez, 1993). The Eastern Domain is also comprised of little deformed rocks, including the North Pole Dome and tilted greenstones around the northern margin of the Shaw Batholith. The Central Pilbara Structural Corridor is a N-S striking zone of tightly folded and variably strained rocks between the lower strain domains, affected throughout by late, sinistral strike-slip deformation (Krapez, 1984; Van Kranendonk and Collins, in press).

Stratigraphy

Coonterunnah succession

Discovered and described by Buick et al. (1995), this sequence of bimodal felsic and mafic volcanics and banded iron formation is tilted to steep dips around, and faces away from, the southern margin of the Carlindi Batholith (Fig. 1). These rocks are locally folded and faulted and contain a weak foliation that predate the overlying unconformity by the ?Warrawoona Group (see below). Folding style is consistent with the tilting of the sequence away from the core of the Carlindi Batholith during its emplacement (see below).

Warrawoona Group

A widely distributed, thick sequence of pillowed and massive, tholeiitic and high-Mg basalts and sills, felsic volcanics, and characteristic black replacement cherts after carbonate or thinly bedded mafic to felsic ash, known as the ca. 3460 Ma Warrawoona Group, occurs in the North Pole Dome and around the Shaw Batholith (Pidgeon, 1978; Buick et al., 1995). Similar rocks, though undated, probably also occur unconformably above the Coonterunnah succession along the northern limb of the Pilgangoora Syncline (R. Buick, U. Sydney, pers. comm., 1996), south of the Soanesville Syncline in the Western Shaw Belt, in the Tambina Complex west of the Shaw Batholith, and around the Tambourah Dome (Fig. 1: see Structural Geology, below). Little of the North Pole Dome was mapped, other than an area in the northwest adjacent to the Lalla Rookh Basin, where rocks of uncertain group affinity occur (see below).

Northwestern margin of the North Pole Dome

In this area, a felsic volcanic eruptive centre and associated volcaniclastic rocks were discovered, that are conformably underlain by heavily carbonate-altered and strongly deformed mafic-ultramafic volcanic rocks, and conformably overlain by well-preserved, little altered, high-Mg and tholeiitic pillow basalts (Fig. 3). The conformable contacts with overlying and underlying mafic volcanic rocks, the exclusively mafic volcanic origin for the overlying rocks, the along strike continuity of the felsic volcanic rocks with dated Warrawoona felsics, consistent facing directions of the folded stratigraphy, and structural considerations detailed below, all suggest that the area mapped is part of the Warrawoona Group of the North Pole Dome.

The oldest rocks in the map area are heavily carbonate-altered, pillowed to massive mafic volcanic rocks of both high-Mg and tholeiitic composition, interbedded cherts, and ultramafic rocks of uncertain extrusive or intrusive origin. Carbonate is extensively developed as both crystalline-scale alteration of mafic-ultramafic rocks and in massive, dark brown-weathering dykes (see below). Pillow facing directions are locally evident, though rare, due to extensive alteration and shearing.

The felsic volcanic rocks have a bimodal structural distribution. In the south, the felsic volcanic centre is generally shallow dipping and weakly deformed into a shallow, doubly-plunging syncline cored by eruptive breccias thick with felsic volcanic fragments and those of a syn-volcanic unit of red banded iron formation which is interbedded with the felsic volcanic rocks around the outer edge of the volcanic centre. An outcrop of eutaxitic rhyolitic ignimbrite with graded pumice lapilli indicates right way up for the core zone. North and northeast of the volcanic centre, steeply dipping to overturned, interbedded felsic tuffs, pummicite, and reworked (cross-bedded) and silicified (cherty) ash flows are deformed into a broad, east-facing, reclined to overturned antiformal fold (D3: Fig. 3, inset). The whole southeastern to eastern limb of the fold structure is overturned to the east, whereas the northern limb varies around 90° dips.

A red banded iron formation dips under the felsic breccias occupying the core of the volcanic centre along the south, in the northwest and northeast. Along the northern margin of the structure, the iron formation is dismembered by felsic volcanic breccia. The western end of the northeastern arm of the iron formation is deformed into a fold which curls south into the felsic volcanic centre, and is then cut off by undeformed felsic volcanic breccia. This fold and the dismemberment of the iron formation are both interpreted to be the result of explosive eruption of felsic volcanic magma from the core of the structure. In this region, extensive autobreccia is present, along with occasional beds of crystal lithic tuff which locally contains sheath-shaped flow folds. Diffuse zones throughout the core contain red jasper fragments in addition to those of felsic volcanic material. A 1 m diameter pipe of red jasper fragment breccia cuts felsic volcanic autobreccia in the central south part of the core zone.

A cross-section through the volcaniclastic sequence in the south shows finely bedded felsic tuffs at the base, capped by a white-and-red layered chemical chert (MBC on Fig. 3) that is lithologically similar to the Marble Bar chert further east, except that the former lacks the ubiquitous veins and sills of black replacement chert contained in the latter. Above the chert, the grain size of the tuffs becomes coarser, culminating in 1-2 m thick beds of red jasper breccia with ≤10 cm diameter fragments. Stratigraphically above this unit, the clast size in the breccias decreases again and passes up into cross-bedded, reworked ash flows interbedded with several fine-grained, silicified ash beds (cherts). Grain size and bedding thickness decreases to the top of the section, which is capped by a thick replacement chert unit after felsic ash. North of the section, the jasper breccias decrease in thickness and fragment size, and do not occur along the northern limb of the fold. This northerly decrease in fragment size suggests that both the eruptive source of the felsic volcaniclastic material, as well as the origin of the jasper was from the south, consistent with the eruptive textures and the presence of banded iron formation observed in the felsic volcanic centre.

Conformably overlying the felsic volcaniclastic rocks are relatively unaltered, weakly strained, pillowed and massive, high-Mg basalts, chert, and tholeiitic mafic pillowed volcanics. Pillow structures are well preserved such that facing directions are commonly evident (Fig. 3). The preserved cross-section of this sequence in the hinge of the fold structure shows that the tholeiitic volcanics were erupted through the felsic volcaniclastic sequence via a series of dykes, extruded out of an asymmetrical vent, and flowed down a paleoslope to the east, as indicated by progressively decreasing pillow size east of the vent (Fig. 3). Dismembered rafts of chert and high-Mg pillow basalt from the underlying sequence occur in the vent, and an erosional base to the tholeiitic flows is preserved to the east. The main feeder dyke to the vent is highly schistose, whereas coplanar dykes off the vent axis are fresh and undeformed, indicating the probability that a large volume of magma passed through the feeder dyke vent.

An unusual feature of this area is that the felsic and underlying mafic volcanic sequence is affected by a high degree of carbonate alteration, but that this alteration is not seen in the overlying sequence. In places, tuff breccias of the felsic volcaniclastic sequence are contained within an almost solid matrix of dolomite after felsic ash, and many dykes of dark brown weathering carbonate were discovered throughout the section. In addition, several distinct, elliptical areas of carbonate hosted breccia occur within felsic breccias in the felsic volcanic centre (Fig. 3). The carbonate matrix in these areas hosts both jasper and felsic volcanic fragments.

Initially thought to represent evidence of both intrusive and extrusive carbonatite magmatism coeval with the felsic volcanism, thin section petrography of carbonate-bearing rocks reveals replacement, rather than magmatic textures for the carbonate minerals, and a lack of commonly associated magmatic carbonatite minerals such as hornblende and olivine. The carbonate principally replaces the fine-grained felsic volcanic matrix in volcaniclastics. For example, in some finely-bedded ash tuffs, replacement of the felsic matrix by carbonate in mm-thick discordant veins was observed. Also, carbonate-hosted jasper breccias in the southern part of the volcaniclastic section pass along strike to the north into white, felsic volcaniclastic-hosted jasper breccias without carbonate. These features suggest a replacement origin for the carbonate. However, the restricted nature of the carbonate to a discrete stratigraphic level, a distinct correlation between the amount of carbonate with proximity to the felsic volcanic centre, and the presence of carbonate dykes suggests that the carbonate may have been introduced *during* felsic volcanism, possibly during volatile-charged episodes. A volatile-rich, gas-charged environment during felsic volcanism is supported by the extensive breccias and volcaniclastic deposits in this area.

Northern limb of the Pilgangoora Syncline

Steeply south-dipping and south-facing mafic volcanic rocks and interbedded chert lie unconformably atop the Coonterunnah succession south of the Carlindi Batholith across the thick, distinctive, Strelley Pool chert (Fig. 1: Buick et al., 1995). Although undated, the chert and overlying volcanic rocks have been correlated with the Towers Formation and upper part of the Warrawoona Group (Salgash Subgroup), respectively, to the east (Hickman, 1990). These rocks are similar to the upper part of the Warrawoona Group in the North Pole Dome in that they contain numerous chert interbeds (Buick et al., 1995), and dissimilar to the upper part of the Gorge Creek Group because they lack interbedded sedimentary rocks; thus a Warrawoona age is considered most likely for this succession.

Tambina Complex

This structurally dismembered zone contains mafic and ultramafic volcanic rocks and numerous, varied cherts (Boulter et al., 1987). Large-scale tight folds and faults have disrupted the stratigraphy, but nearly pristine pillow and other volcanic textures are widely preserved within kilometre-scale coherent panels on fold limbs. A steeply south-plunging synformal anticline, overturned to the north, is outlined by a thrust-bound sliver of Gorge Creek Group and unconformably overlying De Grey Group sedimentary rocks (see Fig. 1: Boulter et al., 1987).

Western Shaw Belt and Tambourah Dome

Wrapping around the Tambourah Dome is a sequence of massive, equigranular amphibolites and less common ultramafic schists, with rare igneous or volcanic textures such as ocelli. Interlayered with the amphibolites in the hinge of the fold north of the dome, are buff-weathering meta-quartz arenites which preserve coarse bedding despite extensive recrystallisation. The meta-sandstones are distinct from metamorphosed cherts, which typically have a blue-grey colour and sucrosic textures. Although distinct from the typical Warrawoona Group of the North Pole Dome and around the Shaw Batholith in having metasandstone beds interlayered with the mafic volcanic rocks (amphibolites), these rocks have been deformed by at least two main episodes of deformation and thus are likely to form part of the Warrawoona Group (see Structural Geology).

This relatively homogeneous sequence passes into a heterolithic assemblage in the eastern half of the Western Shaw Belt, across a high strain zone containing sheared quartz porphyry intrusives. Common felsic schists in this part of the belt are largely derived from the silicification of mafic rocks, though some may be of felsic volcanic origin. In the centre of the belt are strongly flattened, steeply dipping sandstones and pebble conglomerates. Facing directions seen in two outcrops give different directions, indicating tight folding. Significantly, however, pillowed volcanic rocks adjacent to the Shaw Batholith face east, towards the batholith, and are overturned. This observation, the lack of a matching panel of homogeneous amphibolites adjacent to the Shaw Batholith as occurs adjacent to the Tambourah Dome, and the presence of a monoclinal panel of the Gorge Creek Group in the southern part of the belt, together indicate the Western Shaw Belt does *not* represent a simplistic greenstone synform, but rather, an east-facing monoclinal panel (though it be strongly sheared and tightly folded).

Strelley succession and the Gorge Creek Group

Strelley succession

West of the North Pole Dome and Shaw Batholith is the well-preserved, though folded and faulted, Strelley succession which comprises wacke, andesitic through dacitic to rhyolitic felsic volcanics, and an overlying unit of grey and white layered "marker" chert after felsic ash (Fig. 1: terminology after Morant, 1995; Vearncombe et al., 1995). The succession is intruded by a syn-volcanic granite laccolith, the Strelley Granite. Both the granite and the intermediate to felsic volcanic rocks have been dated as ca. 3240 ± 5 Ma by U-Pb SHRIMP dating (unpublished zircon data, U. Western Australia *in* Morant, 1995; R. Buick, pers. comm., 1996). Associated massive sulphide (Cu-Zn) deposits at the top of the felsic volcanics give a slightly older ^{207}Pb-^{206}Pb age of 3257 +8/-6 Ma, suggesting contamination by older crust (Vearncombe et al., 1995). The sequence is underlain by mafic pillow basalts, ultramafic rocks and chert, which may either form a lower part of the same sequence, or belong to an older succession (Warrawoona Group?: see below). Overlying the sequence are the predominantly clastic rocks (plus mafic volcanics and gabbro sills) of the Gorge Creek Group. In this report, the Strelley succession refers only to the well-defined part of the sequence above and including the wacke, up to and including the marker chert.

The Strelley succession is best developed around the Strelley Granite, where it dips and youngs 60° eastward, is remarkably well preserved at very low metamorphic grade, and is host to numerous, massive Cu-Zn deposits (Fig. 1: Morant, 1995). North of the Strelley Granite, the succession and underlying pillowed high-Mg basalts are deformed into a drag fold associated with a regional sinistral fault. A west-facing panel of andesitic to felsic volcanics and the marker chert occurs west of the regional sinistral fault located immediately west of the granite (Fig. 1: see below). Through deconstruction of movement across the sinistral fault, this panel is seen to have been attached to the drag folded part of the sequence north of the granite.

South of the Strelley Granite, the Strelley succession and overlying Gorge Creek Group are folded into a tight, NE-plunging syncline, whose southern limb is strongly sheared and transposed to the northeast into a km-wide zone of ultramafic-hosted tectonic breccia. Rafts and lenses of Strelley material occur within this breccia zone (P. Morant, pers. comm., 1995), but north of this, the Strelley succession is cut out by the fault marking the southeastern boundary of the Lalla Rookh Basin. South of the sheared synclinal fold limb is another panel of the Strelley succession, which lies ?unconformably upon mafic and ultramafic volcanic rocks and interlayered chert of probable(?) Warrawoona Group age (see below). These relationships suggest a continuity of stratigraphy across the Central Pilbara Structural Corridor, thus implying that this structure does not represent a terrane boundary within the Pilbara Craton (see structure section below: Van Kranendonk and Collins, in press).

The base of the Strelley succession is not yet defined. A major change in assemblage from mafic-ultramafic volcanic rocks and interbedded chert to the intermediate and felsic volcanics of the Strelley succession is marked by a unit of lithic wacke. The relationship of the wacke to the underlying volcanic rocks is uncertain west and north of the Strelley Granite. However, mapping to the south of the granite in the area of Figure 4 has uncovered what is most likely an unconformable relationship of Strelley succession wacke and felsic volcanic rocks atop pillowed and massive high-Mg and tholeiitic basalts and interbedded cherts of what is inferred to be the southwesterly extension of the Warrawoona Group of the North Pole Dome sequence (see Fig. 1). Around the west-facing antiformal fold of the underlying rocks in this area, the Strelley succession apparently cuts down section from west to east, eliminating three distinct chert beds and intervening volcanic rocks in the underlying sequence. Along the northern limb of the west-facing anticline below the Strelley succession, interbedded mafic volcanic rocks and cherts are overlain, at generally a low, but locally at a high, angle, by a 1-2 m thick felsic volcanic flow distinguished by an eutaxitic base and flowtop breccia. Immediately below this unit, in different places, are lenses of pure quartzite, beds of red shale, felsic ash beds, and locally, channels, or palaeovalleys, filled with volcanic rubble and cut by felsic volcanic dykes (Fig. 5a). These units lie directly atop the underlying mafic volcanics, whereas the felsic volcanic flow is the first regionally developed unit of the overlying sequence. At one locality, a 40 cm thick unit of thinly-bedded quartz arenite at the base of the section is deformed into an upward-facing "blow", ostensibly caused by the forceful extrusion of volcanic gases from below (Fig. 5). The upturned, breached limbs of this structure are unconformably overlain by 10 cm of black-weathering breccia interpreted to represent a haematitic regolith, then 25-30 cm of the eutaxitic felsic volcaniclastic, a 15 cm thick bed of red shale, an eastward thickening lens of pure white quartzite, and finally by a vesicular felsic volcanic flow. Upsection, the felsic volcanic flow is conformably overlain by massive to weakly bedded wacke, interlayered felsic volcanics, massive and columnar-jointed dacites, and the regional marker chert (Fig. 4).

The underlying mafic volcanic sequence comprises pillowed and massive, interlayered high-Mg and tholeiitic mafic volcanic rocks and several distinctive cherts. This assemblage is very similar to the volcanics of the North Pole Dome and thus thought to be part of the Warrawoona Group. Further, their orientation in a SW-facing antiformal fold is compatible with their being the southwestern, higher stratigraphic equivalents of the North Pole sequence, which is also folded into this pattern. A possible Warrawoona affinity for this assemblage is supported by the fact that it differs from mafic and ultramafic volcanic rocks thought to represent the lower part of the Strelley succession on the southern limb of the Pilgangoora Synform (see "Supracrustal rocks of uncertain affinity" below). Furthermore, the rocks below the postulated unconformity display local evidence of having been folded prior to deposition of the Strelley succession, suggesting they have been affected by an early deformation not present in the overlying rocks. Such a relationship is indicated in Figure 5a by the truncated folds of chert C3, and by the discontinuous outcrop of chert C2. If true, the unconformity present in this area may represent a hiatus of some 210 Ma.

Gorge Creek Group

Overlying the dominantly volcanic Strelley succession are sandstones, shales, iron formation and mafic volcanic rocks of the Gorge Creek Group (Hickman, 1983). The group differs in composition and unit thicknesses across the map area (see Eriksson et al., 1994 and references therein), and has not been widely studied in detail for this report. However, two sections are important regarding the regional structure and the extent and continuity of stratigraphic successions in this area.

i) East of the Strelley Granite. Disconformably overlying the east-facing Strelley succession east of the Strelley Granite, is a basal onlap sequence of sandstones and conglomerate, shales, gabbroic sills, and high-Mg volcanic rocks (Honeyeater basalt: Hickman, 1983). The marker chert at the top of the Strelley succession is locally cut by Neptunian sandstone dykes of the overlying Gorge Creek sandstone, indicating a hiatus at least long enough for the chert to become lithified and fractured.

The basal part of the Gorge Creek Group immediately above the marker chert at the Sulpher Springs gossan contains a kilometre-scale olistostrome breccia in which pebble to house-sized blocks, and even kilometre-long rafts, of the marker chert, banded iron formation, and felsic volcanic rocks from the underlying sequence are supported within a fine-grained, grey, silty matrix (Figs. 6, 7a). Bedding is uncommon in the matrix. The km-scale raft of iron formation is deformed into an east-facing tight fold, the southern limb of which is cut out by breccia. Numerous, non-cylindrical, randomly-oriented folds in the iron formation, and textural evidence of syn-sedimentary deformation in the unit (Fig. 7b), attests to deformation in the sedimentary environment and not as a result of later tectonism. The core of the syn-sedimentary fold is occupied by a felsic volcanic porphyry sill, interpreted as having been emplaced into, and nucleating the fold. Pepperite textures around the margins of the sill, irregular intrusion shapes, and the presence of floating blocks of felsic material in the breccia are used to interpret a syn-deformational emplacement age for the felsic sill in the core of the breccia.

Sedimentary rocks overlying the olistostrome breccia at Sulpher Springs fine upwards from chert pebble conglomerate, through metre-thick bedded sandstone, to red shale of the Corboy Formation (Fig. 8), and form an onlap sequence on the marker chert. The onlap sequence thickens from the olistostrome breccia to the southeast and has been interpreted as a canyon fill sequence (Wilhelmij and Dunlop, 1984). Above the breccia, however, the coarse clastic sequence thins and pinches out, where it is replaced by 10 m of very finely bedded green shale (Fig. 8). West of the fault which cuts up through the Strelley Granite, the breccia unit thins to the west, as does the overlying green shale, and these rocks are once more overlain by granule conglomerates, coarse-grained wackes and thick-bedded rusty orange-weathering sandstone before passing up to the red shales of the Cleaverville Formation. This symmetrical east and west distribution of grain size and bed thicknesses across the olistostrome breccia is used to infer that this unit formed a topographic high during deposition of the basal part of the Gorge Creek Group. Valleys on either side of the high were filled by coarse sands derived through erosion of the breccia (cf. Wilhelmij, 1986), while the topographically high, thickest part of the breccia was covered by a thin veneer of shale.

Panels of tightly folded banded iron formation occur within felsic volcanic rocks located north of the Strelley Granite and west of the fault which cuts up through it. Much of the deformation in the iron formation is of the soft-sediment type, similar to that affecting its counterparts to the east. These features and the interbedded nature of the banded iron formation with felsic volcanic rocks suggest that the iron formation was a product of the felsic volcanism and that the felsic volcanic environment was an unstable one which caused deformation of the iron formation soon after its deposition. Such a scenario is supported by observations, described in the structure section below, that the Strelley Granite laccolith caused extensive folding and deformation of overlying rocks during its intrusion into the volcanic pile only some 1.5 km below the rock-water interface. Further, a geochemical link between the granite and overlying Strelley felsic volcanics (Vearncombe, 1996) show that granite emplacement, felsic volcanism, and deformation were synchronous. The presence of felsic volcanic blocks and syn-sedimentary sills in the olistostrome breccia above the maker chert provide a link between the Strelley succession and the Gorge Creek Group, and indicate that they are not separated by a large time gap.

ii) Pilgangoora Synform. The sedimentary rocks in the core of the Pilgangoora Synform belong to the Gorge Creek Group and include iron formation, turbiditic sandstone, shale, and interbedded conglomerate and sandstone deposited as an uplapping submarine fan in a fault-controlled, rapidly subsiding basin (Wilhelmij and Dunlop, 1984; Wilhelmij, 1986). These rocks locally lie in unconformable contact above pillowed mafic volcanic rocks of the Pilgangoora Belt (Wilhelmij and Dunlop, 1984; Wilhelmij, 1986), as shown, for example, in a section across the mafic volcanic-sediment contact along the southern margin of the basin (Fig. 9a). In this area, a steeply-dipping, but otherwise well preserved, sequence of pillowed, high-Mg mafic volcanic rocks becomes distinctly red stained by haematite alteration approximately 10 m from the contact with overlying sediments. At the contact, relict topography at a scale of metres can be observed in the underlying rocks, although bedding in rocks above and below the contact are roughly coplanar, indicating a disconformity. Immediately above the contact, on palaeohighs, is a section comprising 50 cm of ferruginized and mildly brecciated high-Mg basalt, ≤1 m of rubbly breccia, and 10's of cm of fine-grained red siltstone. The brecciated and altered rocks are interpreted to represent a palaeoregolith. In palaeovalleys, 50 cm of regolith-type breccia is overlain by 4-5 m of pale yellow, thinly-bedded (mm-25cm) felsic(?) ash. Overlying both sections is 1-2 m of chert pebble conglomerate and an unusually-textured, irregular unit of chert (?silicified carbonate: Fig. 9b). Pending petrographic and geochemical confirmation, it is suggested that the tuff beds are distal correlatives of the Strelley felsic volcanics, and the irregular unit of chert the equivalent of the marker chert at the top of the Strelley succession.

