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Gliding and overthrust nappe tectonics in the Barberton Greenstone Belt

MAARTEN J. DE WIT*

Bernard Price Institute of Geophysical Research, University of the Witwatersrand, Johannesburg 2001, South Africa

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Abstract—Regional recumbent folds, inverted stratigraphy, nappes and olistostromes are described from the southern part of the 3.3–3.5 Ga Barberton Greenstone Belt. Overthrusting of thin rigid silicified slabs with minimum dimensions of 25 km^2 and up to 500 m in thickness, occurred over minimum distances of 86 km. More ductile and coherent units were overfolded up to at least 2 km during their emplacement. The glide planes on which these nappes travelled were zones of high fluid pressures related to hydrothermal fluid circulation patterns, driven by heat sources from igneous intrusions. The upwelling areas of the geothermal convection cells were sites of mud-pools and hydrothermal vents which may mark the trailing edges (pull aparts) of the overthrust units. Progressive silica and carbonate precipitation due to decreasing temperatures, within the zones of fluid migration distant from the areas of high heat flow, probably acted as built-in braking systems below the travelling slabs. Active sedimentation and metasomatism during this tectonism indicates a protracted history for the evolution of the greenstone belt. The recognition of nappe and overthrust tectonics in the Barberton Belt, processes which may have been commonplace in Archaean terrains, necessitates a re-evaluation of the stratigraphy of this belt.

INTRODUCTION

MOST of the current stratigraphic working models of the oldest Archaean greenstone belts (> 3.0 Ga) assume little or no tectonic repetition within sections ranging between 12 and 23 km in thickness (Viljoen & Viljoen 1969a, Anhaeusser 1973, Glikson 1976, Wilson 1979, see Windley 1976 for overview). In a critical reappraisal of these stratigraphic successions, Burke *et al.* (1976) question the validity of this assumption which is incorporated into many tectonic models dealing with the early history of the Earth. They predicted that such unusually thick and undisturbed greenstone-belt successions would in time prove to be paradoxical because of the unrecognised importance of internal tectonic boundaries and possible stratigraphic breaks. One of their points (op cit., p. 125) touches the roots of a very serious problem in Precambrian greenstone belt studies; namely there is *no* stratigraphic fossil control, whilst at best, geochronologic isotope studies in the Archaean have a resolving power of two orders of magnitude too large to replace such a vital control on stratigraphy:

> "... there is the awful lesson of the Col de Genet section in the French Alps where, until foraminifera were collected, a continuous stratigraphic sequence was recognised from Jurassic carbonates through a schistes lustre section with 'interstratified' ophiolites and flysch. Once *fossil* repetition was detected, very thin mylonite zones were found."

The Barberton Greenstone Belt (Fig. 1 inset) is one of the oldest (~3.3–3.5 Ga) preserved belts, where the heterogeneous nature of the deformation, the low grade of regional metamorphism and hence the exceptional state of preservation of many stratigraphic sections along excellent outcrop, offers an opportunitv to evaluate this problem. The stratigraphic succession of the Barberton Greenstone Belt, collectively referred to as the Swaziland Supergroup, consist of an essentially volcanic sequence about 15 km thick (Onverwacht Group) which is in turn overlain by the largely sedimentary Fig Tree and Moodies Groups, of about 2 and 3 km thick, respectively. The general geology and detailed stratigraphy of the belt has been described by many authors, and the reader is referred to Viljoen & Viljoen (1969a), Anhaeusser (1973) and Williams & Furnell (1979) for summaries.

One of the first to document the polyphase and heterogeneous nature of the deformation and strain in the northern part of the Barberton Greenstone Belt was Ramsay (1963). He recognized a period of heterogeneous deformation during which a vertical cleavage developed and which thus involved considerable horizontal shortening. This was later quantified by Gay (1969) and Anhaeusser (1969), and recently elaborated on by Fripp *et al.* (1980). The latter have suggested that the zones of intense deformation are slides or root zones of low-angle thrusts. In the southern part of the belt, Viljoen & Viljoen (1969a,b,1970) recognized vertical schistose zones and heterogeneous cleavage development, which they equated with similar features in the northern part of the belt and related them to granite emplacement. However, they placed little stratigraphic significance on these zones, which was duly criticised (Burke *et al.* 1976, Williams & Furnell 1979, Philpot 1979). In particular, suspected significant tectonic breaks that formed during a period of horizontal shortening have been identified during the course of a recent field mapping programme (Philpot 1979, de Wit *et al.* 1982, in review). Strain analysis in one such vertical schist zone about 5 km west of locality V (Fig. 1, western margin), and about half-way up the stratigraphic succession established by Viljoen & Viljoen (1969a), indicates horizontal shortening in excess of 80% (de Wit & Stern 1980, de Wit *et al.* 1982, in review), which

* Present address (until end 1982): Department of Geological Sciences, Queens University, Kingston, Ontario, Canada K7L 3N6.

was previously attributed to lateral, east–west stratigraphic thinning (Viljoen & Viljoen 1969b).

Significantly, Ramsay's (1963) observations of bedding—cleavage relationships led him to conclude that the above cleavage-forming event post-dated a fabric-free fold forming event. Fripp *et al.* (1980) have revived this conclusion and suggest that it may relate to an early period of recumbent folding and thrusting, since downward-facing structures have been recorded in the northern part of the Barberton Belt (Fripp, personal communications, 1980).

This paper sets out to demonstrate that large-scale horizontal translations, in the form of fold and thrust nappes along glide planes, have drastically affected the stratigraphy in the type sections of the southern part of the greenstone belt, prior to the upright cleavage and associated shortening. In lieu of fine-scale chronological control, the basic tools used to test this were sedimentological and structural in nature. Sedimentary structures used to establish younging directions were grading (specifically of accretionary spheroids, Fig. 2 and Stanistreet *et al.* 1981), ripple drift cross-laminations, cross-bedding and, rarely, convolute bedding, channelling and shelves in hollow layered pillows. At any one locality a high confidence level for younging directions, as indicated on Fig. 1, reflects at least two conforming structures (e.g. grading and cross-bedding) observed together.

FIELD RELATIONSHIPS

Recumbent folds and inverted stratigraphy

Figure 3(a) shows a sectional view of a flat-lying isoclinal fold, whose limbs can be traced over a distance in excess of 2 km (Figs. 1 and 4). Sedimentary structures in the cherts indicate a lower upward-younging and an upper downward-younging limb separated by carbonates and serpentinites which occupy the core of the fold. Figure 5(a), taken at about 60° to the right of Fig. 3 (as indicated on Fig. 4), shows the trend of the fold axis, as well as the refolding of this recumbent fold by open upright folds shown in Fig. 3(b). The orientation of the structural elements are shown on a stereogram (Fig. 5b). The upright folds have a sporadically-developed axial planar cleavage, and present along the limbs are vertical shear zones in which schistose to gneissose fabrics are developed. The two-dimensional plan view of this interference folding (Figs. 1 and 4) compares with theoretical and experimentally produced patterns of refolded recumbent folds (type 2, Ramsay 1967) and can be readily appreciated in the field (Fig. 5). This area will be referred to hereafter as the recumbent fold area.

Figure 1 shows the younging directions in relation to the major upright antiforms and synforms over a larger investigated region. Open upright folds are usually flanked by vertical schist zones. Particularly towards the west, beyond the boundary of Fig. 1 these folds grade into near-isoclinal folds both along and across strike, with concomitant widening of the schist zones and development of a penetrative axial-planar cleavage in the cores of the folds. Younging and bedding–cleavage relationships indicate that the fold at locality C (Fig. 1, central) is an antiformal syncline (that is an antiform in which the sediments young towards the axial surface, i.e. downward; Ramsay 1967, Hobbs *et al.* 1976), whereas at locality D (Fig. 1, north-centre) the structure is the converse of this, a synformal anticline (Fig. 6). Both structures are downward-facing (Shackleton 1958). At these localities, the stratigraphic succession was therefore overturned prior to the upright folding, although at locality D, the plunge of the fold hinge is much closer to vertical and the nearest younging directions are some 100 metres away from the hinge zone of the fold.