Supracrustal rocks of uncertain affinity

The ages of the volcanic and sedimentary rocks on either side of the clastic sedimentary core to the Pilgangoora syncline are undated, but interpreted to represent two monoclinal panels of different age facing towards one another (Fig. 1; see also Krapez, 1993). To the north, basalts and interbedded cherts of suspected Warrawoona Group age (ca. 3460 Ma) lie unconformably on the Coonterunnah succession across the distinctive Strelley Pool chert, whereas to the south, high-Mg pillow basalts and komatiites of suspected Strelley succession age (3240 Ma) crop out.

The interpretation of a Warrawoona Group age for the northern panel is based on published correlations of the Strelley Pool chert with the Towers Formation elsewhere in the eastern Pilbara (Hickman et al., 1990) and a similarity of lithology with Warrawoona Group rocks elsewhere (R. Buick, pers. comm., 1996). The interpretation of a Strelley age for the southern panel is based in part on structural considerations outlined below, and on the presence of a unit of wacke at the base of the panel that is identical to that at the base of the Strelley succession in the type area around the Strelley Granite (P. Morant, SIPA Resources, pers. comm., 1996). Regional sedimentological constraints have also been used to come to the same conclusion (Krapez, 1993; Eriksson et al., 1994).

Supracrustal rocks within the Tambina Complex and Western Shaw Belt are also undated and thus of uncertain stratigraphic affinity. The sequence which flanks the Tambourah Dome and wraps around the Tambina Granite is distinct from previously described sequences of the Warrawoona Group because it contains a number of sandstone interbeds. However, it is more like the Warrawoona Group than the Strelley succession and contains thick, fault-bound panels of felsic volcanic rocks which are similar to the Duffer Formation. The rocks of the Tambina Complex immediately west of the Shaw Batholith contain many chert interbeds and are considered to belong to the Warrawoona Group. An older age for all of these rocks is supported by the observation that they contain two distinct sets of structures, in contrast to the younger rocks of the Strelley succession and Gorge Creek Group which only contain one set of structures (regional D3: see below).

De Grey Group

Coarse clastic boulder conglomerate, sandstone, and mudstone of the De Grey Group crop out within two small fault- and unconformity-bound basins northeast and northwest of the Strelley Granite (Krapez, 1984; Wilhelmij and Dunlop, 1984), and in the Tambina Complex (Boulter et al., 1987) (Fig. 1). In the Lalla Rookh Basin, the group fines upward from boulder conglomerates deposited above an unconformity developed on the Gorge Creek Group along the southwestern margin of the basin, in which a relict topography can be recognised (Krapez, 1984). In the Strelley block, the De Grey Group lies paraconformably to unconformably above the Gorge Creek Group (Wilhelmij, 1986) whereas in the Daltons complex, the De Grey Group was deposited above the Gorge Creek Group across an angular unconformity. The Lalla Rookh Basin was previously interpreted as a strike-slip pull-apart basin on the basis of rapid lateral facies variations within the basin and marginal alluvial fans interpreted to have been deposited during strike-slip movement on the bounding faults (Krapez, 1984, 1993). However, as described below, an analysis of the bounding faults and regional structures indicate that the Lalla Rookh Basin is a mini foredeep-style basin which formed in advance of the northerly-moving Strelley Granite during late strike-slip deformation within the Central Pilbara Structural Corridor (see Structural Geology: Van Kranendonk and Collins, in press).

Fortescue Group

This has not been mapped.

Granitoid rocks

Seven main areas of granitoid rocks crop out in the map area. The Carlindi Batholith, southern parts of the Shaw Batholith, and the North Pole Dome have been only cursorily examined. A detailed mapping project on the Strelley Granite is currently underway as part of a Ph.D. study by Carl Brauhart at the University of Western Australia. A Ph.D. study of the Shaw Batholith margins by Zegers (1996) accompanies previous studies by Bettenay et al. (1981) and Bickle et al. (1980, 1985, 1989, 1993).

Carlindi Batholith

This body is composed largely of medium-grained, well-foliated biotite (±hornblende) granite-granodiorite-tonalite and subordinate migmatite. A late phase of massive porphyritic microgranite along the southern margin of the batholith, that cuts an older, non-foliated granodiorite phase of the batholith and intrudes the Coonterunnah succession, is 3468±4 Ma (Buick et al., 1995). This phase is unconformably overlain by the Strelley Pool chert of the Warrawoona Group, and thus provides a maximum age for the volcanic rocks of the Strelley Belt (Salgash Subgroup?).

North Pole Adamellite

This body is a medium- to coarse-grained, undeformed granite (adamellite), with porphyritic textures and miarolitic cavities indicative of a high level of emplacement. Dated at 3459±18 Ma (Thorpe et al., 1992a), the North Pole Adamellite is interpreted to represent the exposed cupola of a syn-volcanic laccolith emplaced at a high-level into mafic volcanic rocks, synchronous with felsic volcanism in the contemporaneous Panorama Formation (3459+9/-4 Ma: Thorpe et al., 1992a). Syn-emplacement doming during laccolith emplacement is indicated by a radial distribution of palaeocurrent directions in Panorama Formation felsic volcaniclastic rocks, with transport of material away from the core of the dome (DiMarco and Lowe, 1989).

Strelley Granite

This body is a zoned, multiphase, syn-volcanic granite intrusion emplaced ~1.5 km below the rock-water interface at ca. 3240±5 Ma (age from R. Buick, pers. comm., 1995). The Strelley Granite represents the magma centre for the felsic volcanic rocks of the Strelley succession and the generator of hydrothermal circulation systems responsible for the formation of Cu-Zn massive sulphide deposits in this region (Vearncombe et al., 1995). Tilted to a 60° E-dipping cross-sectional view, the intrusion is composed of a coarse-grained interior rimmed by a medium- to fine-grained granophyric margin. A later, heavily-altered phase occupies the core of the body (C. Brauhart, U.W.A., pers. comm., 1996) and caused the intrusion to expand from a sill into its present, roughly triangular cross-sectional shape. Adjacent supracrustal rocks are deformed into tight folds (as outlined below), and affected by contact metamorphism as seen by cm-scale rosettes of epidote in felsic schists north of the granite. The geometry of the Strelley Granite and associated structures, combined with the geochemical and temporal link between the granite and overlying felsic volcanics and hydrothermal Cu-Zn deposits of the Strelley succession (Vearncombe, 1996), all point to the Strelley Granite having been emplaced as a syn-volcanic laccolith, and more particularly, as an asymmetrical sphenolith (cf. Dixon and Simpson, 1987).

Tambina Granite

This is a complex body comprising several undated phases in two sublobes, that is enveloped by supracrustal rocks and highly strained. The western lobe is a shallow-dipping, dish-shaped body which is composed of many phases and contains evidence of a complex history. The eastern lobe is a doubly-plunging dome, reclined to the southwest with an overturned southwestern limb. It is composed of porphyritic granite-granodiorite that is texturally similar to the 3430 Ma South Daltons Pluton of the Shaw Batholith (McNaughton et al., 1993). The parallel nature of the contacts between the granite lobes and the surrounding stratigraphy, combined with structural complexity of the lobes which have been deformed by both sets of regionally-developed structures, suggests that they were emplaced as sills into the volcanic sequence, probably at the time of volcanism (age unknown, but probably 3460-3430 Ma).

Daltons Creek Granite

This previously unnamed granite is a medium-grained, partly recrystallised intrusion which cuts upturned, amphibolite-facies greenstones along the western margin of the Shaw Batholith (A on Fig. 1). It is deformed by a weak, but penetrative greenschist-facies foliation and is variably altered. Unpublished SHRIMP data indicates an igneous age of 2936±5 Ma for this body (M. Van Kranendonk, 1995). This age is intepreted as a minimum age for tilting of the supracrustal rocks which are cut by the granite, and associated amphibolite-facies metamorphism associated with formation of the Shaw Batholith dome (see below).

Shaw Batholith

This is a highly complex body, comprising several distinct phases ranging from migmatitic gneisses to massive, post-tectonic granite, that spans an age range of 3500-2830 Ma (Bettenay et al., 1981; Bickle et al., 1985, 1989, 1993; Zegers, 1996). The homogeneous North Shaw Suite, dated as ca. 3460 Ma, is coeval with Duffer Formation volcanism in the Warrawoona Group (Pidgeon, 1984). Much of the batholith has recently been found to be composed of ca. 3460 Ma tonalite-trondjheimite-granodiorite (TTG) intrusions (Zegers, 1996), some of which may have been emplaced during dome formation and (?)core complex formation (Zegers et al., 1996. However, published Pb-Pb ages of ca. 3300 Ma from the northern margin, and 2950-3000 Ma in the central part indicate a complex, multiphase history (Bickle et al., 1989). Recent dating (unpublished SHRIMP data; Van Kranendonk, 1996) has found evidence for leucosome vein development and amphibolite-facies metamorphism at ca. 3400 Ma close upon the end of TTG magmatism at ca. 3430-3410 Ma. Furthermore, concordant zircon from the amphibolite-facies Mulgandinnah shear zone along the western margin of the batholith indicate episodes of zircon growth at ca. 3300 Ma and 2930 Ma, both of which are associated with pulses of shearing, the latter of which was sinistral (see also Zegers,1996).

Yule Batholith and Tambourah Dome

Four main granitoid phases have been recognised.

1) Migmatitic orthogneisses; includes grey tonalitic gneiss and porphyritic mesocratic tonalite gneiss, both with granitic leucosome veins. These gneisses occur as thin remnants along the northeastern margins of the Yule Batholith and Tambourah Dome, where they are intruded by younger granitoids. Tonalitic gneiss in the batholith has been dated as ca. 3470 Ma, whereas porphyritic mesocratic tonalite in the Tambourah Dome is ca. 3420 Ma (unpublished SHRIMP data; M. Van Kranendonk, 1996).

2) Homogeneous, leucocratic, medium-grained granodiorite to granite, emplaced as sheeted sill complexes and intrusions into the older gneisses, synchronous with granitoid dome formation at 3240 Ma (unpublished SHRIMP data; Van Kranendonk, 1996: see below). These rocks form 95% of the Tambourah Dome and a large portion of the Yule Batholith (see also Hickman, 1983).

3) Small pegmatitic granite intrusions and dykes marginal to the Tambourah Dome, contemporaneous with sinistral shear at ca. 2950 Ma.

4) Undeformed, K-feldspar porphyritic granite underlying large areas of the batholith. Undated, but probably <2950 Ma.

Structural geology

Three sets of structures were recognised across the map areas, although not all structures are present in all rocks. The first set of structures (D1) were formed in response to the generally passive doming of supracrustal rocks during the syn-volcanic emplacement of granitoid rocks, which occurred at ca. 3468-3459 Ma in the Carlindi Batholith and North Pole Dome, and up to 3430 Ma in the Shaw and Yule Batholiths. The development of a set of leucosome veins in the Shaw and Yule batholiths at 3430-3400 Ma is due to unknown tectonic events, but is herein lumped together with this early stage of deformation. The two later structure sets are the result of granitoid doming (D2) and sinistral strike-slip deformation (D3), respectively. Post-D1 granitoid doming occurred at 3240±5 Ma in the Western Domain (Yule Batholith and Tambourah Dome), and over a much longer, poorly defined, period in the Shaw Batholith, between ca. 3359-3150 Ma (Zegers, 1996; unpublished zircon data, M. Van Kranendonk, 1996). Sinistral strike-slip deformation occurred at ca. 2950 Ma across the map area, as described in Van Kranendonk & Collins (in press). Table 2 shows how sets of structures relate to one another temporally across the map area.

Table 2: Ages of structures across the map area (data from Thorpe et al., 1992a; Buick et al., 1995; Zegers, 1996; Davids et al., in press; unpublished U-Pb SHRIMP data, M. Van Kranendonk, 1996).

	Yule batholith/Tambourah Dome	Carlindi/North Pole	Shaw Batholith
3470	Granitoid (?syn-volcanic) magmatism (3470-3420Ma)	Syn-volcanic doming (3468-3459) **D1**	Syn-volcanic doming
	(?migmatisation?)		
3400			migmatisation
			zircon recrystallisation
3300			
	doming & volcanism; clastic sedimentation in Gorge Creek Gp.	**D2** domal re-activation(?)	Ar-Ar ages, E. Shaw zircon recrystallisation
3200			
3100			
3000			
	D3 Sinistral strike-slip deformation, granite intrusion & clastic sedimentation (De Grey Gp.)		
2900			
			granitic plutonism
2800			

D1 deformation: syn-volcanic dome formation

3470-3400 Ma

The oldest known structures in the map area are those caused by the emplacement of syn-volcanic granitoid intrusions into the Warrawoona Group volcanic pile. Structures include folds and faults in the tilted Coonterunnah succession that are not developed in, and are unconformably overlain by, the Warrawoona Group and are thus ≥3468 Ma (Buick et al., 1995). Rocks of the Coonterunnah succession along the NE-SW striking margin of the batholith dip steeply to the west, face east, and are folded into non-cylindrical structures about NE-SW plunging axes (Fig. 10a). Reconstruction of the subvertically-dipping, east-west striking overlying rocks to the palaeohorizontal indicates that prior to tilting, the underlying rocks dipped moderately to steeply to the east, away from the Carlindi batholith contact, while folds plunged to the NNE and SSE on axial planar foliations subparallel to bedding (Fig. 10b).

The formation of structures in the older rocks is interpreted to have formed through tilting of the strata away from the granite batholith during its emplacement, probably at ca. 3468 Ma, which is the age of a late phase of the batholith that is unconformably overlain by the Warrawoona Group

(Buick et al., 1995). Although impossible to prove, the geometry of the folds is consistent with their having formed when material was shed off the rising granitoid dome, following a model based on similar relationships adjacent to the Mt. Edgar Batholith (Collins, 1989), the Strelley Granite (see below), and those commonly developed adjacent to salt diapirs (Jackson et al., 1990).

In the North Pole Dome, intrusion of the 3459 Ma North Pole Adamellite caused doming of the volcanic pile during Panorama Formation felsic volcanism, as indicated by a radial distribution of palaeocurrent directions in the felsic volcaniclastic rocks away from the core of the dome (DiMarco and Lowe, 1989). Intrusion of the granite magma and doming of the volcanics is thought to mimic, in plan view, similar features preserved in cross-section around the 200 Ma younger Strelley Granite (see below).

The Carlindi Batholith and North Pole Adamellite are relatively simple, high-level intrusions without internal evidence of a complex history. This is in contrast with the Yule and Shaw batholiths which contain migmatitic gneisses and several phases of younger granitoid intrusions. In these more complex, re-activated batholiths, the unravelling of their early history is difficult and the focus of further research. Strong re-activation and doming of the Yule Batholith at 3240 Ma (see below) precludes conclusive evidence for early syn-volcanic doming in this body. However, the presence of migmatitic leucosomes in ca. 3430 Ma rocks in both the Yule and Shaw batholiths suggests that the D1 deformation continued to younger ages than in the simpler domes to the north (see below). Rare F1 folds occur in deformed greenstones around the batholiths (eg. Western Shaw Belt and that west of the Tambourah Dome [see Fig. 16]).

Recently, Zegers et al. (1996) have proposed that the Shaw Batholith formed as a core complex at ca. 3460 Ma, based on the discovery of syn-volcanic extensional structures in the overlying Duffer Formation to the east of the batholith, east-side-up kinematics in the sheared eastern margin of the batholith (Split Rock Shear Zone), and the occurrence of sheets of granitoid material within the shear zone that cut some of the deformation fabrics and have 3460 Ma zircons (Zegers, 1996). They suggest that the eastern shear zone curls westward across the top of the batholith, south of the North Shaw Suite and thus formed part of a bowed, top-to the-west structure that accommodated extensional detachment of the cover sequence during doming. However, mapping in the northern part of the batholith failed to find the western extension of the shear zone. Furthermore, the fabric elements used to describe the shear were found to consist of deformed, amphibolite-facies metamorphic mineral assemblages and solid-state deformation fabrics, as opposed to the syn-magmatic features required by the core complex model at 3460 Ma, contemporaneous with voluminous TTG magmatism. Ar/Ar age data on metamorphic amphiboles from the Split Rock Shear Zone and the adjacent Coongan Belt, and recrystallisation of zircon in rocks from within the shear zone, indicate recrystallisation during metamorphism at ≤3308 Ma (Zegers, 1996; Davids et al., in press), and we suggest that the kinematics observed by Zegers et al. (1996) inside the batholith are probably of this age, different from the from syn-volcanic extensional structures in the Duffer Formation, thus indicating that the proposed core complex model of formation of the Shaw Batholith is constructed from a composite of different events. An alternative model (Van Kranendonk and Collins, in prep.) suggests instead that the Shaw Batholith represents an early, very large, syn-volcanic dome, and that it has been amplified by regional successive deformation events at ca. 3300-3240 Ma and at 2950 Ma (see below).

3430-3400 Ma

A distinct porphyritic hornblende tonalite in the Tambourah Dome is 3430 Ma, some 20-30 m.y. younger than the main phase of grey tonalitic orthogneiss in the Yule Batholith (unpublished SHRIMP data, Van Kranendonk, 1996) but the same age as the granodioritic South Daltons Pluton in the western margin of the Shaw Batholith (McNaughton et al., 1993; unpublished SHRIMP data, Van Kranendonk, 1996). In the Shaw and Yule batholiths, the ca. 3430 Ma intrusions have been migmatised, indicating deformation younger than this age. One such leucosome vein, cutting an amphibolite xenolith within grey leucotonalite orthogneiss from the western margin of the Shaw Batholith, contains 3400 Ma clear zircon overgrowths on 3430-3410 Ma zoned zircon cores, indicating migmatisation soon after magmatism and thus a continuum of magmatism through 3470-3400 Ma (unpublished SHRIMP data; Van Kranendonk, 1996). Unfortunately, similar leucosome veins collected from the Yule Batholith and Tambourah Dome failed to yield zircon. Pb-Pb model ages of 3392 and 3406 Ma on galena from the Normay Mine in the North Pole Dome (Thorpe et al., 1992a), combined with the suggestion of such an age for metamorphism of gneisses in the Mt. Edgar Batholith (ca. 3385 Ma; Williams and Collins, 1990), support the inference of a widespread magmatic and/or metamorphic/structural event at this time.

D2 deformation: batholith re-activation, granite doming, and volcanism

ca. 3325-3300 Ma

This age is well-known from granitoid rocks in the Mt. Edgar and Corunna Downs batholiths immediately east of the map area, and is related to granitoid doming through active magmatic diapirism, coeval with the deposition of felsic volcanic rocks of the Wyman and Kelly formations (Cooper et al., 1982; Williams and Collins, 1990; Thorpe et al., 1992a; McNaughton et al., 1993; Van Kranendonk, 1995; Collins and Van Kranendonk, 1995). Although no direct structures associated with this age of events have been recognised in the map areas, ca. 3300 Ma ages are known from the Shaw Batholith in the form of a Pb-Pb age of 3338±51 Ma on undeformed leucotonalite in the northern part of the batholith (Bickle et al., 1989), concordant zircons in two dated samples from the western margin of the batholith (unpublished SHRIMP data; M. Van Kranendonk, 1996), and recrystallisation ages of zircon from the south-central part and eastern margin of the batholith (Zegers, 1996). The ca. 3300 Ma ages in these latter samples have been obtained from a range of zircon types, including metamorphic zircons in demonstrably older rocks, to oldest zircons in shear-related leucosome veins from the sheared western margin of the batholith. Thus it is inferred that an event at ca. 3300 Ma affected the Shaw Batholith, that it was probably related to domal reactivation of the batholith, but that it did *not* involve extensive plutonism. Considering the importance of this age event in rocks immediately east of the map area, it is surprising that this age is not recorded in rocks west of the Shaw Batholith.

3240 Ma

The main phase of granitoid diapirism in the Yule Batholith and Tambourah Dome occurred at 3240 Ma (see below), at the same time as deposition of the Strelley succession, intrusion of the syn-volcanic Strelley Granite laccolith (R. Buick, U. Sydney, and C. Brauhart, U.W.A., pers. comm., 1996), and amphibolite-facies metamorphism of mafic volcanics adjacent to the Tambourah Dome in the Western Shaw Belt (Wijbrans and McDougall, 1987). Similar ages from further east in the craton, obtained by Rb-Sr whole rock dating (deLaeter and Martyn, 1986; Collins and Gray, 1990), from metamorphic zircon (McNaughton et al., 1993; Zegers, 1996), and from Ar/Ar ages on hornblende in the Coongan Belt (Davids et al., in press), attest to the widespread nature of this event.