Between localities D and F, there is a general convergence of younging directions towards the contacts of the silicic volcanic–plutonic unit in the central region of the area shown in Fig. 1. Consequently, younging directions and bedding–cleavage relationships, i.e. facing directions (Fig. 1) indicate that the entire stratigraphic succession in this central region is probably downward facing. To the north and south of these localities, reliable bedding–cleavage relationships are not abundant, and facing directions cannot always be confidently established. In the vicinity of locality A, the succession is upward facing. Nevertheless, the succession north of locality D, which consists of almost identical northward younging stratigraphic sections, separated by vertical shear zones and mylonites (e.g. at locality E) may, by extrapolation from the recumbent fold area, in parts also be downward facing. Additionally, there may be major tectonic repitition due to later displacements along the vertical schistose shear zones; that is these zones may represent post-recumbent fold slides or thrusts. Many of the schist zones undoubtedly represent rotated and reactivated early dislocation planes associated with the recumbent structures (see below), but this cannot be verified everywhere at present for lack of good stratigraphic control. Similarly, the facing directions of the succession south of locality F are not clear.

In the eastern part of the area south of the silic plutonic–volcanic unit, the NNE-striking chert units and probably the entire Middle Marker are upward facing with regard to the vertical cleavage (Fig. 1). Therefore at least one early large northwest-verging recumbent anticline must have occupied the area between these upward-facing chert units and the similar chert unit which defines the antiformal syncline at locality C. A more complete structural analysis in the intervening area is hampered by the cross-cutting, shallow-level silicic plutons (Fig. 1). Additionally, the age relationship between the early deformation and the silicic igneous rocks are not fully understood. In the northwestern part of Fig. 1, the stratigraphy is also again upward facing.

There is surprisingly little evidence of deformation in the cherts along either limbs or in the serpentinite that occupies the core of the recumbent fold shown in Figs. 3 and 4. Accretionary spheroids are perfectly spherical throughout the chert sections and the serpentinites are rarely schistose. Away from areas of steep dip, where no vertical cleavage is present such as at locality G (Fig. 3b),

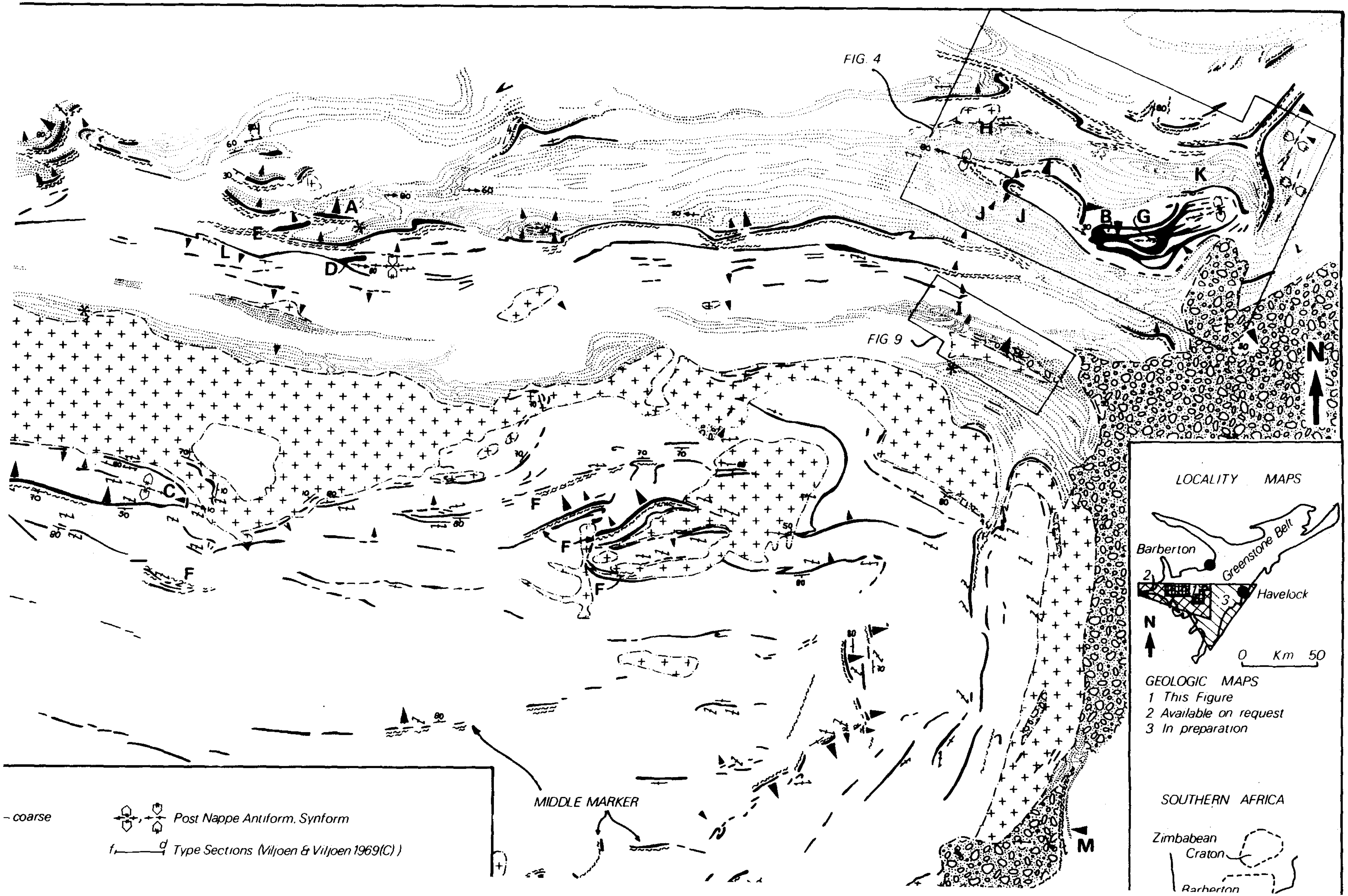


FIG. 4
FIG 9
A
B
C
D
E
F
G
H
I
J
K
L
M
N
MIDDLE MARKER
– coarse
Post Nappe Antiform, Synform
Type Sections (Viljoen & Viljoen 1969(C))
LOCALITY MAPS
Barberton
Greenstone Belt
Havelock
0 Km 50
GEOLOGIC MAPS
1 This Figure
2 Available on request
3 In preparation
SOUTHERN AFRICA
Zimbabean Craton
Barberton