Numerous Scrapes area of the Yule Batholith

The northeastern margin of the Yule Batholith contains a kilometre-scale dome-and-basin map pattern comprising panels of 3470 Ma tonalitic orthogneiss preserved along the outer margins of 3240 Ma granitoid domes, and a complexly folded, shelf-type sequence of quartzite-pelite-rhyolite-banded iron formation that is unique within the eastern part of the Pilbara Craton (Fig. 11: Van Kranendonk, submitted; ages from unpublished SHRIMP zircon data, Van Kranendonk, 1996).

The northeastern margin of the Yule Batholith may be subdivided into a dome-and basin subdomain and a northwestern, granite-dominated subdomain (Fig. 11). The structure of the dome-and basin subdomain is dominated by a central, elliptical dome which is cored by an irregular-shaped intrusion of homogeneous granitoid and outlined by a ring fault (Fig. 12). Inside of the ring fault, rocks are deformed into a refolded antiform, reclined to the east, while outside of the ring fault is a rim syncline that wraps around the dome in a horseshoe shape from the northwest, around the southern margin of the dome, and north along the eastern side of the dome (Figs. 12-14). Porpoising variations in the plunge of fold axes outside of the rim fault are controlled by the geometry of promontories and re-entrants in the homogeneous granitoid intrusion coring the dome, suggesting a control on structure by granitoid magmatism. Promontories around the southern half of the central intrusion correspond with fold plunge culminations and sites of preferred granitoid intrusion outside the ring fault, again suggesting a contemporaneous development of intrusion and structure (Fig. 12). Such map-scale relations of syn-kinematic granitoid intrusion are also seen at the outcrop scale (Fig. 15a).

A large-scale feature of the dome-and-basin subdomain is that fold plunges change from south-plunging in the south, to north-plunging in the north across an WNW-ESE axis which strikes subparallel to the granite-greenstone boundary. In the northern half of the subdomain, north-plunging folds in granitoid rocks are accompanied by a penetrative, N-S striking, subvertical foliation defined by biotite, that is absent in rocks to the south. The absence of N-S striking, S2 foliations in the southern half of the dome-and-basin subdomain, combined with the fanned distribution of fold axial traces out from the culmination in the northern apex of the subdomain, suggests constrictional strain in the northern half during folding and northerly flow of material during deformation (see below).

The dome-and-basin subdomain is bound in the northwest, north and northeast, by a 200-1500 m wide zone of greenschist-facies ultramafic schist and tectonic megabreccia (Figs. 11, 13). Contributions from both hanging wall volcanics and footwall 'shelf-sequence' metasediments occur in the zone, which dips everywhere away from the subdomain. The zone separates the dome-and-basin subdomain from low-grade mafic volcanics to the north and northeast, and from a granitoid-dominated subdomain of the batholith to the northwest. Rare kinematic indicators, dip-parallel mineral elongation lineations, and the juxtaposition of low-grade hangingwall rocks against high-grade footwall rocks indicate a normal sense of displacement across the zone, which relates to the extensional detachment of the greenstone cover.

In the northwestern, granitoid-dominated subdomain of the batholith, granitoid rocks are moderately foliated and folded into irregular shapes, some of which have nappe shapes indicating a northerly vergence (Fig. 11). Infolded supracrustal rocks are of mafic volcanic origin, and there is no sign of the shelf sequence present to the southeast. The granitoid rocks are separated from the low-grade greenstones to the north by a narrow, low-strain, gradational (?intrusive) boundary. North of the batholith, weakly-strained, low-grade volcanic and sedimentary rocks dip steeply N- to NE, away from the batholith.

Folding and granitoid doming within the batholith was accompanied by the syn-kinematic growth of high-T, low-P metamorphic mineral assemblages, including sillimanite-muscovite and garnet-biotite (one locality) in metapelites, and hornblende-plagioclase in amphibolites. The high-T, low-P character of the metamorphism is interpreted to reflect heat input from syn-kinematic granitoids emplaced during the folding.

Model of formation. Van Kranendonk (submitted) showed that the geometry of the central dome of the dome-and-basin subdomain and associated structural and magmatic features of this area are identical to that of some salt diapirs in the Great Kavir of central Iran (Jackson et al., 1990: Figs. 15b, 15c), and suggested a diapiric origin for the Numerous Scrapes area of the Yule Batholith. The time of dome formation and re-activation of the pre-existing granitoid basement is constrained by U-Pb zircon ages on the main central intrusion of the dome-and-basin subdomain, and on a syn-kinematic, axial planar granitoid dyke cutting migmatitic, ca. 3470 Ma orthogneisses in the northeastern hook of the same intrusion (Fig. 15a), that both give concordant, single zircon population igneous intrusion ages of ca. 3240 Ma (unpublished SHRIMP data by M. Van Kranendonk, 1996).

Tambourah Dome

Mapping of the Tambourah Dome (Fig. 16a) has shown that, as in the Numerous Scrapes area, its northern edge is composed of a thin rind (<200 m) of interlayered migmatitic leucotonalite and porphyritic hornblende tonalitic orthogneiss, the latter of which is ca. 3430 Ma (unpublished SHRIMP data by M. Van Kranendonk, 1996). The gneisses are cut by layer-subparallel, to highly discordant sheets and dykes of homogeneous, leucocratic granodiorite-granite that coalesce to form the interior of the dome (Fig. 16b). A sample of one such sheet has yielded zircons which give concordant U-Pb data at ca. 3240 Ma and some older (>3400 Ma) xenocrystic zircons (unpublished SHRIMP data by M. Van Kranendonk, 1996).

Three generations of structures were recognised in this area. Second generation folds (F2) include the prominent fold of greenstones around the northern apex of the dome and associated minor folds, that overprint rare, rootless isoclinal folds of bedding (F1 on Fig. 16a). D2 folds are transected by axial planar foliations and have axes colinear with mineral elongation lineations defined by amphibolite-facies minerals. The orientation of these structures varies around the dome. Southeast of the area shown in Figure 16a, axial planes dip steeply east, away from the dome, and folds have subhorizontal axes parallel to the margin of the dome. Folds in this area verge asymmetrically to the east, suggesting formation due to the shedding of the greenstones off the rising dome. North and northeast of the dome, folds plunge moderately NNW through to SSE on steeply-dipping axial planes (Fig. 16c). Greenstones along the western limb of the fold are much more highly strained than to the east, are characterised by L>>S fabric elements that plunge to the SW (Fig. 16d), and are deformed into non-cylindrical folds.

At the northern apex of the dome, the rind of older, migmatitic orthogneisses are cut by homogeneous granitoid sheets which show a progression of intrusion from gneissosity-parallel sills to younger sheets oriented at a progressively higher angle to the layering, to highly discordant, metres-wide, N-S striking dykes located within the axial planes of upright D2 folds (Fig. 16b). The high angle dykes extend north into the greenstones where they feed a triangular-shaped (plan view) intrusion located in the hinge region of the steeply-plunging antiform north of the dome.

D2 structures located north of the margin of the Yule batholith and along its strike extension to the southeast are overprinted by a D3 generation of folds and lineations developed under greenschist facies conditions (Figs. 16a, e, and f). The D3 structures are related to sinistral strike-slip shear, as described in the following section.

Model of formation. The above features are interpreted to indicate the syn-kinematic emplacement of granitoid magmas during the formation of the Tambourah Dome. In this model, granitoid magmas are interpreted as having first been emplaced as sills along the pre-existing interface between greenstones and orthogneisses. Progressive addition of magmas initiated doming, and subsequent generations of sills cut the tilting gneiss-greenstone contact at increasingly higher angles. In the late stages, dykes cut through the gneiss-greenstone interface at the apex of the fold and fed a saddle reef intrusion within dilated layering within the folded greenstone carapace to the granite dome (Fig. 16b). The gradual change in the orientation of fold axes around the dome, from NW-plunging in the east, to SW-plunging in the west, is thought to reflect differential strain intensity during doming, with the more transposed rocks in the narrow, tight syncline between the Tambourah Dome and the Yule Batholith having fold axes that are more colinear with the finite extension direction.

An inference of the model is that the age of the granitoid rocks should reflect the age of the D2 deformation, as well as the age of associated amphibolite-facies metamorphism of adjacent greenstones produced through conductive heating from the intruding granitoid magmas. U-Pb SHRIMP data (unpublished, M. Van Kranendonk, 1996) from a homogeneous granitoid sheet along the northern rim of the dome, that is subparallel to the gneiss-greenstone contact, indicate an age of intrusion of 3240 Ma. This age is similar to the oldest age from blue-green hornblende adjacent to the eastern margin of the Tambourah Dome (≥3234 Ma: Wijbrans & McDougall, 1987). A sample of a syn-D3, ca. 2930 Ma granite intrusion cutting older, folded granitoid rocks in the saddle reef of the D2 fold north of the Tambourah Dome, contains a concordant xenocrystic zircon which is 3240 Ma, interpreted to originate from the older, syn-D2 granitoid intrusion and thus dating the time of D2 folding (Fig. 16e: unpublished SHRIMP data by M. Van Kranendonk, 1996).

Strelley Granite laccolith

The complexly folded rocks north of the Strelley Granite laccolith (Fig. 6) contain evidence of two generations of folds. At Sulpher Springs, the second generation include open folds, trending to the NE, that are colinear with folds in the Lalla Rookh Basin (Fig. 17a) and thus interpreted to result from the D3, strike-slip deformation (see below). Similar open folds further west trend approximately NNW-SSE and plunge moderately (Fig. 17b). Associated with the late folds are a conjugate set of dextral E-W and sinistral N-S faults (Fig. 6).

The earlier, D2, set of structures include syn-sedimentary slump folds in the olistostrome breccia north of the Sulpher Springs gossan and a series of tight, "S"-asymmetric folds of a prominent chert unit immediately above the basal wacke located further to the west (C1 on Fig. 6). As described previously, the olistostrome breccia is a syn-sedimentary structure. Fold axes within it plunge into the northern hemisphere (Fig. 17c) and the overall vergence of the structure is to the east (Fig. 6). Similarly, the tight "S" folds of the C1 chert show a scatter of trends, generally to the north, that are roughly colinear with the calculated b-axis derived from folded bedding planes (Fig. 17d). These folds have highly irregular axial planes and occur only within the chert unit; adjacent units are not folded. Several panels of the chert, either folded or not, are isolated from one another across felsic intrusions, some of which are up to 200 m wide and display columnar jointing (Fig. 6). The S-asymmetric shape of the folds of the chert indicate a westerly vergence, away from the Strelley Granite. Note that folds in both the chert and the olistostrome breccia have fold axes at a high angle to the direction of vergence, and that the vergence direction is opposite on opposite sides of the thickest cross-sectional part of the granite.

In order to generate a better view of the original geometry of the early structures, an attempt was made at deconstructing the effects of the D3 strike-slip deformation (Fig. 18). The result of restoring the fault panels shows that the bed length of felsic volcanic rocks and overlying sediments at the top of the Strelley succession on the northern side of the Strelley Granite (fault panels A-E) is shorter than that of the underlying chert which retains the train of "S"-asymmetric folds (Fig. 18c). This discrepancy can be resolved if one removes the thickness of the syn-volcanic Strelley Granite (Figs. 18d-e), which shows that the folds in the chert most likely formed as a result of the shedding off of partly lithified cover off the ballooning Strelley Granite.

Model of formation. The origin of folds and dismemberment of the chert, the reason for its sole deformation relative to adjacent, undeformed units, and the opposite facing directions of folds in the chert and the olistostrome breccia can all be related to emplacement and ballooning of the Strelley Granite laccolith. In this model (Fig. 19), the folds are interpreted to have formed during shedding off of material from atop the inflating intrusion, as they occur only in that part of the chert located along the steepest slope of the laccolith. The dismemberment of the chert by felsic intrusions is interpreted to reflect injection of felsic magmas off the main intrusion during the deformation. Similarly, the preferential, plastic deformation of the chert relative to adjacent units is thought to be due to its being deformed within the high temperature contact metamorphic aureole around the granite and the high silica content (low temperature of plastic deformation) of the chert. In the fourth step of the deconstruction (Fig, 18d), the late, heavily-altered phase is removed from the granite, deflating it to more of a sill-like body. In this reconstruction, it can be seen that both the folds in the chert and in the olistostrome verge downslope from a topographic high caused by the asymmetric intrusion beneath. The topographic high (and later locus of faulting) is the site of an anomalous thickness of felsic volcanic rocks and associated iron formation, and is thus interpreted to represent a felsic volcanic centre fed from the top of the underlying magma chamber (Fig. 19). Disruption by magma replenishment and expansion of the laccolith caused syn-sedimentary folding of strata at the margins of the eruptive centre, particularly the olistostrome breccia above Sulpher Springs, which is interpreted as having formed due to gravity slumping off the topographically highest point above the asymmetrical laccolith (Fig. 19). Later infilling of the adjacent valley was completed by the onlap sequence of sandstones of the Gorge Creek Group (cf. Wilhelmij and Dunlop, 1984).

The geometry of the syn-volcanic laccolith and associated structures preserved in cross-section in the 3240 Ma Strelley succession, is used as an analogue to explain the formation of older (ca. 3460 Ma) syn-volcanic domes of the eastern Pilbara Craton, which are preserved in plan view, such as the North Pole Dome and possibly also the earliest phases of the Shaw and Carlindi batholiths. This is not to say that the syn-volcanic granitoid intrusions coring the larger domes are necessarily underlain by more greenstones. Rather, the laccolith model is presented as a way to rationalise the relationship of the domes themselves to the overlying greenstones. As previously described for the Tambourah Dome and central dome of the Numerous Scrapes area, the granitoid magmas forming the domes are interpreted to have been emplaced as sheeted sills within the greenstone stratigraphy at a gravitationally stable level in the crust, but that with continued extension and more voluminous intrusion, the domes started to rise and subsequent intrusions were emplaced at progressively higher angles to the originally subhorizontal stratigraphy.

Regional model

The similar ages of granitoid magmatism in the Yule Batholith and of adjacent ultramafic-felsic volcanism in the Strelley succession leads to the conclusion that diapirism and basin formation were contemporaneous, and linked, events in this area, as it was further east at ca. 3325-3300 Ma (Van Kranendonk and Collins, 1995). A model to explain the evolution of this region (Fig. 20) is based on modern core complexes from the D'Entrecasteaux Islands, off Papua New Guinea (Baldwin et al., 1993) and on theoretical considerations of the effects of the interaction between a mantle plume and pre-existing sialic crust (cf. Campbell and Hill, 1988).

In the following model (Van Kranendonk, submitted), extension is proposed as the driving mechanism for formation of the Yule Batholith through magmatic diapirism and re-activation of pre-existing crust and for coeval deposition of the sequence preserved across the southern limb of the Pilgangoora Syncline (basal shelf sequence, mafic through komatiitic volcanics, andesitic to rhyolitic volcanics of the Strelley succession, and the sedimentary rocks of the Gorge Creek Group), through the following, continuous sequence of events:

1) thinning of pre-existing sialic crust, erosion of the Warrawoona Group along the southern limb of the Pilgangoora Synform, and deposition of a shelf-type metasedimentary sequence on Yule orthogneiss;

2) core complex style re-activation of pre-existing basement gneisses, accompanied by voluminous mafic volcanism during the main phase of basin formation;

3) active magmatic diapirism during coeval extensional detachment of the overlying mafic volcanic sequence, and;

4) deposition of the overlying clastic sedimentary basin sourced by uplifted granitoids in the Yule batholith during the waning phase of extension and decreased heat input.

In this model, the significant heat input required from the lower crust or mantle to drive the extension is interpreted to be derived either from a mantle plume, or from backarc spreading following an analogy with modern core complexes in the D'Entrecasteaux Islands, Papua New Guinea (Baldwin et al., 1993). In either scenario, pre-existing sialic crust became re-activated to such an extent that it became thinned enough to rupture and allow the formation of a new greenstone belt. Asymmetrical extensional thinning caused the overlying Warrawoona Group to become completely eroded along the southern margin of the Pilgangoora syncline, where the thickest section of supracrustal rocks was deposited. As the volcanic section thickened, it started to become detached off the reactivating sialic crust of the batholith, following an analogy with the D'Entrecasteaux Islands core complexes. Syn-extensional granitoid magmas, probably derived from remelting of older gneisses (cf. Collins, 1993), were first emplaced above the boundary between the shelf sequence and overlying mafic volcanic sequence (northwestern subdomain of the batholith), but with continued extension and concomitant thinning of the overburden, were emplaced lower down in the section, at and below the shelf/mafic volcanic sequence boundary. As the mafic rocks extended off the batholith and the lithostatic load was released, magmas were able to drive diapirism of the reactivated basement. Diapirism occurred in distinct nodes, probably as a function of original heterogeneities in the thicknesses of either, or both of, the basement sialic crust and overlying volcanic pile: specific nodes include the dome-and-basin subdomain of the Yule batholith, and the Tambourah Dome. In the final stages of the extensional event, detachment of the greenstones off the batholith was completed to the extent that granitoid rocks were exposed to erosion and fed the clastic basin in the Pilgangoora Syncline.

Evidence supporting a coeval development of the greenstone basin and reactivated granitoid batholith comes from recent SHRIMP dating of granitoid rocks in the dome-and-basin subdomain of the batholith, which are the same age as the Strelley Granite and associated intermediate volcanic rocks (unpublished data, M. Van Kranendonk, 1996). Further dating of material from lower and higher in the supracrustal sequence, and of granitoids from a variety of settings in the batholith, is underway in order to constrain the timespan of the proposed extensional event.

D3 deformation (ca. 2950-2930 Ma): sinistral strike- slip shear

A north-south corridor between the Shaw Batholith and North Pole Dome in the east, and the Yule and Carlindi batholiths in the west, is strongly deformed by sinistral strike-slip shear at ca. 2930 Ma. Referred to as the Central Pilbara Structural Corridor (CPSC: Van Kranendonk and Collins, in press), it is bound to the west by a curviplanar, greenschist-facies zone of shear and fault breccia that extends north through the centre of the Western Shaw Belt, along the curved northeastern margin of the Yule batholith, due north through greenstones immediately west of the Strelley Granite, around the top of the Strelley Granite where it curves to the east, and then northeast where it forms the northwestern bounding fault of the Lalla Rookh Basin (Figs. 1, 21). Sinistral kinematic indicators were observed in association with shallow-plunging mineral elongation lineations at several places along the western boundary fault, most notably in the Western Shaw Belt, within the margin of the Yule Batholith, and within the fault north of the Strelley Granite and northwest of the Lalla Rook Basin. Sinistral kinematic indicators were also observed in the amphibolite-facies (hornblende-biotite) shear zone developed along the entire western margin of the Shaw Batholith, which forms the eastern boundary to the CPSC (Fig. 22).

The eastern margin of the CPSC is a ragged series of splays formed along the western margin of the Shaw batholith, the southern limb of the Soanesville Syncline, the southern limb of a syncline southeast of the Strelley Granite, and the southeastern boundary fault of the Lalla Rookh Basin. On the NE-striking splays north of the Shaw Batholith, rare kinematic indicators of dextral shear were observed.

In between the western boundary shear and the eastern margin of the CPSC, rocks are deformed into a series of tight, upright antiforms and synforms with NE-SW striking axial planes and NE-plunging axes (Fig. 21: see also Fig. 17a). The tight to isoclinal, NE-plunging, "S"-asymmetric folds in Gorge Creek Group supracrustal rocks along the concave margin of the Yule batholith are interpreted to be drag folds associated with sinistral shear along the western boundary fault. Similarly, the tight fold of pillowed, high-Mg basalts adjacent to the western boundary fault north of the Strelley Granite is also interpreted to represent a drag fold (Fig. 6). Resolution of shear sense indicators and the geometry of the folds indicates a contemporaneous development for these structures as a result of a NW-SE oriented maximum compressive stress (Fig. 21).

Within the CPSC, rocks of the Soanesville Syncline, the Strelley succession, and Lalla Rookh Basin are preserved at low to very low metamorphic grade and show only the one set of tectonic structures, related to the D3 strike-slip deformation. In contrast, retrogressed amphibolite-facies greenstones south of the Soanesville Syncline contain an earlier amphibolite-facies set of structures, (D2) and a later set of greenschist-facies, D3 structures, suggesting a probable older protolith age for these rocks. In the Tambourah Dome, early, amphibolite-facies structural fabric elements related to doming at ca. 3240 Ma are folded and retrogressed within the CPSC by the greenschist-facies sinistral shear event (see also Wijbrans and McDougall, 1987). Structures formed by this event include small-scale folds with a consistent "S" sense of asymmetry, subhorizontal mineral elongation lineations and fold axes, and foliations in granitoid rocks northeast of the Yule Batholith margin that are defined by muscovite and chlorite (Figs. 16a, f). Syn-kinematic granite pegmatites emplaced into dilatational spaces opened up by counterclockwise rotation of the saddle reef intrusion north of the Tambourah Dome (Fig. 16e) contain abundant sinistral kinematic indicators (eg. winged K-feldspar porphyroclasts, asymmetrical extensional shear band foliations, and S-C textures).

As described in Van Kranendonk and Collins (1995, in press), the CPSC of sinistral strike-slip deformation is thought to have resulted from northerly indentation of the Shaw Batholith into the relatively undisturbed Strelley succession and North Pole Dome (Fig. 23a). Indentation caused the lateral westerly extrusion of the originally subhorizontal Strelley Granite laccolith and associated volcanic and sedimentary rocks, much like an orange pip may be squeezed out from between one's fingers. With continued compression, the Strelley Granite - acting as a rigid body in deforming greenstones - ramped up the concave, western margin of the Yule Batholith and became tilted on its side, similar to a car tilting over on two wheels as it drives over a ramp on one side (Fig. 23b). Westerly movement of the granite was transferred to the north as it slid around the concave arc of the Yule batholith. Northerly translation was halted by its impingement against the rigid Carlindi batholith. Crust following the Strelley Granite to the south became crumpled up behind it, in a manner similar to cars piling up in a highway crash (Fig. 23c: Van Kranendonk and Collins, 1995, in press). Progressive compression exhumed deeper-level, amphibolite-facies rocks to the south across high angle reverse shear zones located along the northern and southern limbs of the Soanesville syncline.

Shear-induced folding occurred across NE-SW trending structures, such as the Soanesville synform, the folds north of the Strelley Granite, and those of the De Grey Group in the Lalla Rookh basin (Fig. 1). Many of the folds characteristically have sheared off limbs. An anomaly to this general geometry, however, is a N-S trending, steeply S-plunging syncline affecting the northern limb of the Soanesville syncline, located west of the west-facing panel of the Strelley succession shown in Fig. 4 (see Fig. 1). This fold represents the southwesterly continuation of the sheared out synform east of the Strelley Granite, which has been deflected counterclockwise to the south as a result of northwesterly indentation of the rigid, west-facing panel of Strelley and possible Warrawoona age rocks. A fault with kilometres of displacement bounds the entire northern, western, and southern margins of this panel, and shows evidence of sinistral shear along the south and west, and dextral shear along the north. The synform west of the rigid Strelley panel and the antiform in the panel itself, are both interpreted to have formed during the regional folding which accompanied sinistral shearing. The narrow panel joining the main outcrop of the Strelley succession to the west-facing panel is itself tightly folded, with overturned, sheared limbs. The complex geometry in this area is interpreted to result from the interference between deforming Strelley succession rocks along a NE-SW striking axial plane and a rigid eastern buttress with a sharp, N-S oriented margin, represented by the southwest-facing continuation of the North Pole Dome.

Sinistral shearing and folding was accompanied by deposition of the clastic sediments of the DeGrey Group (Fig. 23b, c: Krapez, 1984). Remnants of the group in the Lalla Rookh basin, the Strelley block (Wilhelmij and Dunlop, 1984) and the Daltons complex (Boulter et al., 1987) indicate an originally more widespread distribution of these rocks than presently preserved. The coarse clastics of the Lalla Rookh basin have previously been interpreted as having been deposited in a strike-slip pull-apart basin (Krapez, 1984). However, several key features of the geometry of this basin, and regional kinematics of the model presented in Figure 23 suggest that such an interpretation is open to question. First, the unconformity with underlying Gorge Creek Group rocks is subparallel to the bedding in the older units, suggesting initiation of De Grey Group deposition while Gorge Creek (and Strelley) rocks were still flat lying. As seen from the model of Figure 23, this occurred well away from the present site of the basin. Second, the previous model shows south-directed movement of the block south of the Lalla Rookh basin (Krapez, 1984; Fig. 41 of Eriksson et al., 1994), but this is incompatible with the well-documented sinistral movement across the western boundary fault. Third, the southeastern boundary fault of the basin terminates at an east-west striking fault at a point where lake sediments were deposited in the latest part of the basin history, indicating a down-to-the-north sense of displacement across the E-W fault. This geometry suggests southerly movement of the block east of the southeastern boundary fault, and thus *dextral*, not sinistral, shear across the southeastern boundary fault. Fourth, rocks near the base of the basin are tightly folded by the regional set of D3 folds developed outside the basin, but become less strongly deformed upsection to the lake sediments. This feature suggests progressive deposition during folding, and that the earliest sediments were affected by most of the D3 strike-slip deformation as they became progressively tilted along with the underlying rocks.

These features related to the Lalla Rookh Basin, combined with the geometry of the regional D3 structures, indicate that the rocks in the Lalla Rookh Basin were deposited as a type of mini foreland basin in front of the Strelley Granite as it travelled north. Sedimentation of the group commenced while the granite and associated supracrustal rocks were still subhorizontal, and over a wider area than presently indicated. As the Strelley Granite and associated supracrustal rocks were tilted and uplifted, sedimentation progressed from coarse conglomeratic rocks to mudstones. The fining upward nature of the sedimentation in the group is thought to reflect the progressive decrease in the amount of displacement of the granite throughout its journey north (Fig. 23). Terminal lake sedimentation in the basin probably occurred through passive erosion of the uplifted rocks to the south when the Strelley Granite stopped moving as a result of becoming pinched between the Carlindi Batholith and North Pole Dome.

Deformation of the De Grey Group within the Lalla Rookh Basin was accompanied by folding of the felsic volcanic dome shown in Figure 3, outside of the basin. The fold axial planes of this structure are coplanar with those within the basin and are thus thought to have occurred during regional D3 deformation and sinistral shear. No earlier structures were observed in these rocks.

Regional sinistral shear along the western margin of the Shaw batholith is interpreted to account for the origin of complex folds in this region, in contrast to the model presented by Bickle et al. (1980, 1985) who interpreted them as early nappes formed in an Alpine-style horizontal tectonic regime and subsequently tilted on their side by diapirism. The principal reason for re-assessing the earlier model was that, as opposed to the dextral kinematic indicators required by the proposed model for the nappe-style origin of the folded granitoid lobes, the widespread development of *sinistral* kinematic indicators was observed. In addition, some of the isoclinal folds indicated by the previous authors were instead found to be intraplutonic screens of altered greenstones within more openly folded granitoids, indicating that the cross-sections of the region presented in the earlier publications were inaccurate.

Results of detailed mapping of this area show that the folds of the northwestern Shaw batholith occur between two contemporaneous splays of a prominent sinistral shear zone developed along the western margin of the Shaw batholith, at the point where the batholith margin takes a sharp jog to the east (Fig. 24). The folds have the same orientation of amphibolite-facies linear fabric elements as both splays of the shear zone, thus suggesting a contemporaneous time of development (Van Kranendonk and Collins, in press). The intervening area between the splays is a natural site for localised thickening as a result of compression in the restraining bend of a shear system (cf. Davis, 1984) and it is proposed that the folds - or at least much of the folding - formed as a result of *localised* thickening within a restraining bend of the regional sinistral shear system (Fig. 25). In this model, southerly displaced greenstones to the north of the folds interacted with northerly displaced granitoids to the south, thus producing the localised folding, thickening and high pressure metamorphism in this area. Deformation caused pre-existing granitoid lobes to be pushed overtop of surrounding greenstones, causing the local overturning of stratigraphy observed around the South Daltons pluton. Granitoid lobes 2-4 show diverse kinematics that indicate counterclockwise rotation and westerly lateral extrusion in lobe 2, the squeezing up and over to the north of granitoid lobe 3, and clockwise rotation and easterly lateral extrusion of lobe 4 as a result of splay of the southerly sinistral shear zone (Fig. 25). Thus, in contrast to the model of Bickle et al. (1980, 1985), the structures in this area are interpreted to be the latest set rather than the earliest (Van Kranendonk and Collins, in press).

The northerly extension of the sinistral shear zone along the western margin of the Shaw batholith is commonly perceived to continue due north and to link up with the fault cutting through the largest remnant of the Fortescue Group and then further north to link up with the zone of tectonic breccia, which itself links up with the southeastern boundary fault of the Lalla Rookh basin (Fig. 1). However, this fault is in part a post-Fortescue feature. The ductile 2950 Ma structures splay out to the north of the western margin of the Shaw batholith and curve around the batholith to the northeast. Another prominent zone of 2950 Ma shear deformation strikes NE-SW along the north edge of the west-facing panel of the Strelley succession, and this fault takes a turn north to join up with the tectonic breccia zone across which the main part of the Strelley succession faces east against the west-facing Warrawoona Group of the North Pole Dome sequence. As shown in the model of Figure 23, these oppositely-facing sequences of different age most likely represent the opposite limbs of a sheared out synform developed across an originally unconformable relationship, and need not represent a terrane boundary.

Post-2950 Ma deformation

Re-activation of ca. 2950 Ma faults is indicated by the faulted contacts of Fortescue Group remnants. Observations along these faults indicate that many of the remnants contain scree slope deposits at their base, indicating that deposition occurred during faulting, in small extensional basins during rifting. In the map area, and on a regional scale, preservation of Fortescue Group remnants occurs over synclines of the older greenstones, in between granitoid domes. This indicates continued gravitational reactivation of the dome-and-basin map pattern first developed at ca. 3460 Ma in the craton, during successive thermal disturbances.

Summary and conclusions

Five distinct supracrustal assemblages and three generations of granitoid rocks have been identified, as indicated in Table 1. The Warrawoona Group occurs around the northwestern margin of the Shaw batholith, in the North Pole Dome and on the northern limb of the Pilgangoora syncline, where it unconformably overlies the 3515 Ma Coonterunnah succession. Warrawoona Group rocks also probably occur south of the Soanesville syncline, around the Tambourah Dome, and between it and the Shaw Batholith.

Unconformable above the Warrawoona Group is the 3240 Ma Strelley succession, which occurs within the north central part of the Central Pilbara Structural Corridor (CPSC), along the southern limb of the Pilgangoora Syncline, and at the southwestern continuation(?) of the North Pole Dome succession. In this area, the clastic sedimentary rocks of the Gorge Creek Group lie conformably, and in a continuous sequence above, the Strelley succession. A complete section may include all of the following: a basal shelf-type sequence, present on the Yule Batholith margin; mafic- to komatiitic volcanic lavas, present in the Pilgangoora Syncline and possibly also beneath the main part of the Strelley succession; andesitic through dacitic volcanics, best preserved in the main part of the succession; and overlying banded iron formation, conglomerate, sandstone, shale, and mafic volcanic rocks (Honeyeater basalt) of the Gorge Creek Group. The total time span of this sequence is constrained only by the upper and lower limits of dated older (3240 Ma Strelley succession) and younger (ca. 2950 Ma De Grey Group) successions.

A third succession is represented by scattered, isolated remnants of coarse clastic sedimentary rocks of the De Grey Group. Their dispersed occurrence suggests a once more widespread distribution than is presently preserved. This group was deposited during the third set of regionally developed structures, related to sinistral strike-slip shearing at ca. 2950-2930 Ma.

The fifth supracrustal sequence recognised in the map area is the ca. 2760 Ma Fortescue Group, which was not mapped.

Three regionally developed generations of structures have been identified through regional and detailed geological mapping. The earliest generation (ca. 3460 Ma) includes structures related to doming of overlying supracrustal rocks of the Warrawoona Group and Coonterunnah succession, by the intrusion of syn-volcanic granitoid plutons. A migmatitic gneissosity developed in these older plutonic rocks may be related to this event through late-stage partial melting, or may reflect the effects of a slightly younger tectono-magmatic event (<3430 Ma, ≥3400 Ma) of unknown significance.

The second generation of structures is related to granitoid doming through reactivation of pre-existing sialic crust and active magmatic diapirism as a result of regional extension. Doming was coeval with basin formation and the deposition of younger supracrustal sequences, but occurred at different times across the amphibolite-facies mylonite zone developed along the western margin of the Shaw Batholith. Within and to the east of this zone, geochronological evidence from both granitoids and greenstones indicates deformation of greenstones, intrusion and doming of granitoid batholiths, and supracrustal deposition at ca. 3325-3300 Ma. West of this zone, the age of magmatic doming of the Yule Batholith, formation of the Tambourah Dome, and deposition of the volcanic Strelley succession all occurred at 3240 Ma. The precision of the age data (±5-10 Ma) indicates the age difference between the two areas is statistically significant. More geochronological work tied in with detailed structural mapping is required in order to determine whether intermediate ages between these two events occur - thus indicating a lengthy episode of D2 deformation over 3325-3240 Ma, or whether the different ages from east and west of the Shaw Batholith reflect discrete events in the evolution of the Pilbara Craton.

In either case, the style of structures developed during the ca. 3300 and 3240 Ma events is similar, interpreted to be related to the magmatic inflation of pre-existing granitoid domes. Overlying greenstones were shed off the rising domes, and new magmas were erupted at surface over the synforms between the domes. In the case of the Yule Batholith, diapiric re-activation of pre-existing sialic crust was possibly accompanied by the formation of an entire supracrustal sequence along the southern limb of the Pilgangoora syncline, from a basal shelf-type sequence, up through mafic and ultramafic volcanic lavas, to banded iron formation and clastic sediments of the Gorge Creek Group. This very typical Archaean greenstone belt stratigraphy is thought to reflect the effects of a mantle plume. Impingement of the plume into the base of a pre-existing sialic crust caused its re-activation through a sequence of events from core complex style arching and shedding off of dense overburden, to full-blown granitoid diapirism (consisting of magmas derived through melting of the older crust) as the lithostatic load was released through extension. A similar model of formation for the large granitoid domes east of the map areas has been proposed (eg. Collins, 1989).

Super-imposed on these earlier sets of structures after a hiatus of ca. 300 Ma, is a set of discrete shear zones, faults, and large-scale folds related to craton-wide sinistral shear at ca. 2950-2930 Ma. Concentrated within the Central Pilbara Structural Corridor between the Shaw Batholith and North Pole Dome in the east, and the Yule and Carlindi batholiths in the west, these structures overprint and re-orient D2 structures. The CPSC is bound by a single, curviplanar fault in the west, but by a series of unconnected shear splays in the east. Folds related to shearing formed on NE-SW trending axial planes and commonly have sheared out limbs. Shearing caused the originally subhorizontal Strelley Granite laccolith and associated supracrustal rocks to be displaced laterally to the west and north, and to become tilted onto their side within the CPSC. The rocks of the Lalla Rookh basin are interpreted to represent a mini-foreland style basin deposited in advance of the northward-moving Strelley Granite laccolith, rather than a pull-apart basin as previously suggested (see Van Kranendonk & Collins, in press).

Subsequent thermal events imposed on the Pilbara Craton, at ca. 2830 Ma and 2760 Ma, resulted in the further amplification of granitoid domes and greenstone synclines, as indicated by the domainal preservation of the 2760 Ma Fortescue Group over synclines of the older supracrustal rocks in between granitoid domes.

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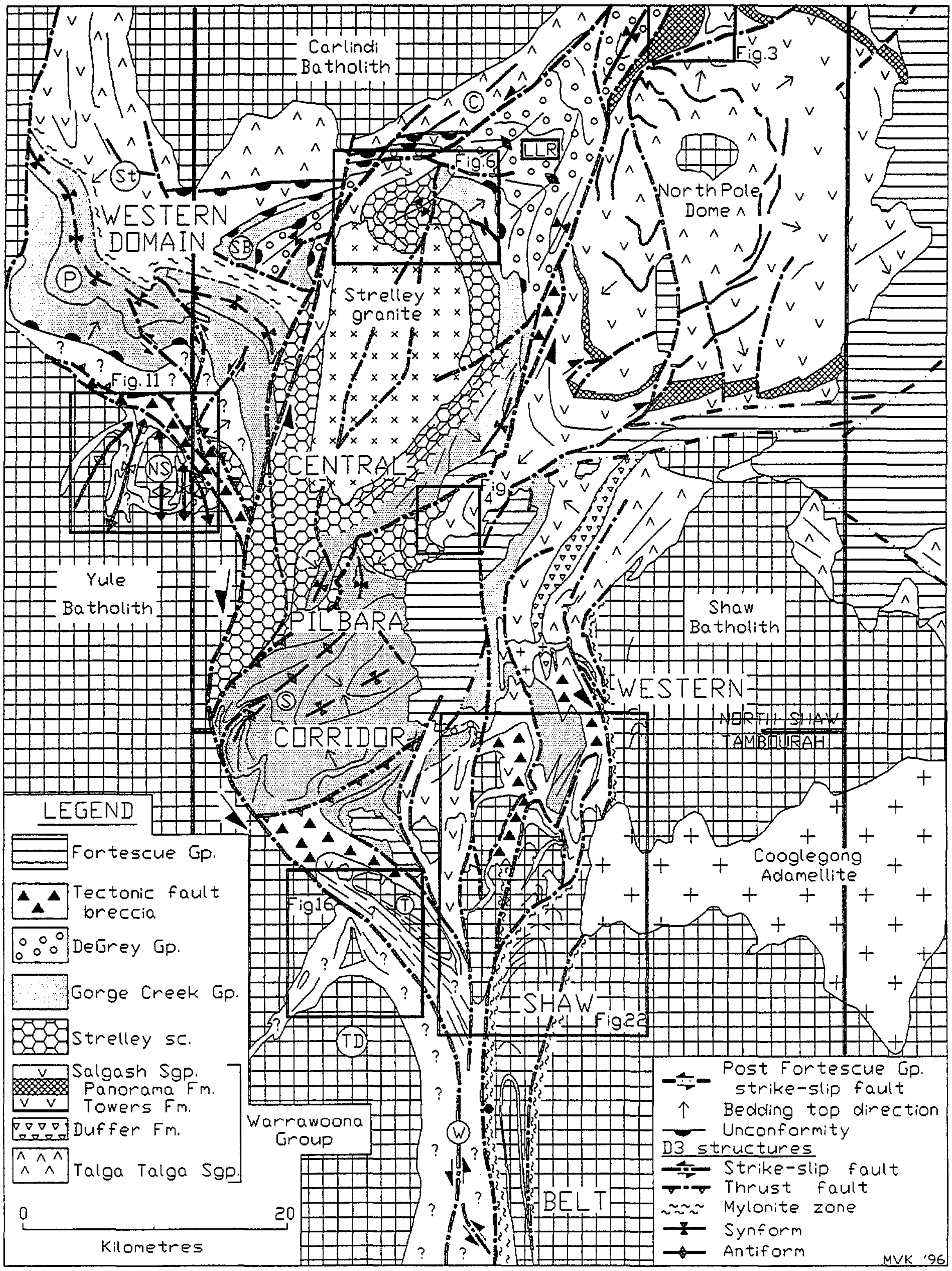


Fig. 1: Geology of the North Shaw and Tambourah map areas (heavy straight lines), showing locations of detailed figures. Grid pattern = 3.45-3.3 Ga granitoid batholiths; x = 3.26 Ga Strelley Granite; + = late tectonic granite intrusion. C = Coonterunah succession; LLR= Lalla Rookh Basin; P = Pilgangoora Belt ("syncline"); S = Soanesville Syncline; Sb = Strelley Block; St = Strelley Belt; T = Tambina Granite; TD = Tambourah Dome; W = Western Shaw greenstone belt.

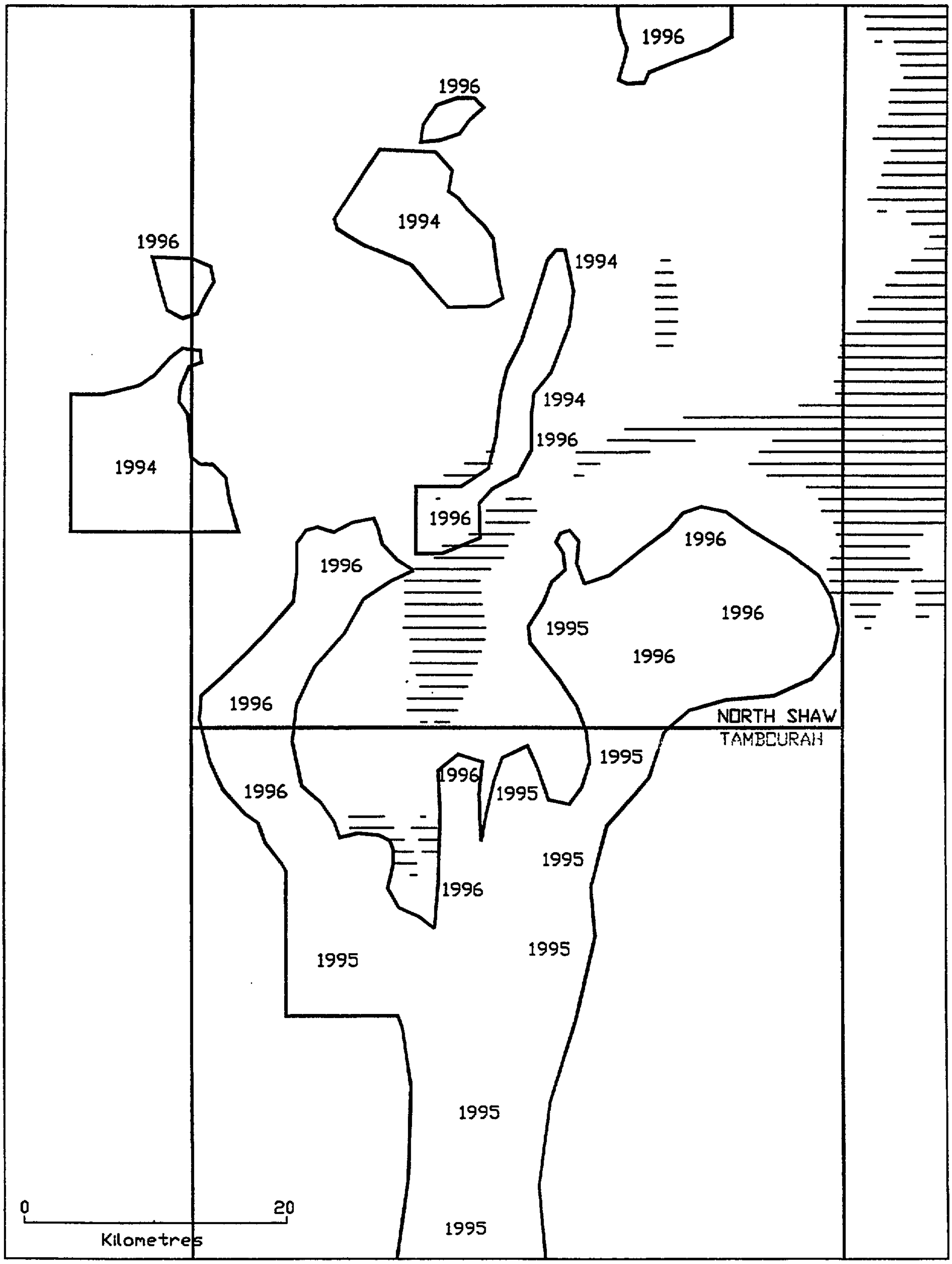


Fig. 2: Areas and year when mapped in the North Shaw and Tambourah map sheets. Ruled areas are those covered by the Fortescue Group.