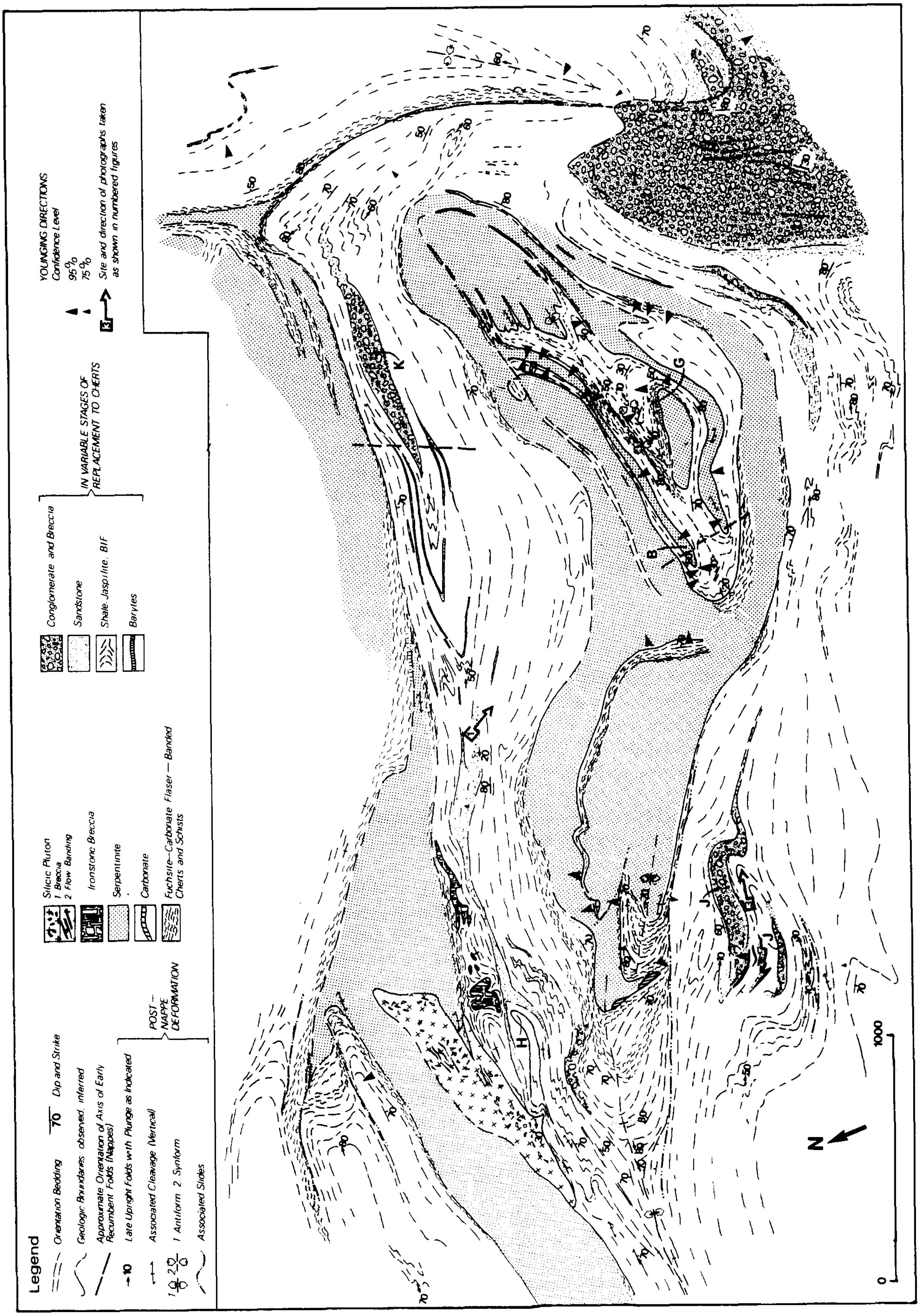


Fig. 4. Detailed geological map of the area referred to in the text as the recumbent fold area. For location see Fig. 1.

horizontally foliated fuchsitic quartz–carbonates occur between the serpentinites and the chert horizon, and have the appearance of recrystallized mylonite (Fig. 7). This quartz–carbonate zone is attenuated and not everywhere present at the chert–serpentinite transitional contact. Early décollement may thus have taken place at the serpentinite–chert interface. Throughout the area similar flaser-banded rocks also commonly occupy the transition between serpentinite and overlying sediments. These rocks define sections up to 50–100 m thick. In places they contain thin breccias with silicified or carbonated clasts of mafic, ultramafic and clastic origin, in various stages of deformation, giving the appearance of melange (*sensu* Hsü 1973). In many places the internal fabric of these units is clearly folded by the upright folds (Figs. 1 and 4).

Vestiges of isoclinal recumbent folds of chert layers in the sediments structurally below the recumbent fold area often show little sign of lithologic thinning over extensive strike distances. For example at locality H (Fig. 4), the limbs of such a fold can be traced over more than 300 m along strike. Figure 8 illustrates this style of deformation with some small scale fold nappes from these sediments. It appears that these rocks must have deformed in a ductile manner, yet the cherts were competent enough to behave coherently. This in turn suggests that they may have been 'set gels' in an unconsolidated ferruginous mud matrix.

Early thrusts and/or glide planes

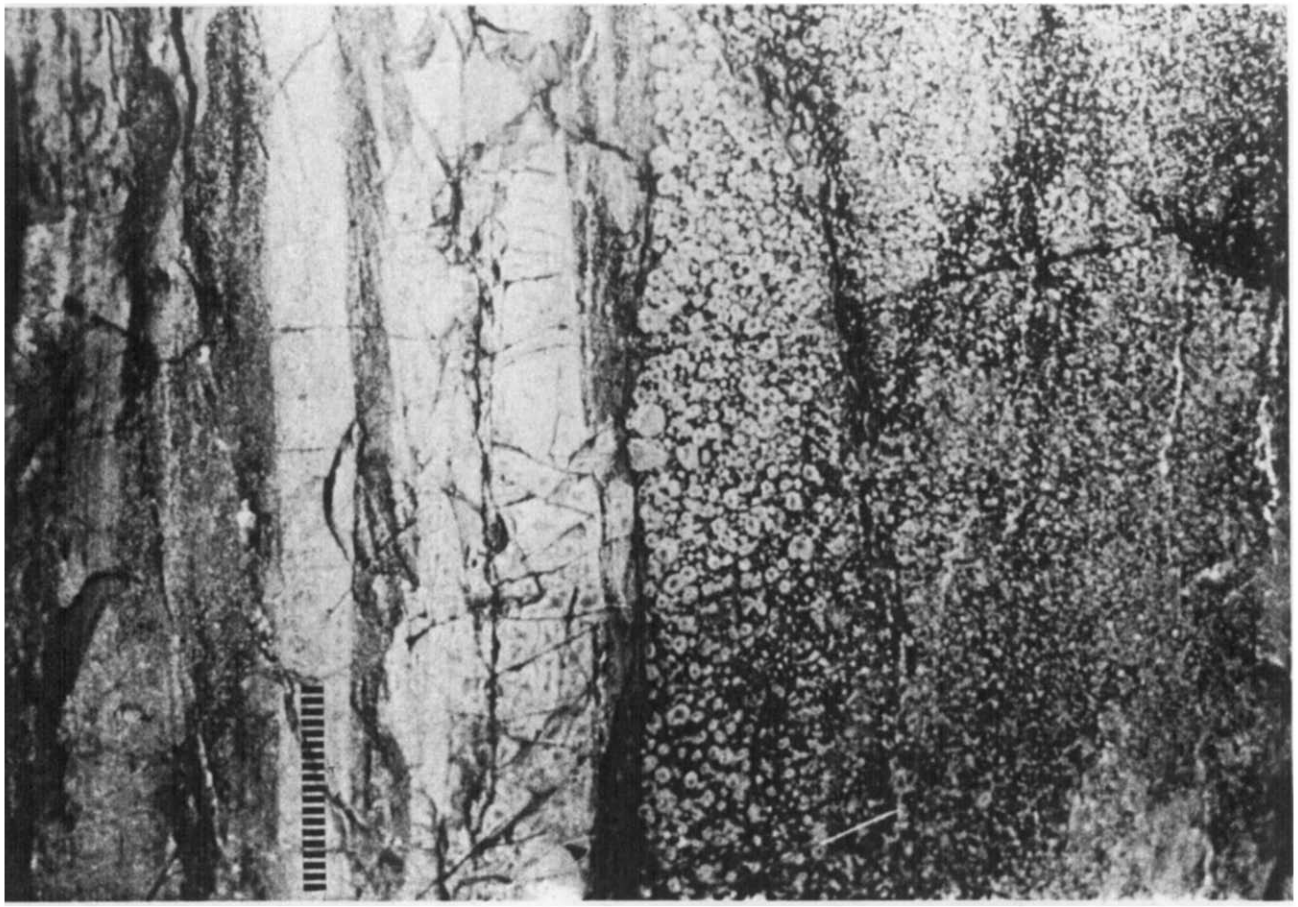
Several sedimentological investigations have documented different modes of deposition for the sediments within and overlying the pillow lavas of the Onverwacht Group (Kuenen 1963, Ramsay 1963, Condie *et al.* 1970, Heinrichs & Reimer 1977, Lowe & Knauth 1977, Eriksson 1980a, b). These studies have clearly revealed both relatively deep and shallow water sedimentation. To reconcile the coexistence of these depositional environments, Heinrichs & Reimer (1977) invoked the presence of a submarine ridge, which they called the "proto-Inyoka zone", separating a shallow marine shelf in the south-eastern half of the belt from a deeper-water basin to the north-west. Eriksson (1980a, b) further developed this into a model of north-westerly prograding submarine fans ahead of prograding deltas, analogous to a stable Atlantic-type continental-ocean margin.