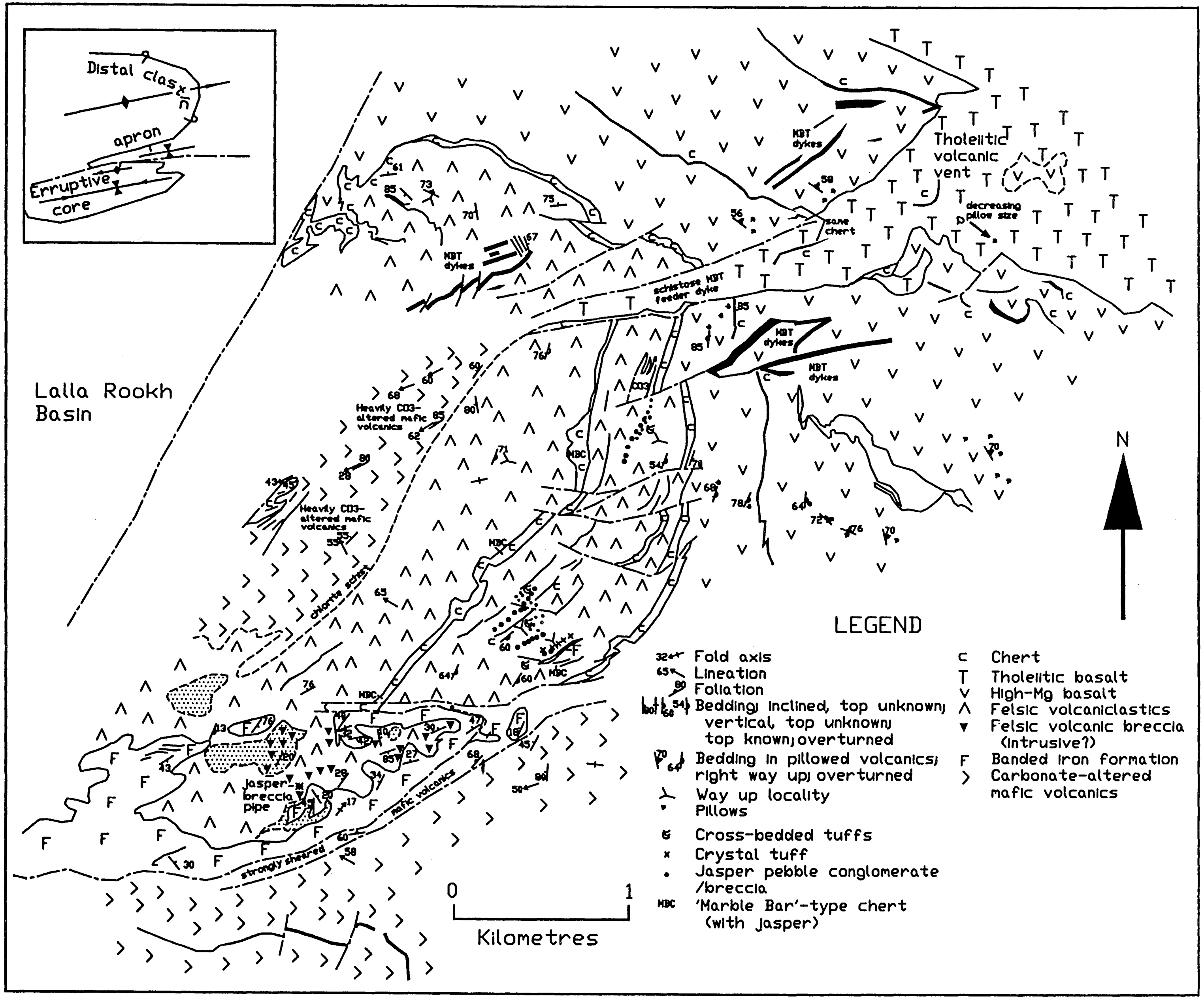


Fig. 3: Geology of the folded felsic volcanic dome along the northern margin of the North Pole Dome. Note the steeply-dipping to overturned nature of the felsic volcaniclastic rocks in the fold structure versus the shallow dips of the iron formation and felsic intrusive breccias in the southwest. Also note the tholeiitic feeder dyke system through the core of the felsic volcaniclastic sequence, that broadens out to an asymmetrical vent from which were extruded pillowed flows. Inset shows the broad structure of the area.

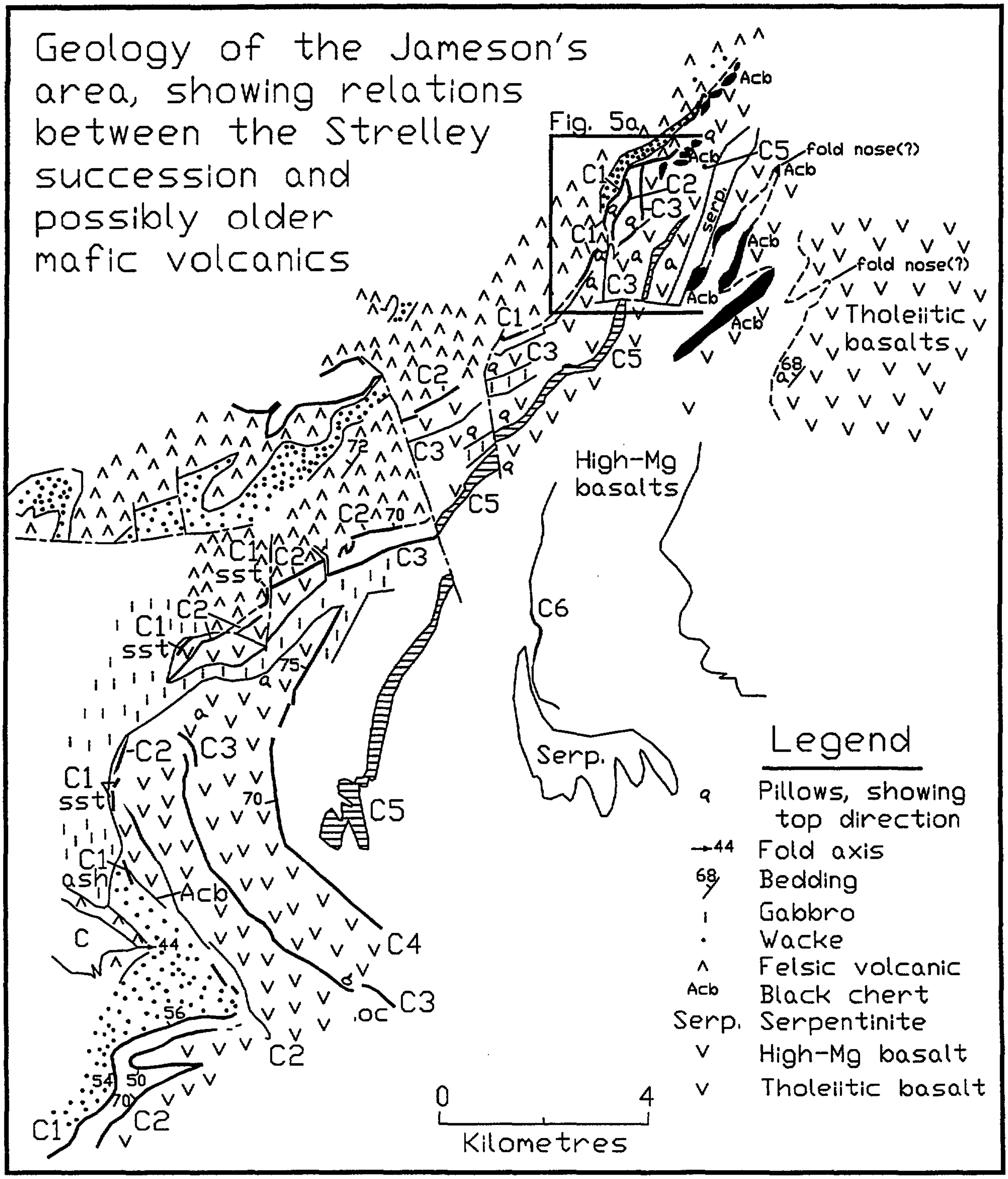


Fig. 4: Geology of the Jameson's prospect area, showing possible unconformable relation between Strelley felsic volcanics and possibly older mafic volcanics and interbedded cherts (Warrawoona Gp?). Chert C1 at the base of the Strelley succession is variable in composition along strike and is discontinuous, having been locally eroded by overlying wacke or felsic volcanic rocks. Note that cherts C2 and C3 in the mafic volcanic rocks are cut out from SW to NE. Location of detailed relations in Fig. 5a indicated by 3-sided polygon.

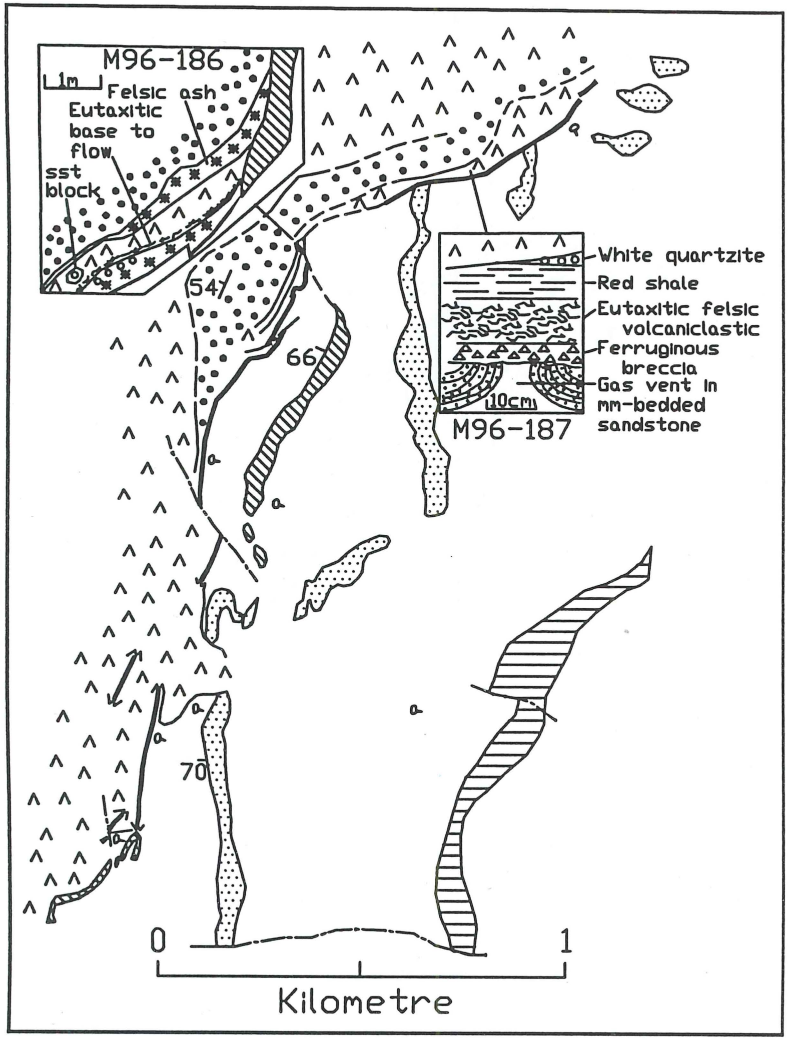


Fig. 5: A) Detailed geology of interpreted unconformity between Strelley felsic volcanics and inerbedded wacke, and older high-Mg basalts and interbedded chert of the ?Warrawoona Group. Horizontal stripes = black and white layered chert C5; Stipple = mm-bedded black, brown and grey chert C3; Diagonal rule = cm-layered grey and blck chert C2; Solid black line = chert C1; ^ = felsic volcanic rocks; = wacke; pillowed flows indicated; w = intrusive contact of felsic volcanics; dash-dot lines = faults.

B) Detailed view, to north, of unconformity at station M96-187 between tilted sandstones of chert C1 and overlying ferruginous breccia. Locality and detailed sketch of relations shown in Fig. 5a.

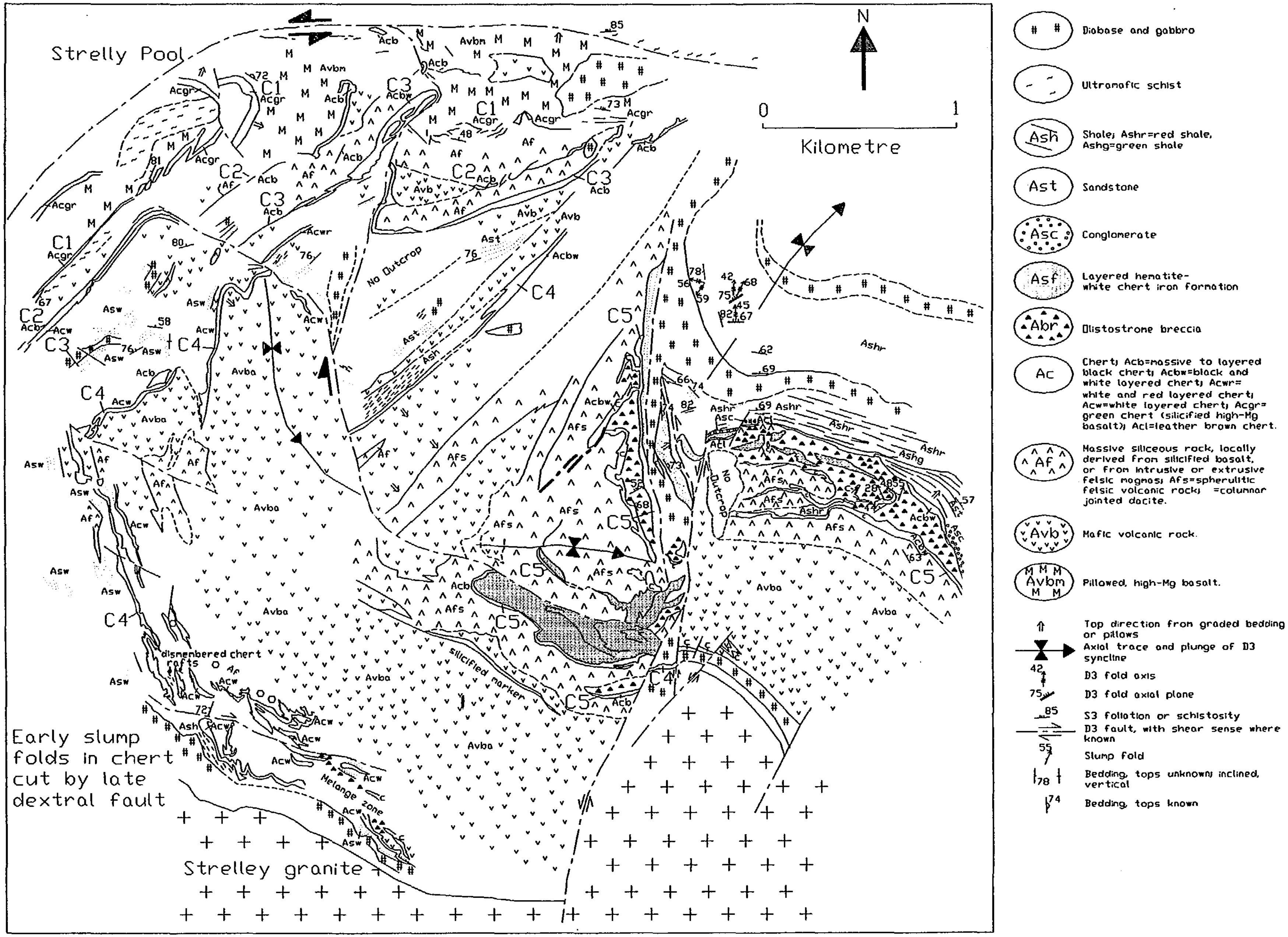


Fig. 6: Detailed geological map of the area north of the Strelley Granite.

Fig. 7: A) Olistostrome breccia north of Sulpher Springs, showing muddy, unbedded matrix and chert clasts.

B) Syn-sedimentary deformation of cm-scale bedding in banded iron formation within the olistostrome breccia north of Sulpher Springs. Rounding of cherty layers occurred prior to total consolidation of the rock.

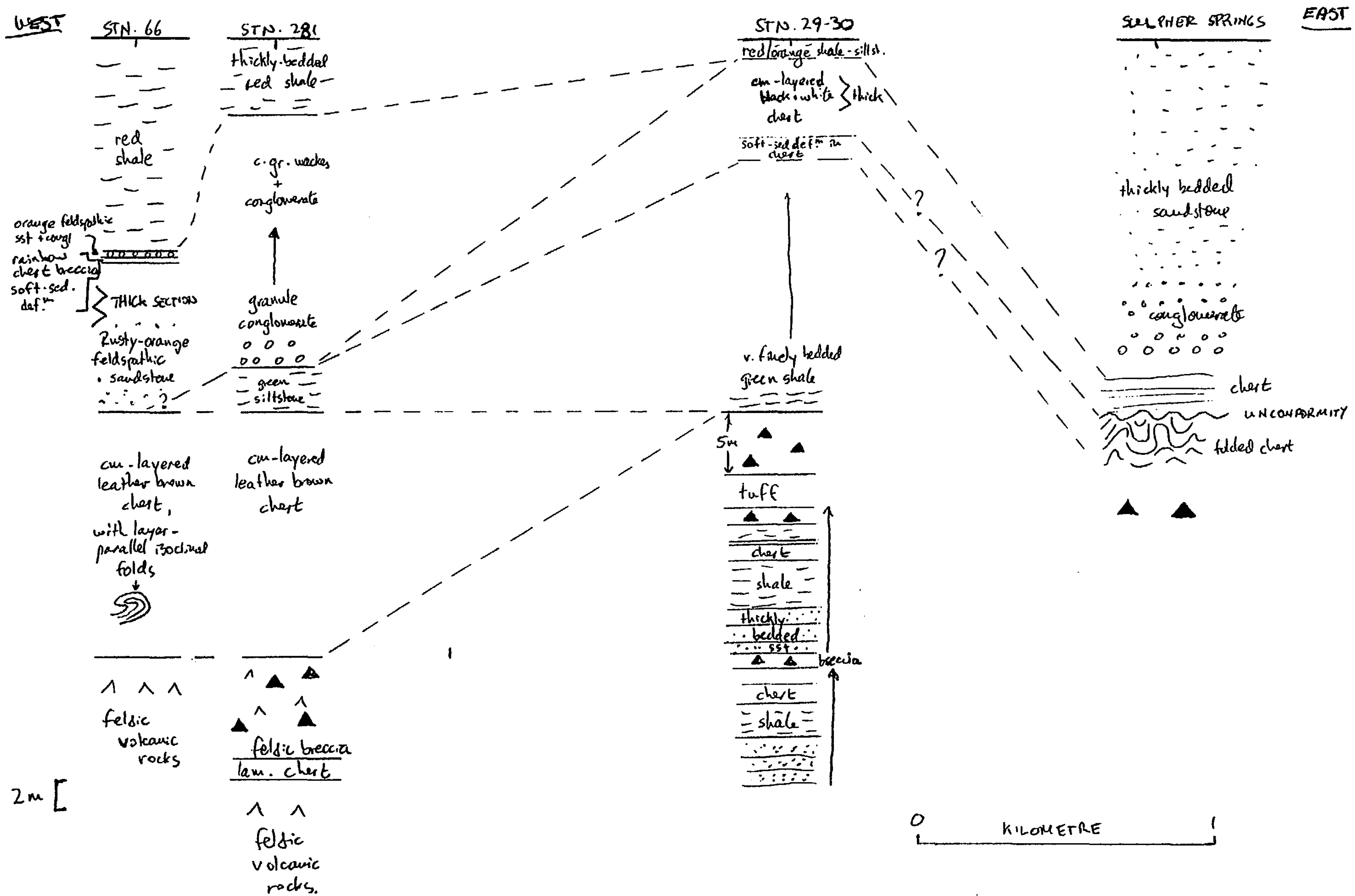
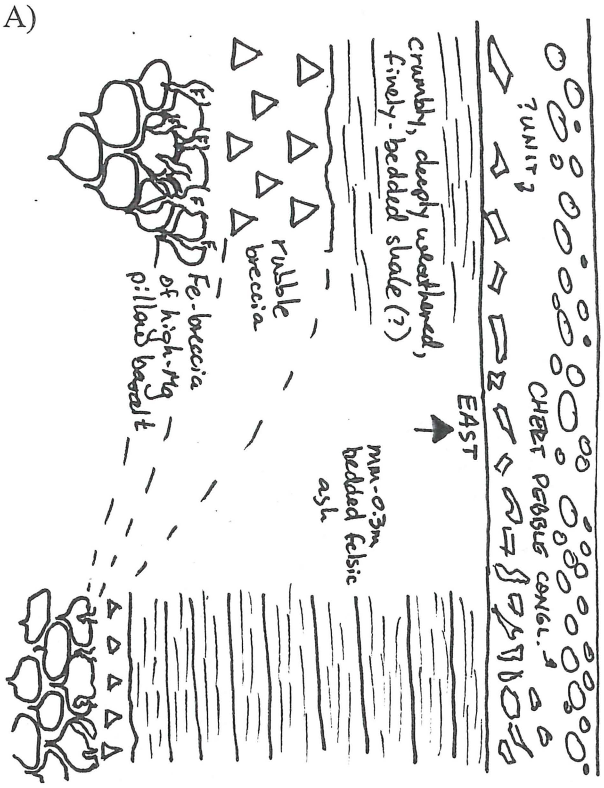


Fig. 8: Stratigraphic sections across the top of the olistostrome breccia above and to the west of the Sulpher Springs gossan. Section station M94-29 and 30 is located immediately east of the N-S fault cutting the Strelley Granite; sections stns. M94-281 and -66 are west of the fault (see Fig. 6).

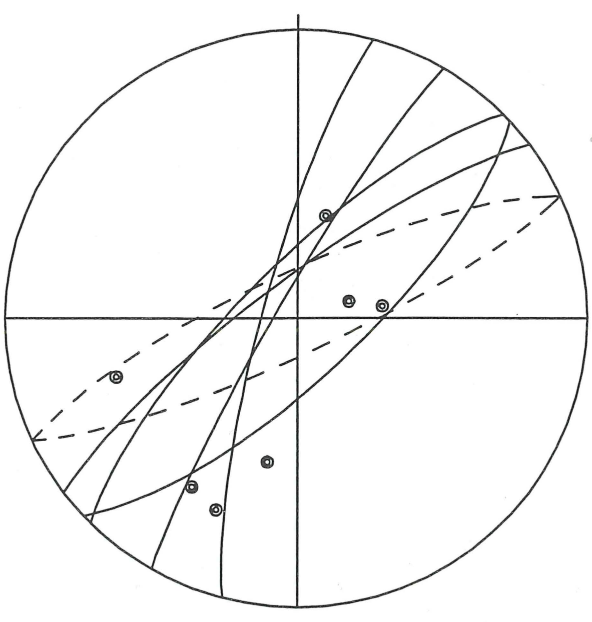
A)



B)

Fig. 9: A) Stratigraphic section across the base of the clastic sequence along the southern limb of the Pilgangoora syncline. B) View south of vertically-dipping chert pebble conglomerate above possible regolith at the base of the clastic Pilgangoora synform, overlying pillowed mafic volcanics. Pencap for scale.

A) Present



B) Backrotated

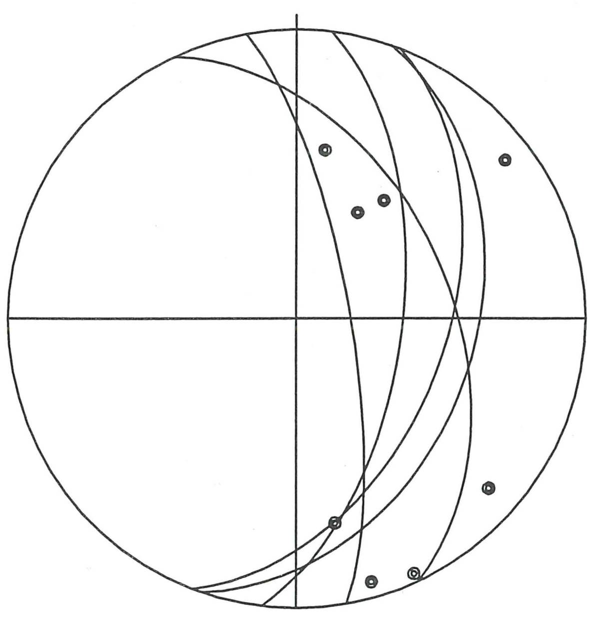


Fig. 10: Equal area stereoplots of structural data from the Coonterunnah succession. A) Measured fabric elements; solid lines = bedding planes, dashed lines = schistosity/fault planes (S1), dots = fold axes. B) Backrotated structural fabric elements of A (except S1 planes), rotated 90° (S-side-up) about an E-W horizontal axis, reflecting restoration of the subvertically-dipping unconformity to the paleohorizontal.

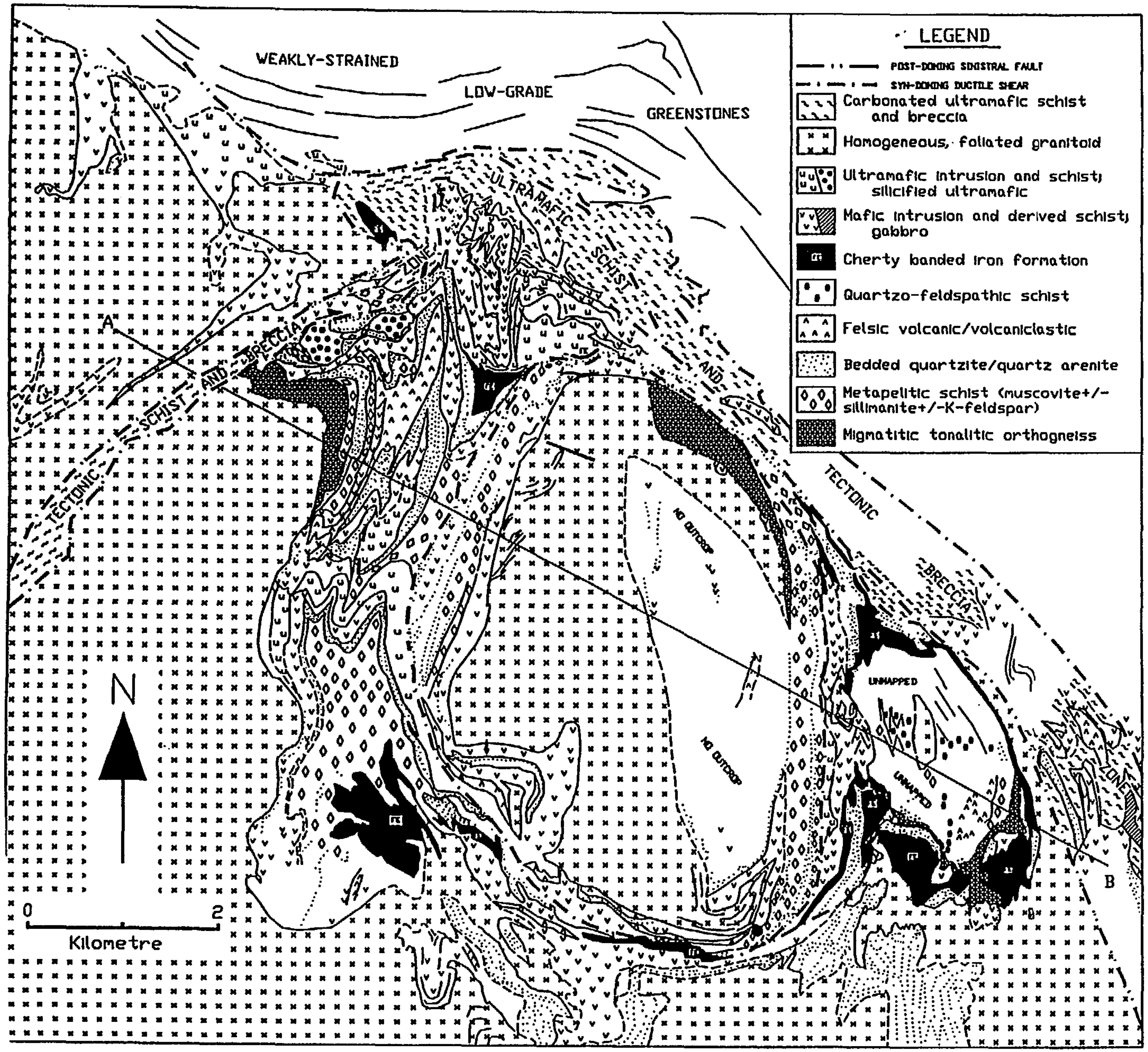


Fig. 11: Geology of the Numerous Scrapes area of the Yule Batholith (for location, see Fig. 1). Line A-B indicates location of cross-section Figure 14.

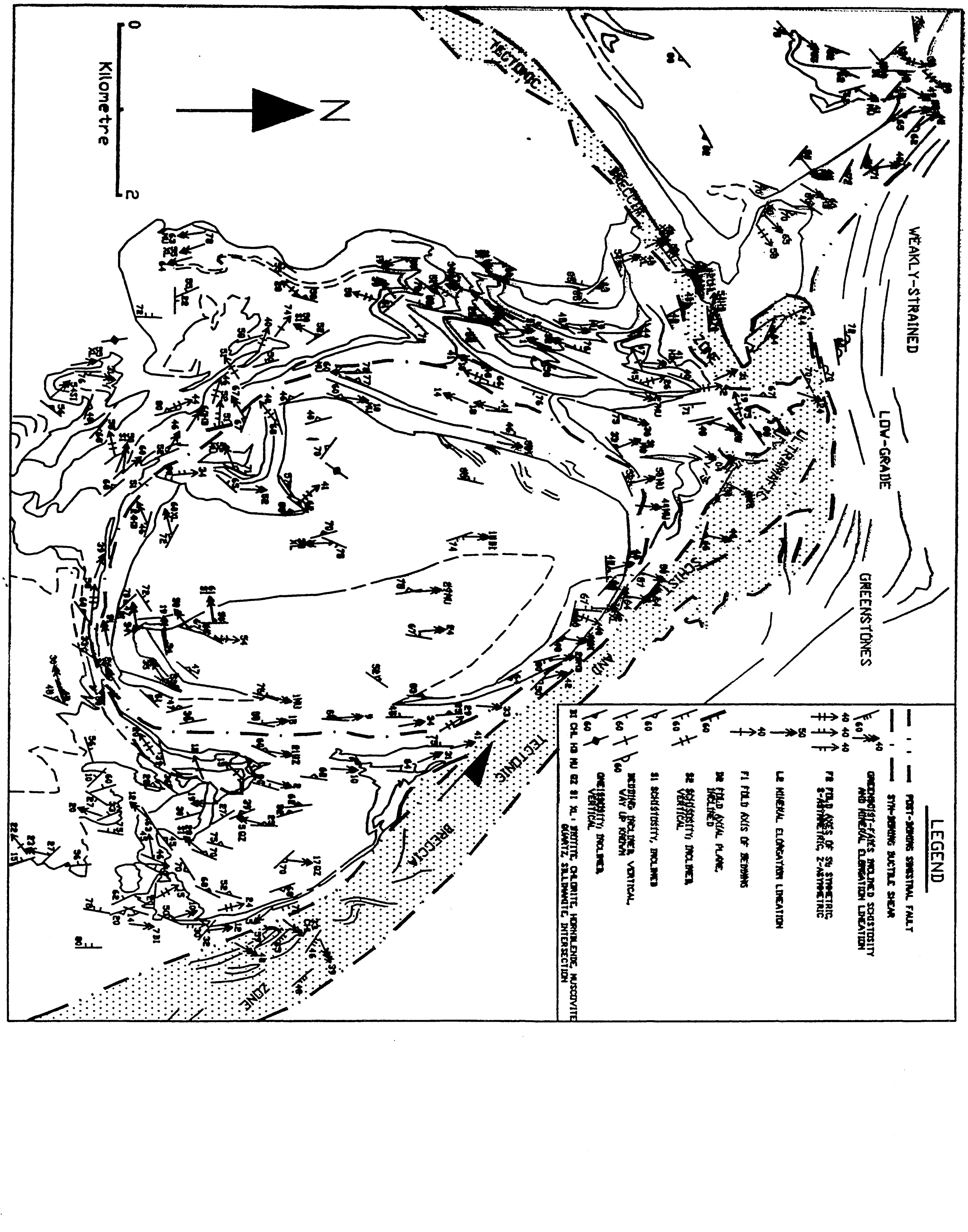


Fig. 12: Structural fabric elements of the Numerous Scrapes area of the Yule Batholith. Note how foliations and lineations wrap around the central elliptical dome, the location of the ring fault around the central dome, and the tectonic schist and megabreccia unit which wraps around the dome-and-basin subdomain of the batholith.

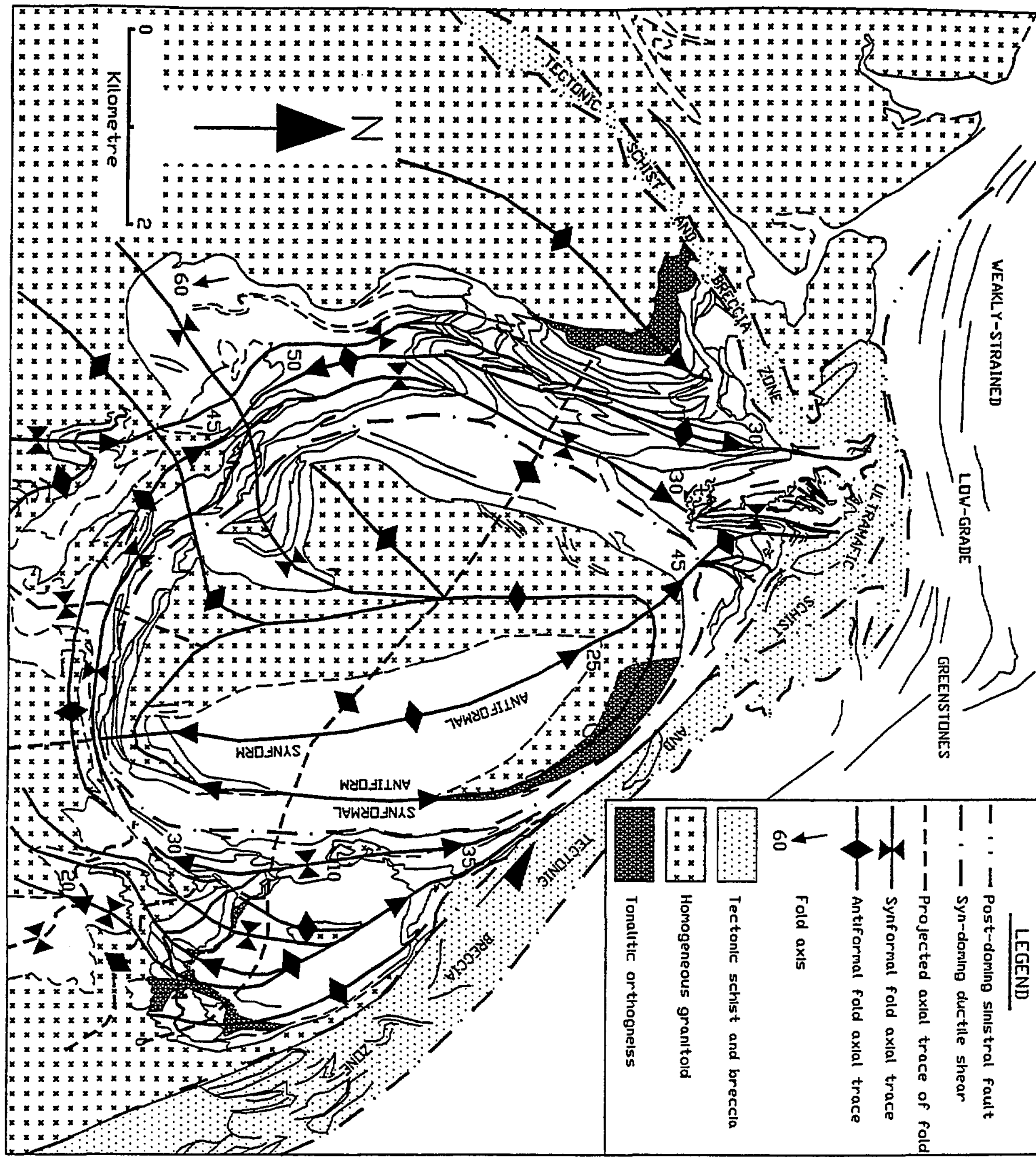


Fig. 13: Fold axial traces in the Numerous Scrapes area of the Yule Batholith. Note that antiformal culminations on promontories in the central dome correspond with sites of preferred granitoid intrusion outside the ring fault.

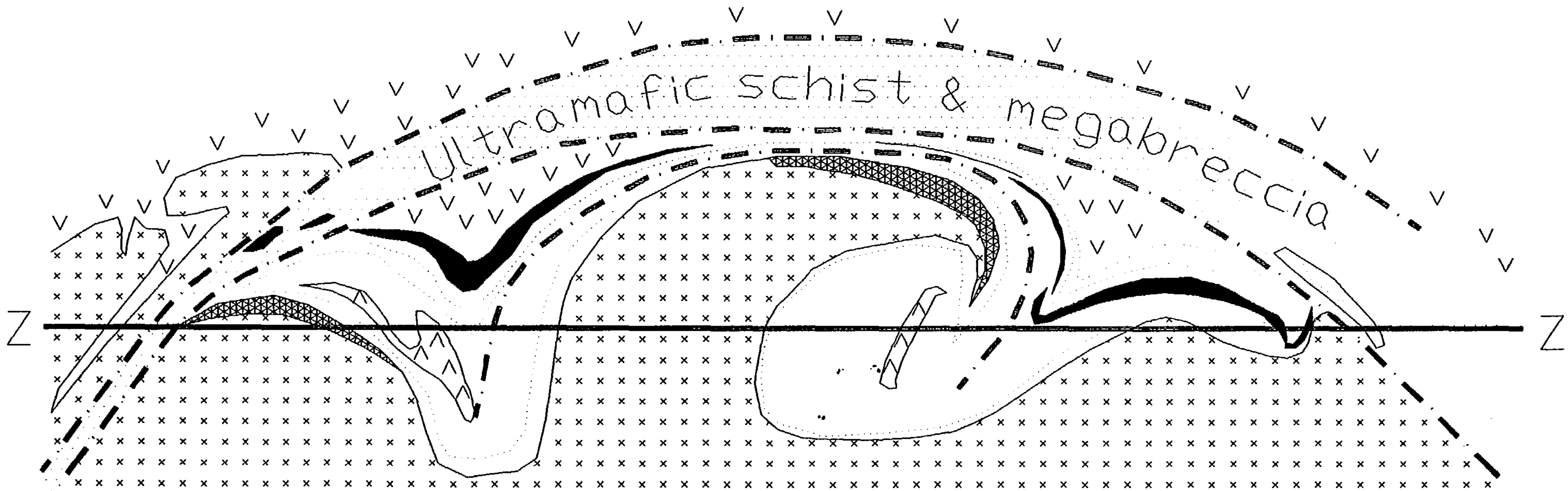
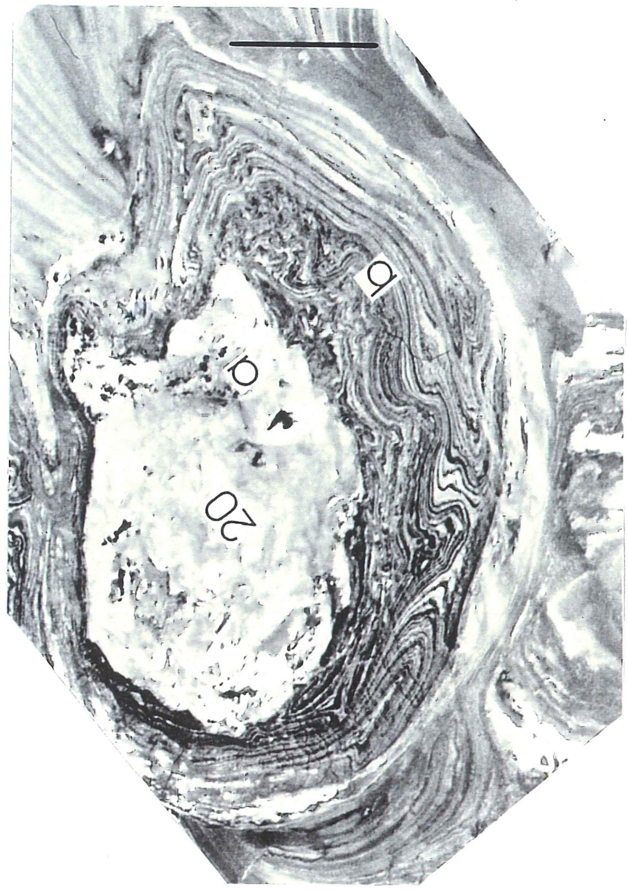


Fig. 14: Cross-section A-B through the Numerous Scrapes area of the Yule Batholith (see Fig. 11 for location). Central dome represents a reclined, refolded synformal antiform bound by the ring fault. Stippled unit represents tectonic schist and megabreccia.

B)



C)

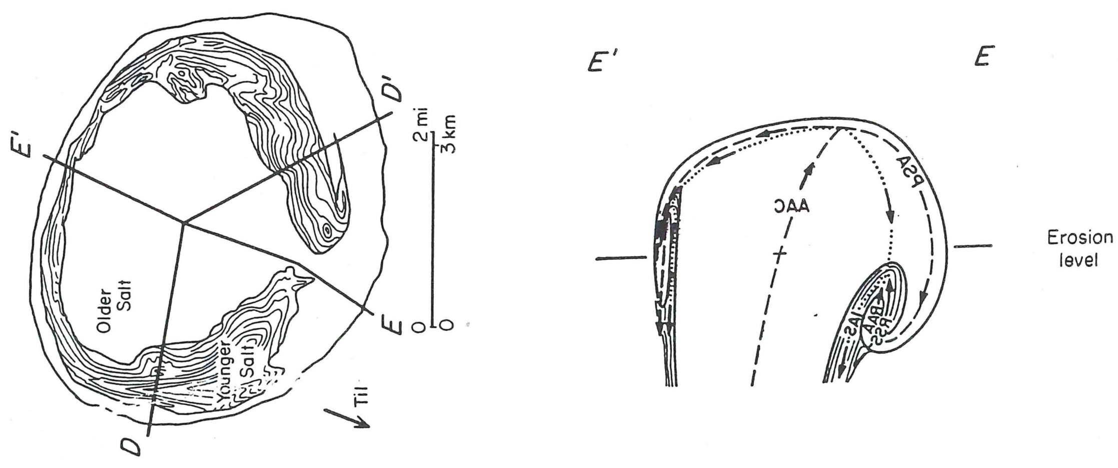


Fig. 15: A) View north of a dyke of homogeneous leucotonalite (3240 Ma) emplaced syn-kinematically into the axial plane of an upright fold of migmatitic orthogneiss (3470 Ma). B) Aerial photograph of salt diapir from the Great Kavir, central Iran, showing hook-shaped folds pointing towards one another on oppositely-plunging fold axes in deformed overburden (striped) to salt core (white). Scale bar = 1 km. From Jackson et al. (1990). C) Plan and cross-sectional view of a salt diapir from the Great Kavir, central Iran, showing reclined asymmetrical structure of diapir. From Jackson et al. (1990).