Recent sedimentological studies, within the area mapped, have also revealed these different modes of deposition from the same general stratigraphic sections. A detailed sedimentological analysis from a continuous section exposed by the Msauli River (locality A, Fig. 1) confirms a juxtaposition of deep water and subaerial deposits. North of an approximately 10-m thick shear zone of fuchsitic gneisses and mylonites, Stanistreet *et al.* (1981) have documented a 20 m sequence of graded units, ranging between 2 and 85 cm thick, with perfect Bouma sequences and erosional bases in accordance with deposition from turbidity currents *sensu-stricto*, with north-easterly paleocurrent directions. They interpret these sediments to have been deposited at oceanic depths of at least 1500 m. Within the sediments, some 20 m south of the shear zone, de Wit *et al.* (1981 in review) have described mud-pool structures indicating deposition (or reworking) of the sediments in a subaerial geothermal environment. Such juxtaposition can be interpreted by invoking rapid vertical basinal oscillations or through tectonic superposition of the two sequences during horizontal translations along thrusts or glide planes. The presence of a substantial shear zone between these units, together with the fact that the sediments up to 30 m below the subaerial deposits are again similar to those above the shear zone, clearly supports the translation model, although both processes may have been operative. Accordingly, it is very likely that tectonic repetition along thrusts and/or glide surfaces has occurred, and the succession contains several allochthonous units.

Within the sediments immediately surrounding and structurally underlying the recumbent fold area, (Fig. 4), the sedimentary structures indicate a shallow water to subaerial environment of deposition (Heinrichs & Reimer 1977, de Wit *et al.* 1981 in review), compatible with the non-graded and poorly sorted accretionary spheroids in the associated cherts. By contrast the accretionary spheroids in the cherts of the recumbent fold itself are graded and units display the same deep-water depositional features as described by Stanistreet *et al.* (1981) for locality A. These cherts were therefore derived from a relatively deep-water environment and emplaced into a shallow water to subaerial depository. Accordingly the recumbent fold is allochthonous and a nappe. This bears resemblance to documented nappe emplacement in several Phanerozoic orogenic belts such as the Appalachians (Bird & Dewey 1970, Stevens 1970, Dott 1978), the Scandinavian Caledonides (Gee 1978) and the Alps (Debelmas & Kerckhove 1973, Dewey *et al.* 1973, Carmignani *et al.* 1977, Kligfield 1979).

Olistostromes

At locality I (Figs. 1 and 9) a large-scale isoclinal fold, similar to that shown in Fig. 3, occurs within a section of coarse clastics overlying thinly layered ferruginous cherts. The latter contain structures indicating a shallow water to subaerial environment of deposition (de Wit *et al.* 1981 in review). The layer defining the fold is composed entirely of a coarse chaotic breccia (Fig. 10a). The breccia clasts are only of the type derived from the underlying ferruginous cherts. At the contact, the banded cherts are cross-cut disconformably by the breccias and become progressively folded into the breccia layer (Fig. 9). The folding was brittle and produced the clasts that make up the overlying breccia bed. These features indicate a large-scale slump or current rip-up structure, and the breccias can be classified as an olistostrome, that is materials derived from other formations in the same sedimentary basin (Elter & Trevisan, 1973). Further east along strike, the ferruginous shales grade into conglomerates and planar-crossbedded sandstones and the succession is conformable. Some steep foresets have incipient slump structures developed on them. Palaeocurrents, as measured from rotated planar cross-bedding, were towards the north-west.



← GRADED UNIT → | CROSS LAMINATED UNIT → |

Fig. 2. Graded accretionary spheroids overlying ripple drift unit. From turbidite sequence at locality A, Fig. 1. Similar graded units have been used in the recumbent fold area (Figs. 1 and 4) to determine younging directions. Scale bar = 10 cm.