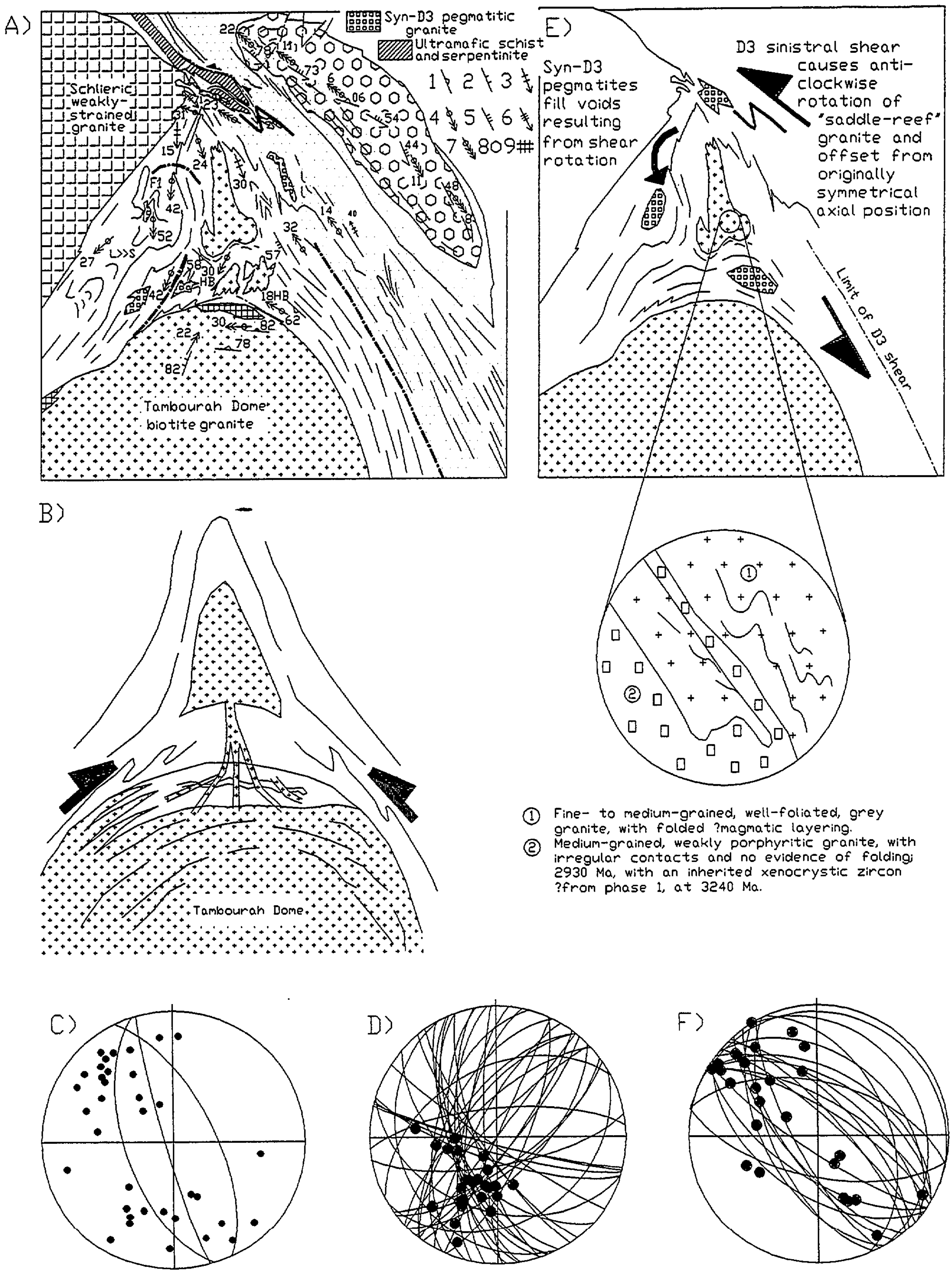


Fig.16: A) Sketch map of the northern closure of the Tambourah Dome: 1. S1 foliation; 2. S2, syn-amphibolite-facies foliation; 3. F2 fold axis; 4. L2 mineral elongation lineation (HB=hornblende); 5. S3 schistosity; 6. F3 fold axis; 7. L3 mineral elongation lineation; 8. Tambina Granite; 9. Migmatitic orthogneisses. B) Sketch of relations in the Tambourah Dome, showing sheeted intrusion of younger, homogeneous granite into remnant of migmatitic orthogneiss, and emplacement of granitoids into the axial plane of upright F2 fold as dykes, feeding saddle reef intrusion in folded greenstones. C) Equal area stereoplot of F2 axial planes and fold axes along the northeastern limb of folded greenstones around the Tambourah Dome. D) Equal area stereoplot of bedding planes and F2 fold axes from the western limb of folded greenstones around the Tambourah Dome. E) Effects of D3 sinistral shear on Tambourah Dome F2 fold, showing how rotation of saddle reef structure caused formation of gaps filled by pegmatitic granite. Inset shows relations of granitoid phases in saddle reef intrusion, and location of sample dated as ca. 2930 Ma. F) S3 and L3 fabric elements from the Tambourah Dome.

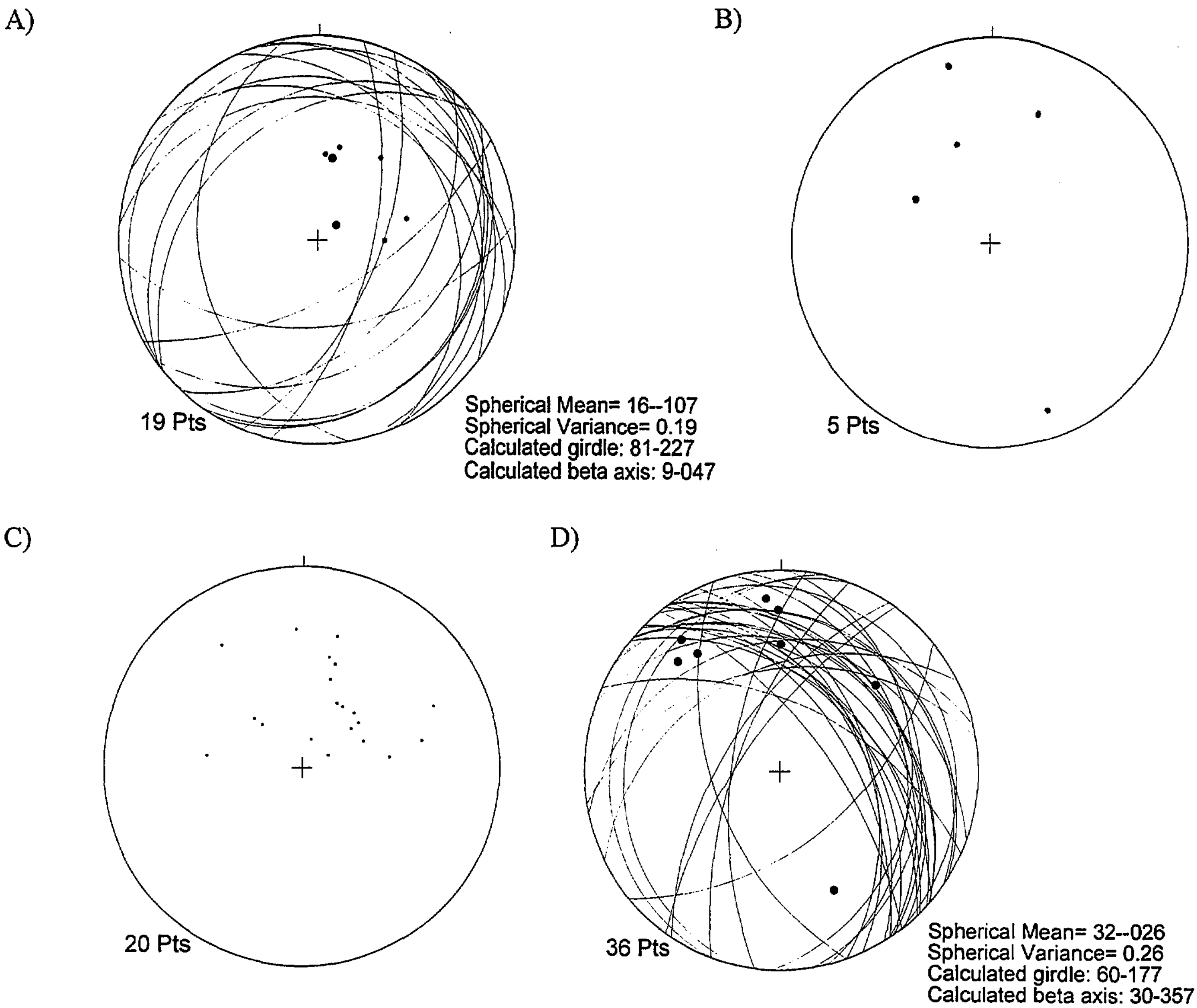


Fig. 17: Equal area stereoplots of structures from the area north of the Strelley Granite: A) Bedding in the southwestern part of the Lalla Rookh Basin, with calculated β-axis indicating average F3 fold axis; large dots = measured F3 fold axes in Lalla Rookh Basin; small dots = F3 fold axes around Sulpher Springs; B) F3 fold axes at Roadmaster; C) Syn-sedimentary fold axes in olistostrome breccia above Sulpher Springs; D) Bedding and pre-D3 (syn-granite emplacement) fold axes in chert at Roadmaster.

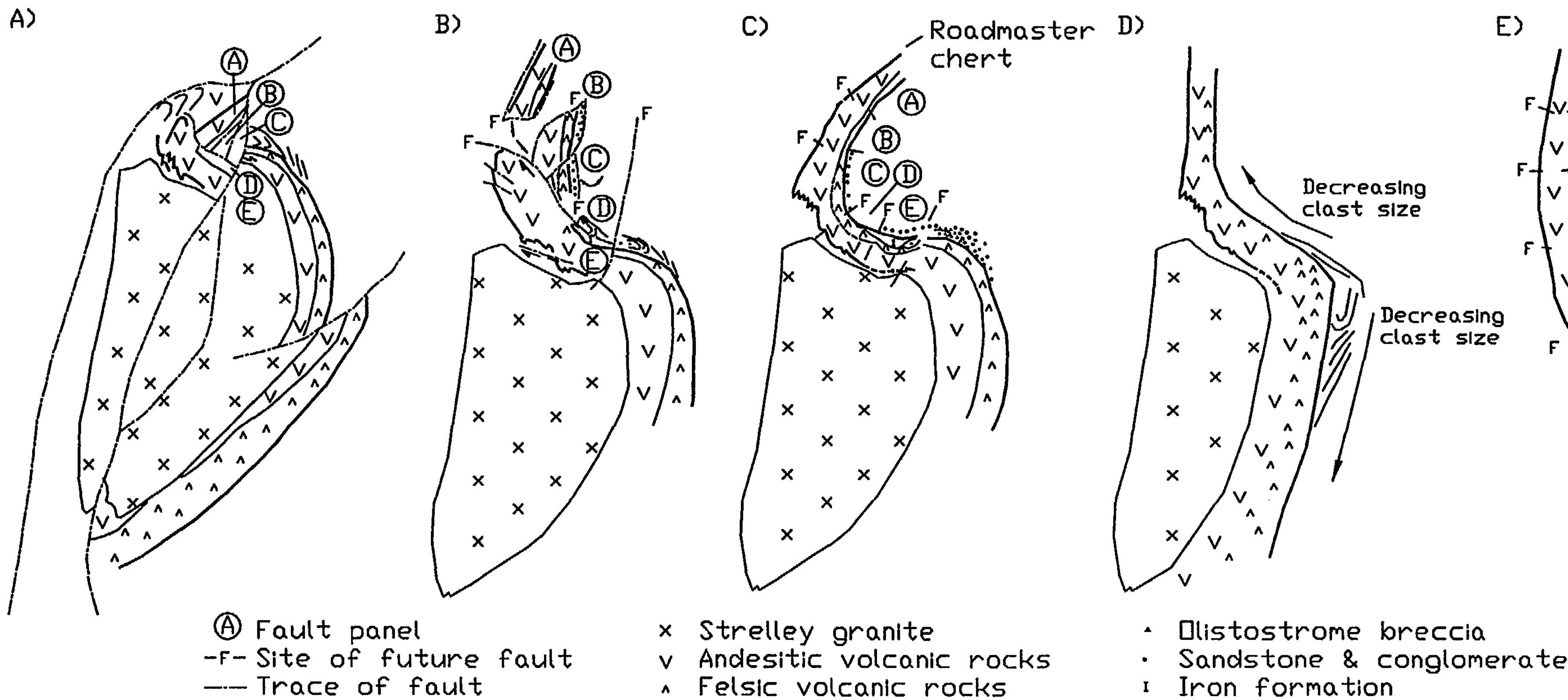


Fig. 18: Deconstruction of structures in the area north of the Strelley Granite. A) Present configuration, dipping 60° east, with identified fault panels A-E indicated; B) Rocks rotated to 90° dip to give true cross-section, faults in Strelley Granite removed, and fault panels in supracrustal rocks partly restored. Arrows demarcate movement path of fault panels A-C; C) Fault panels A-E fully restored. Note that train of "S"-asymmetric folds of chert unit remain. D) Heavily altered, internal phase of Strelley granite removed, deflating the laccolith. Note that folds in E and F now face away from the topographically highest point of the laccolith; E) Granite fully removed and "S"-asymmetric folds fully restored, to reveal relatively flat-lying supracrustal sequence prior to granite intrusion.

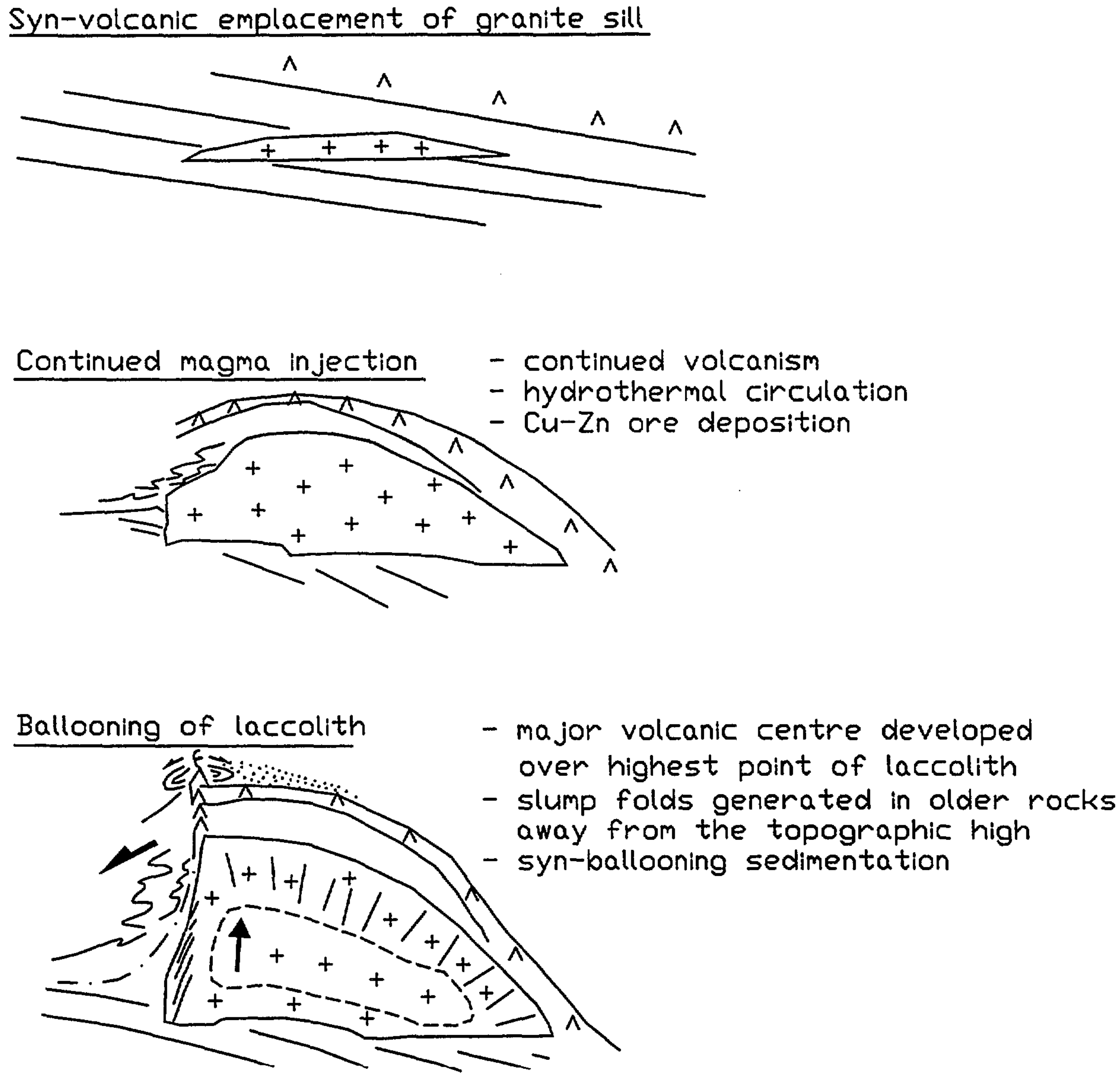


Fig. 19: Schematic evolution of the Strelley Granite laccolith, showing its progressive inflation and resultant formation of slump folds in adjacent chert and syn-sedimentary folds in overlying olistostrome breccia.

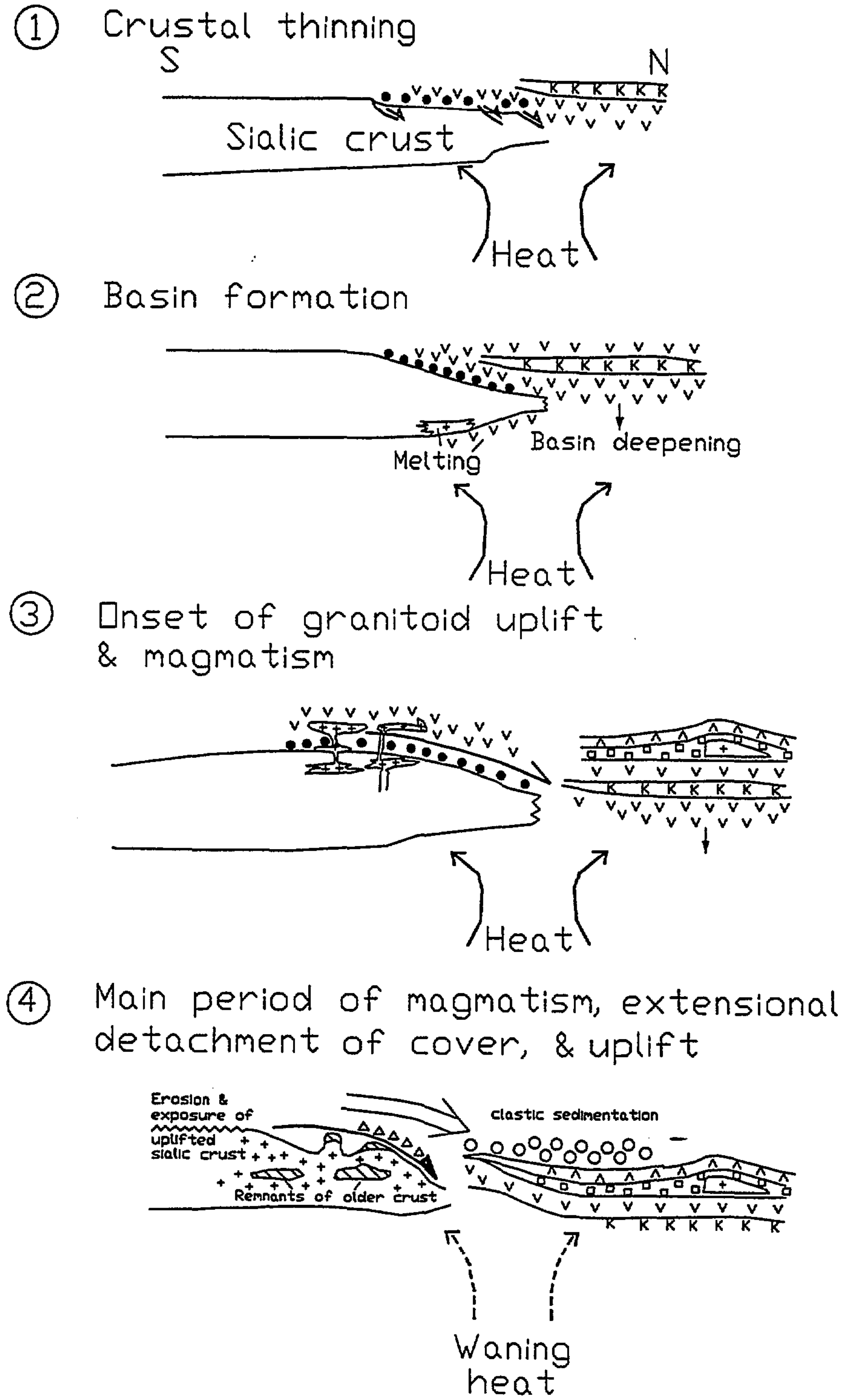


Fig. 20: Schematic evolution of the Yule Batholith and Pilgangoora Synform at ca. 3240 Ma. 1) Onset of extension causes thinning of a pre-existing (ca. 3470 Ma) sialic crust, asymmetrical erosion of overlying Warrawoona Group supracrustal rocks, and deposition of a shelf-type supracrustal sequence (dots). Basin initiation was accompanied by mafic (v) and komatiitic (K) volcanism. 2) Basin deepening during continued extension was accompanied by mafic underplating and onset of granitoid magmatism. 3) Extensional detachment of mafic overburden accompanied by uplift of sialic crust and emplacement of syn-kinematic granitoid sills. Andesitic (□) through felsic (∧) volcanism in basin accompanied by intrusion of the syn-volcanic Strelley Granite laccolith (+). 4) Continued uplift through diapirism causes erosion of granitoid crust which supplies clastic detritus (o) to the Pilgangoora Basin. Older sialic crust largely melted, with only minor relics preserved near surface.