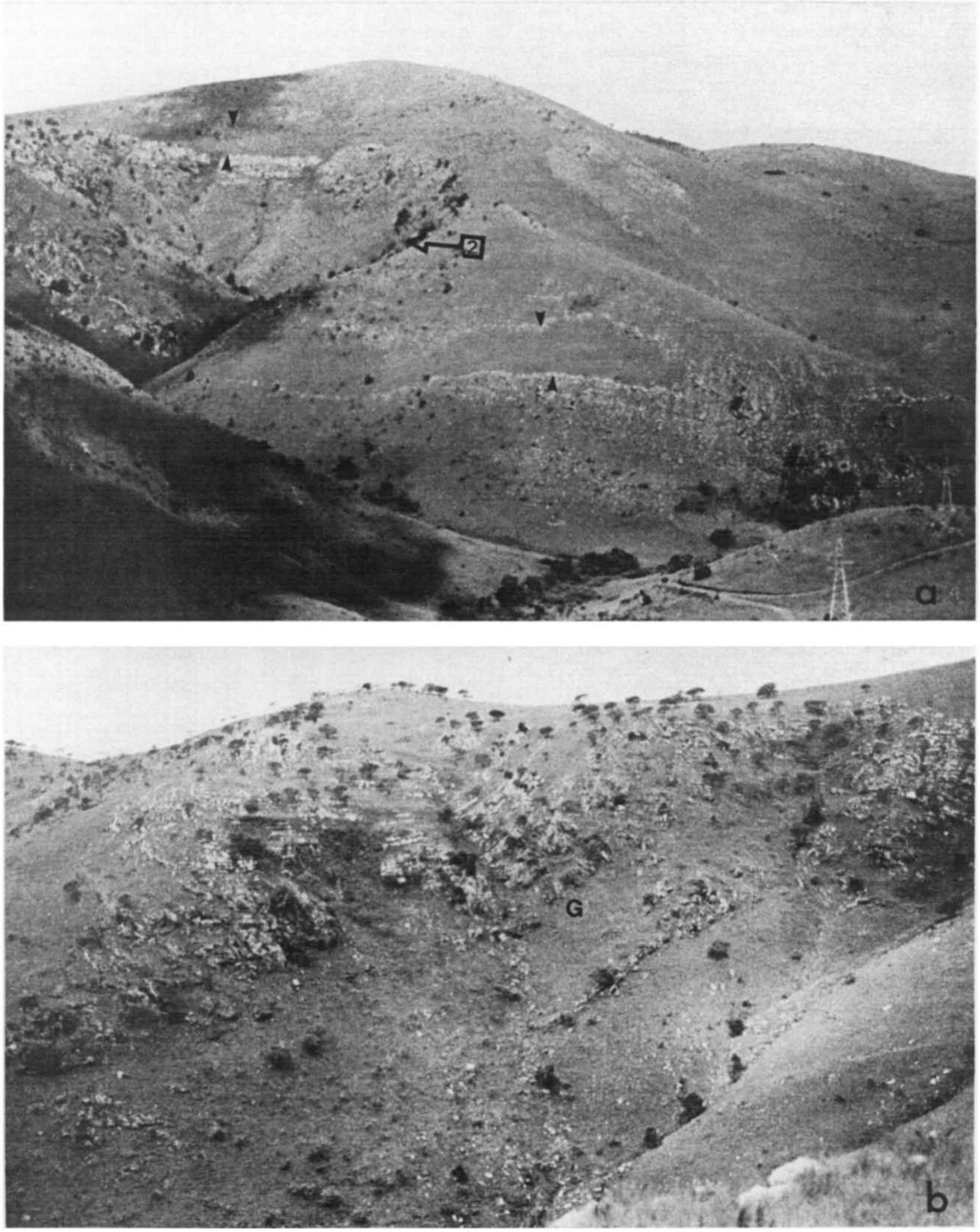


Fig. 3. (a) Recumbent fold at locality B, Figs. 1 and 4. Upper and lower limbs consist of bedded cherts (pale grey) which can be traced for more than 2 km without significant distortion of accretionary spheroids. Carbonates and serpentinites (dark grey) occupy the core. Younging directions as indicated by arrows. Electricity pylons are about 30 m high and spaced at 200 m intervals. Arrow on Fig. 4 shows the position (1) and direction from which this photograph was taken. (b) Open upright folding in the central part of the recumbent fold area, taken from position (2) as indicated on Fig. 3(a) and 4. Abrupt steepening of the fold limb, into a thick fuchsitic shear zone, starts on the extreme left of this photograph.

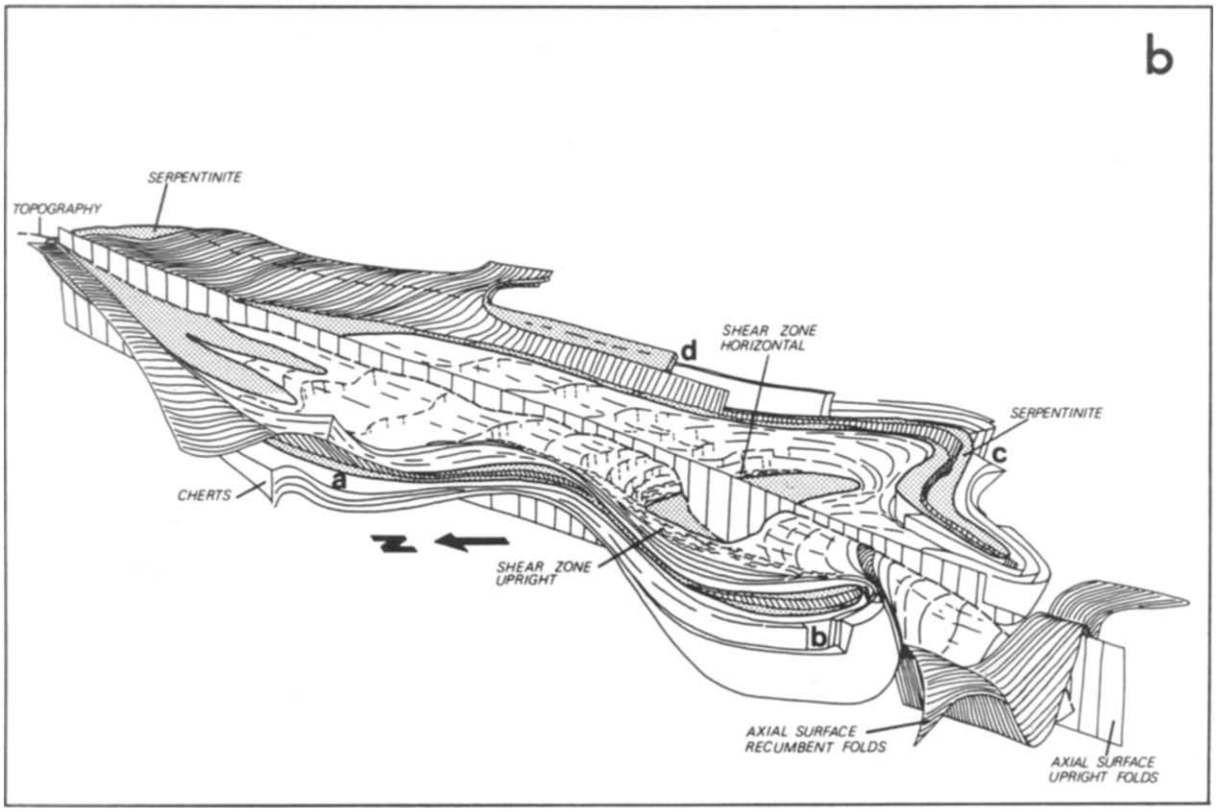


Fig. 5. (a) Recumbent fold area photographed from position (3), Fig. 4. (b) Structural elements from Figs. 4 and 5(a), showing refolding of recumbent fold by later upright folds. Lack of continuous outcrop across and along strike does not allow a completely satisfactory analysis of the geometry of the interference fold pattern. Points a, b, c and d correspond to those shown in (a).

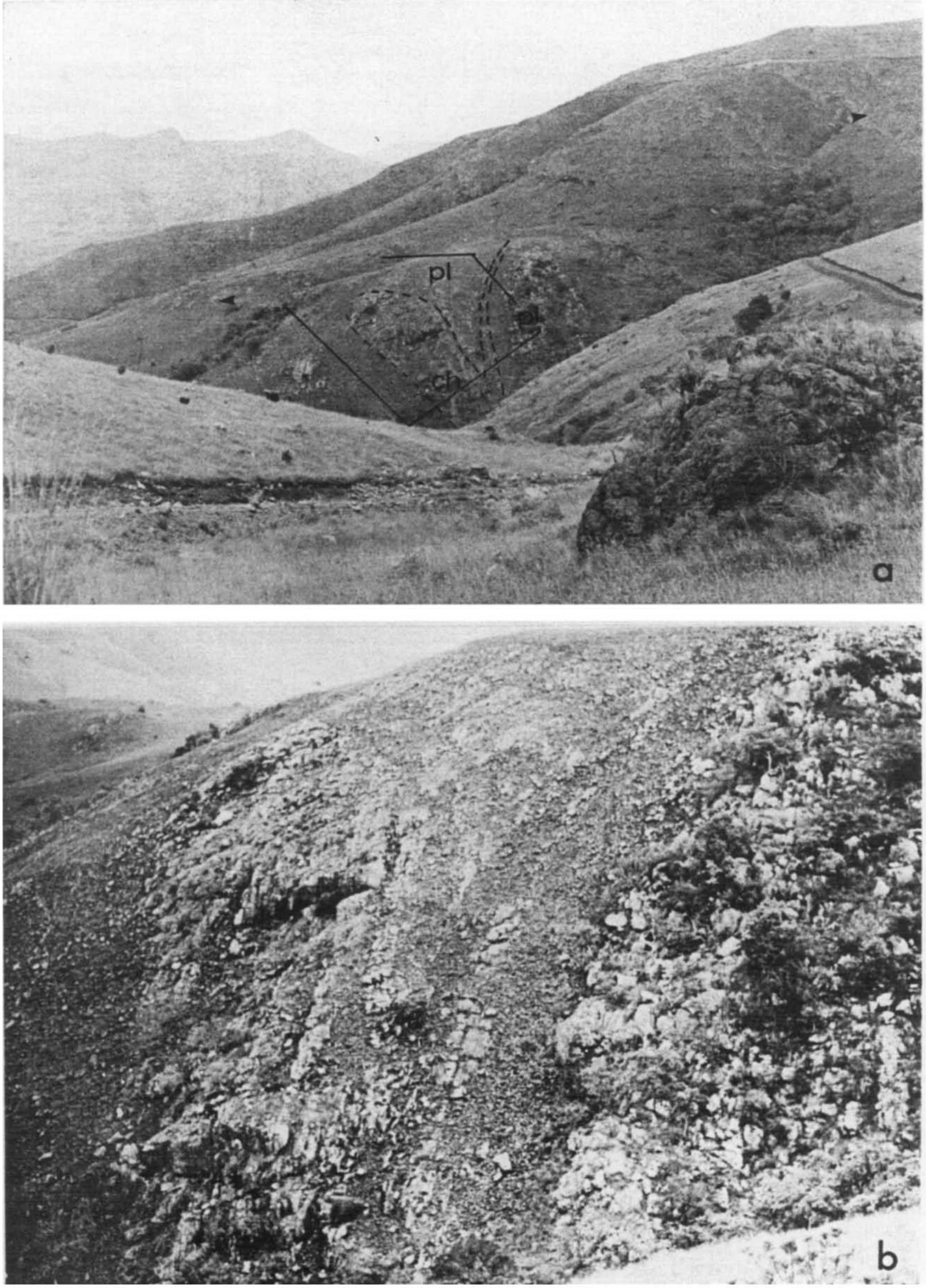


Fig. 6. (a) Upright synformal anticline of locality D, Fig. 1. Younging directions as indicated by arrows; pl, tholeiitic pillow lavas; ch, chert. Photo taken toward the west from locality E, Fig. 1. The central area outlined is shown enlarged in (b).

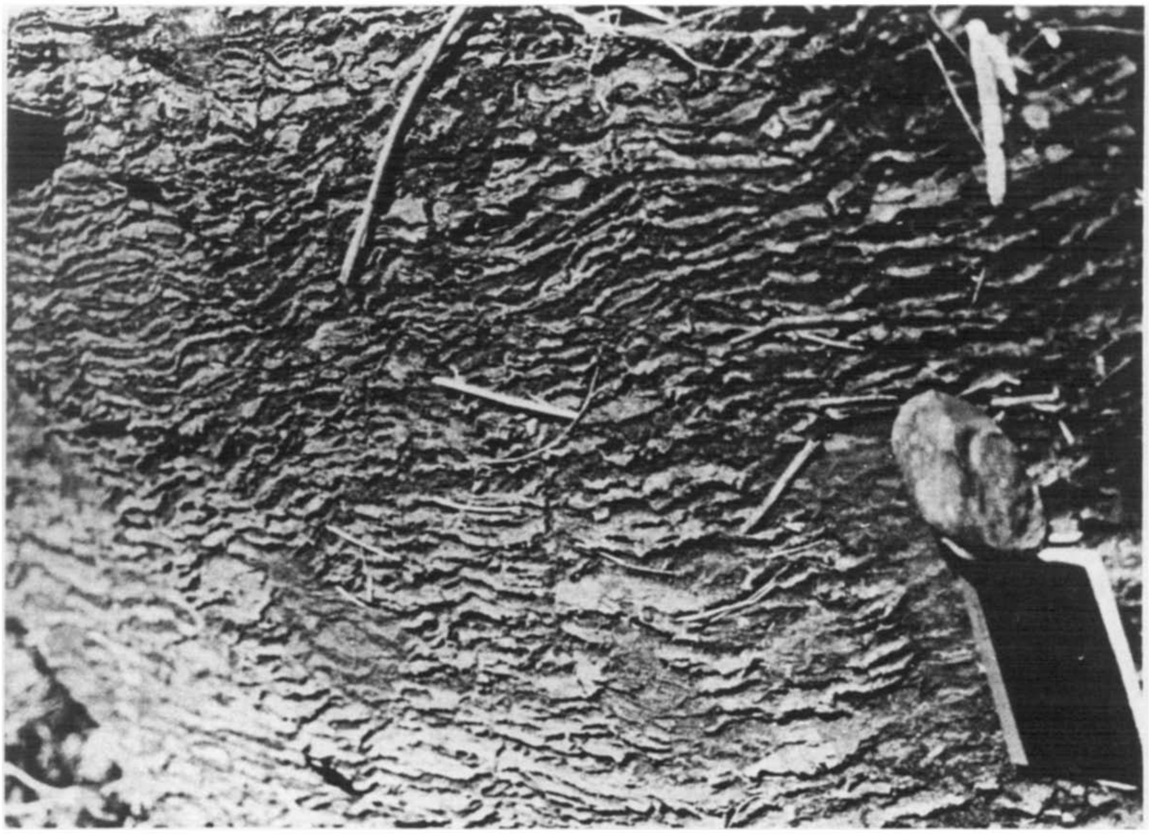


Fig. 7. Flaser-banded quartz–carbonate–fuchsite tectonite (quartz layers, pale grey; carbonate, dark grey) similar to those found in the recumbent fold area between the cherts and serpentinites, (locality G, Figs. 1 and 3b), and which are folded by the late upright folds (Figs. 1 and 4). Note the vestiges of fold closures, indicating that these rocks are tightly folded and flattened.