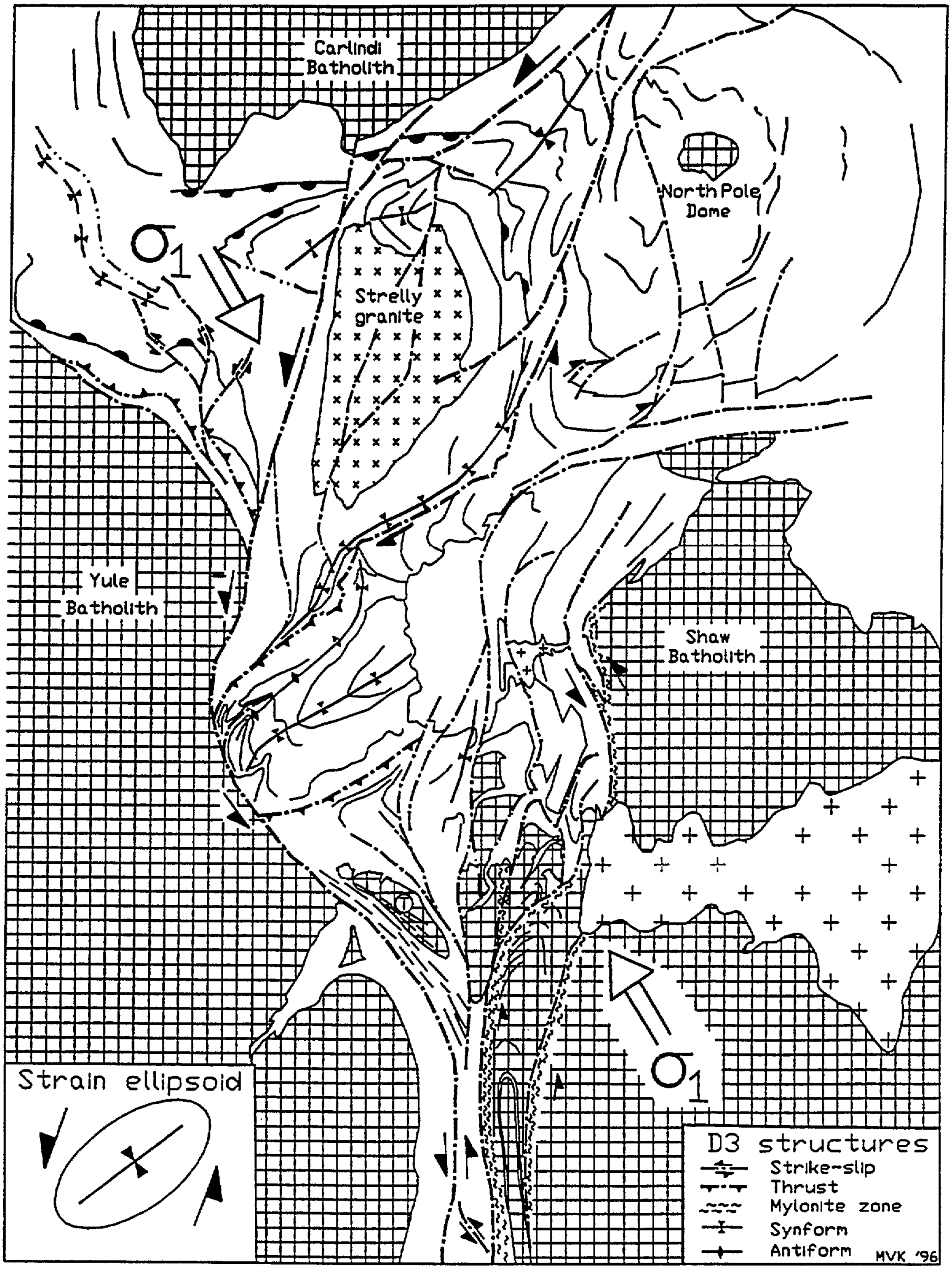


Fig. 21: Principal D3 structures in the Central Pilbara Structural Corridor as a result of NW-SE oriented maximum compressive stress (σ_1). Note that the NE-SW trending F3 fold are bounded by, and do not affect, the western and eastern boundary faults. Folds can be related to sinistral wrenching during D3 deformation, as shown in the inset sketch.

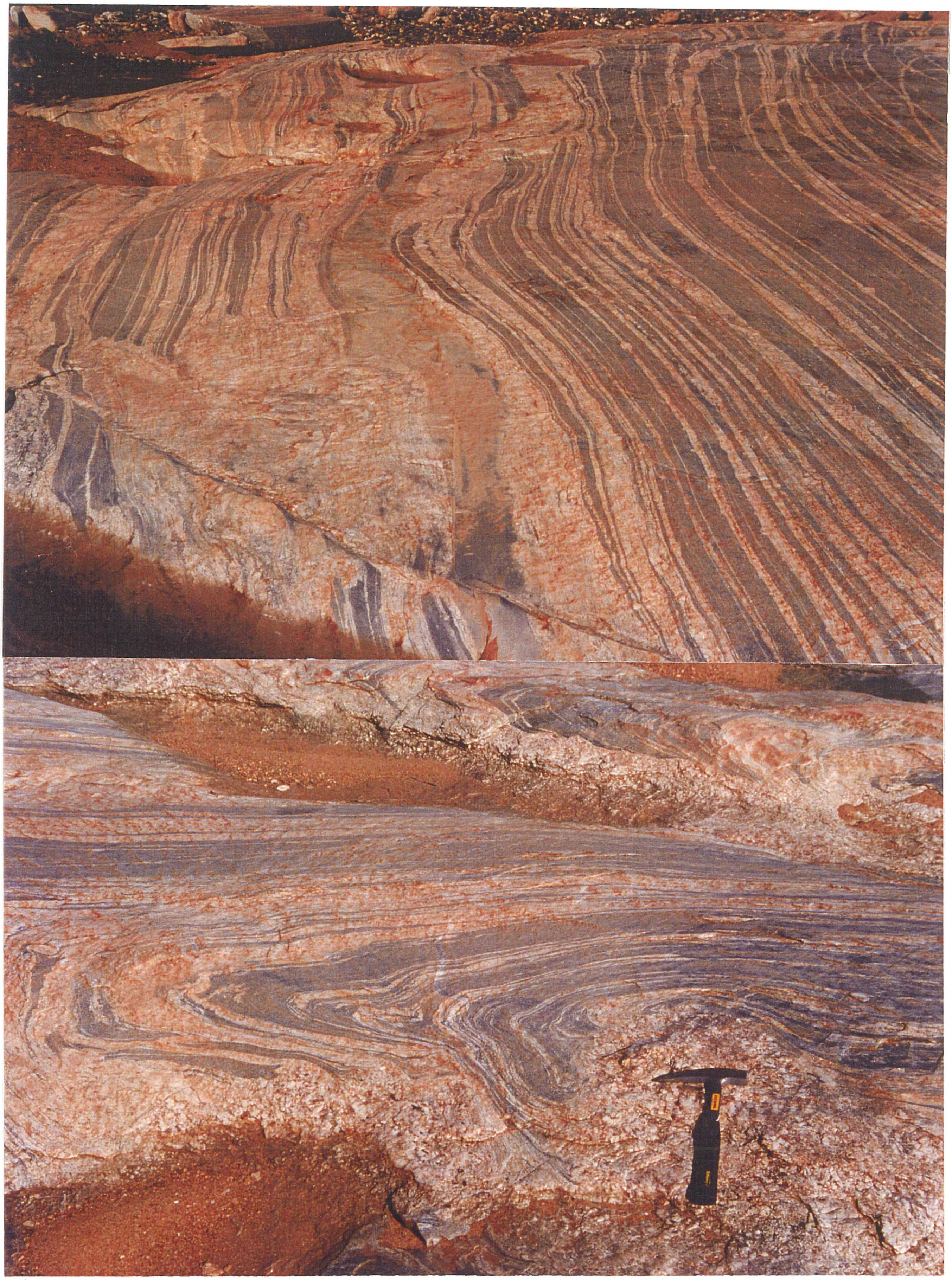


Fig. 22: Features of the amphibolite-facies mylonite zone along the western margin of the Shaw Batholith. A) View south of straight mylonitic orthogneisses, showing two phases of granitic veins; an early, transposed set and a later, coarser-grained set with irregular margins. Zircons from the early leucosome phase yielded ca. 3300 Ma ages, interpreted to be a time of early shear zone development. B) View west of syn-kinematic pegmatite dyke emplaced within the axial plane of an S-asymmetric isoclinal fold related to sinistral shearing. Zircons from one such pegmatitic dyke yielded 2930 Ma ages, interpreted to be the time of late shearing in the zone.

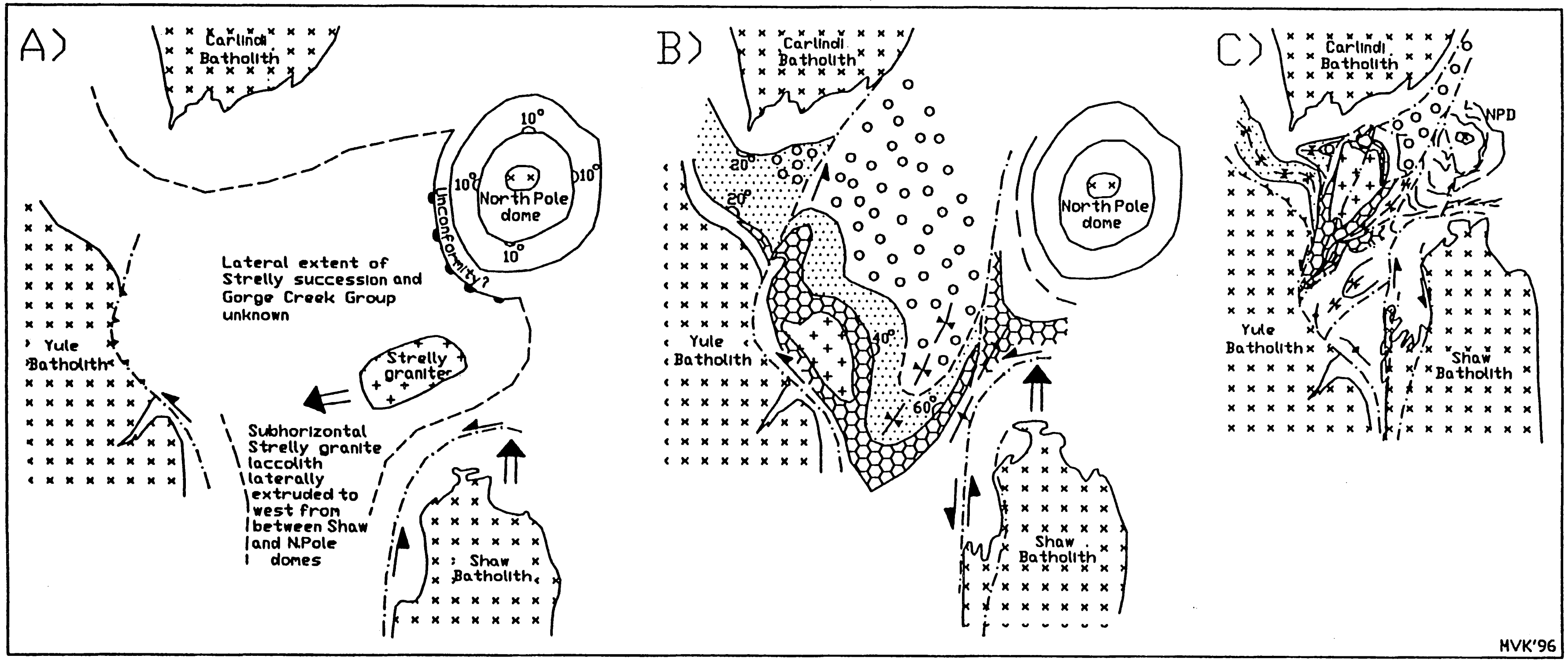


Fig. 23: Structural evolutionary sketch diagram of the D3 structural evolution of the the Central Pilbara Structural Corridor at ca. 2950 Ma. A) Northward movement of the Shaw Batholith causes westerly lateral extrusion of the originally flat-lying, 3.26 Ga Strelley Granite and associated volcanic rocks. Basin formation and deposition of the Gorge Creek Group in the Pilgangoora Syncline may have initiated at, or before, this stage. The relationship of the Strelley succession to the Warrawoona Group of the North Pole Dome is interpreted to be an unconformity (see Figs. 4, 5). B) Continued northerly movement and uplift of the Shaw batholith causes tightening of the grenstones immediately to the northwest, forming a synform into which coarse clastic sediment was deposited (De Grey Group). The Strelley Granite encounters the concave margin of the Yule Batholith and tilts over onto it's side as it starts moving north. C) Present day configuration: from B-C, the Strelley granite travelled north until it impacted the Carlindi Batholith and stopped. Crust following to the south piled up against the Strelley Granite and exposed deeper rocks across thrust faults and tight folds. The De Grey Group continued to accumulate during northerly displacement of the Strelley Granite, with the last components deposited within a strike-slip fault bounded basin (cf. Krapez, 1984). A regionally developed set of NE-trending folds overprints all previous structures. Note in series A-C the change in shape of the NW Shaw Batholith margin and the progressive tightening of the Pilgangoora Synform.

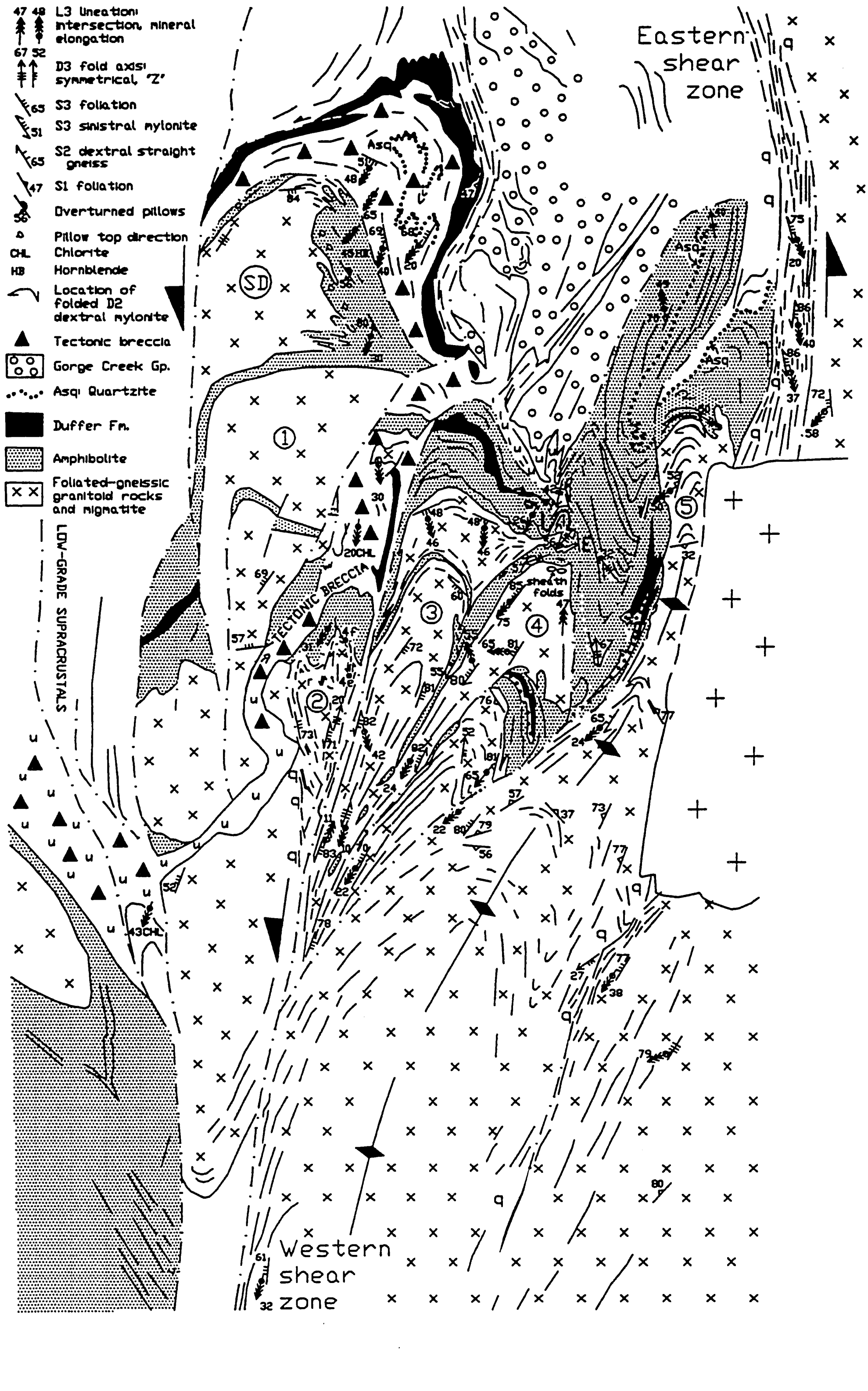


Fig. 24: Geology of the NW Shaw Batholith area (see Fig. 1 for location). Granite lobes 1-5 indicated by circled numbers. SD = South Daltons Pluton; q = quartz vein; u = ultramafic schist. Shaded areas with trend lines only represents greenschist-facies mafic schist. Width of figure is 15km.

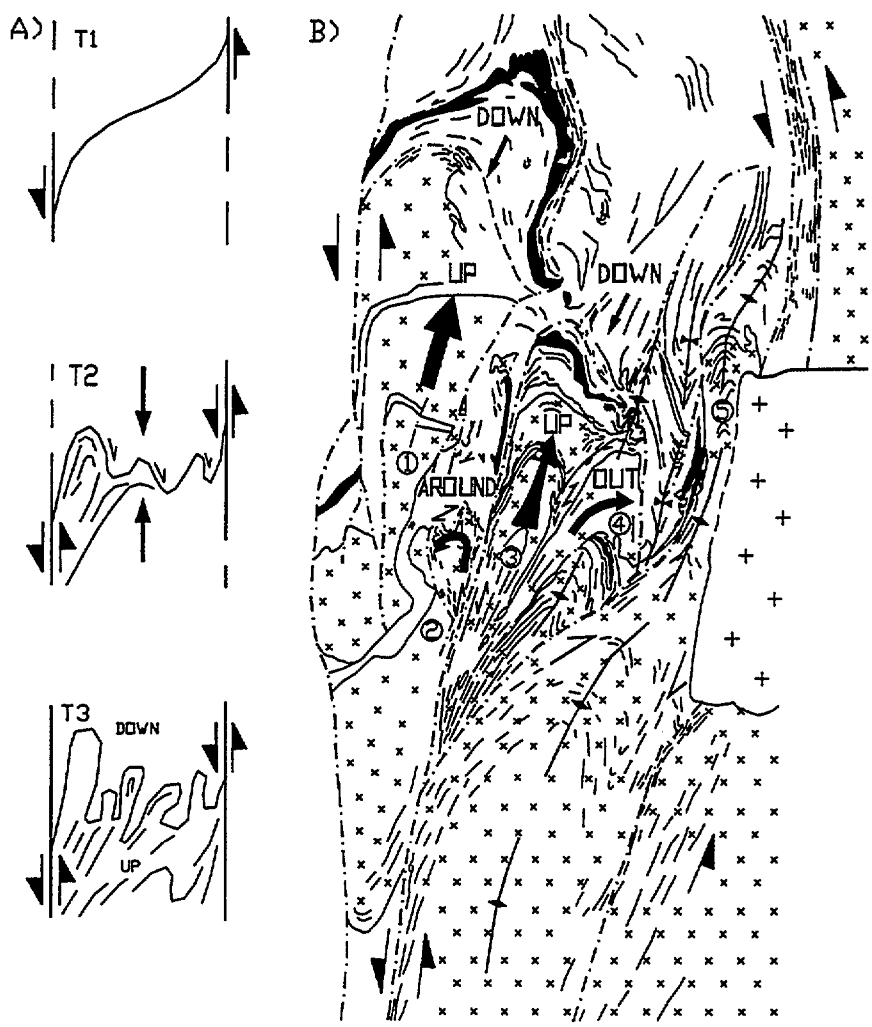


Fig. 25: A) Schematic evolution (time segments T1-T3) of the folds in the NW Shaw Batholith area through formation in a restraining bend between two splays of a sinistral shear system. Same area as Fig. 24 (see Fig. 1 for location). T1: Original flexure in the batholith margin causes deformation focussed along the granite-greenstone contact to localise in two splays. T2: Shortening of the margin flexure between the shear splays initiates fold development, with dextral shear on the eastern limbs of folds as they oversteepen with respect to σ_1. T3: Continued shearing causes fold amplification, uplift of the thickened rocks in the restraining bend, and overturning of greenstones and metamorphic isograds. B) D3 deformation features in the NW Shaw Batholith area. Western and eastern shear zone splays form en-echelon boundaries to a restraining bend (see text). Penetrative D3 strain, characterised by SW-plunging lineations, is focussed in lobes 2-4 and bound to the east by the Emerald Mine shear zone (E). Constriction within the restraining bend of the sinistral shear system causes lobe 2 to rotate counterclockwise to the west, lobe 3 to squeeze up and to the north, and lobe 4 to rotate clockwise to the east.