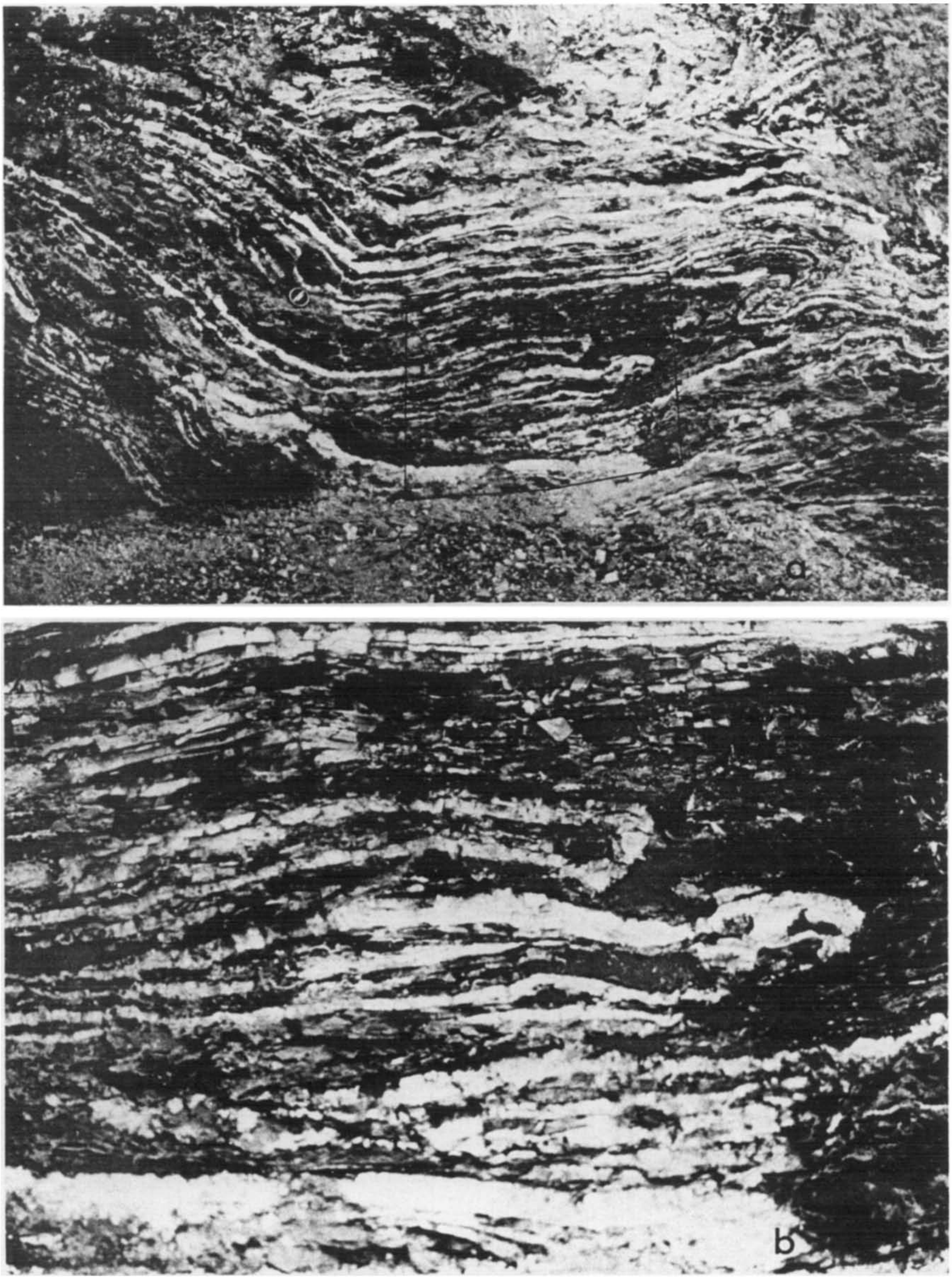


Fig. 8. (a) White chert layer outlining part of early recumbent folds in ferruginous shales (Fig Tree Group) structurally underlying the recumbent fold area. Note the coherence and consistency of thickness of chert layer of the upper recumbent fold over several metres and the shearing discontinuities within lower limb of middle 'nappe'. Fold hinge of lower 'nappe' is missing. Open folds in layers above the recumbent folds are coaxial with the fold nappes. Locality along the Barberton–Havelock road almost 6 km north-east of the recumbent fold area. The central area outlined is shown enlarged in (b).

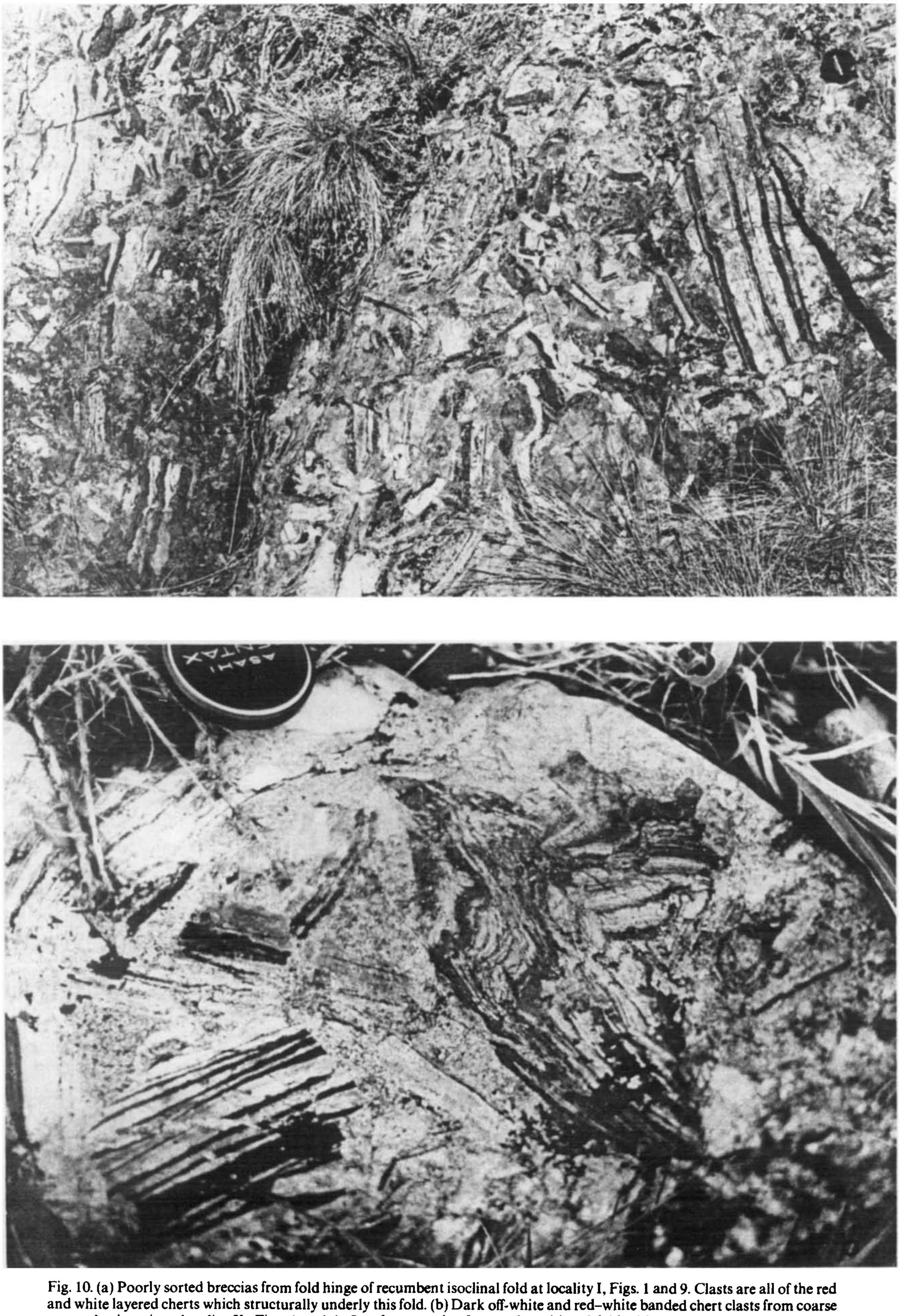


Fig. 10. (a) Poorly sorted breccias from fold hinge of recumbent isoclinal fold at locality I, Figs. 1 and 9. Clasts are all of the red and white layered cherts which structurally underly this fold. (b) Dark off-white and red–white banded chert clasts from coarse clastic unit at locality K, Figs. 1 and 4. One fragment is of a pre-depositional fold with an incipient cleavage.

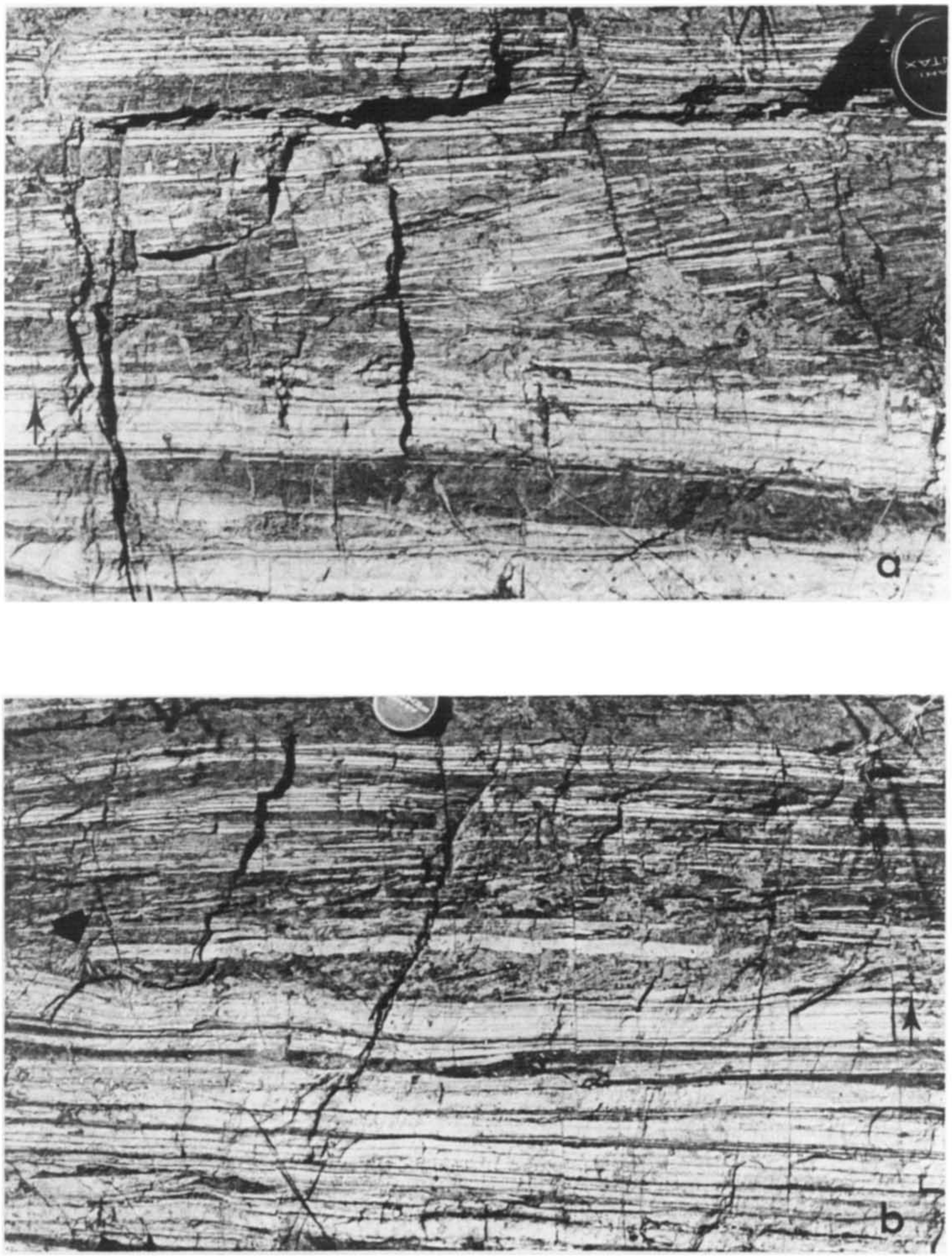


Fig. 11. (a and b) Intraformational breccias, showing fine imbricate nature of thin slabs of chert derived from the breaking up of the thin cherts laminae from chert horizon at locality L, Fig. 1. Top indicated by fine arrows. Note the imbrication which may simulate cross-bedding, and the disruption of such imbrication by the central slab in (b), shown by the stout arrow.

Fig. 12. Abiogenic structures formed during gas evasion in subaerial mud-pools overlying fossil-geothermal systems (locality N, Fig. 1). Such structures, described in detail elsewhere (de Wit *et al.* 1981 in review), occur at numerous localities (Fig. 1).

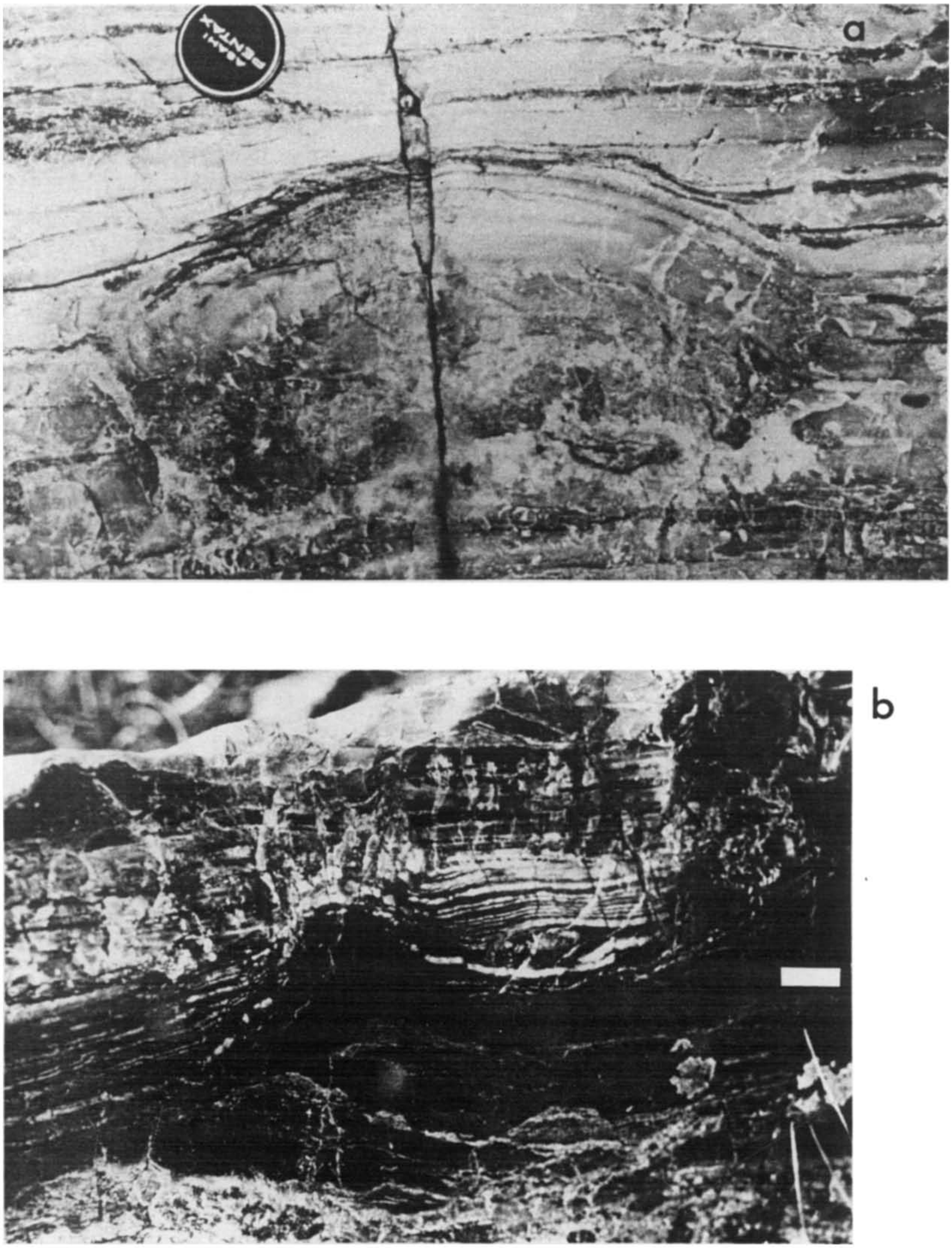


Fig. 13. Diapiric structures common in lithologies which contain fossil mud-pool structures shown in Fig. 12. Note variations between ductile attentuation (necking, boudinage) of layers approaching the diapir in (a), and the more brittle fracturing and displacements (pull-apart grabens) of such layers in (b). Scale bar = 1 cm.

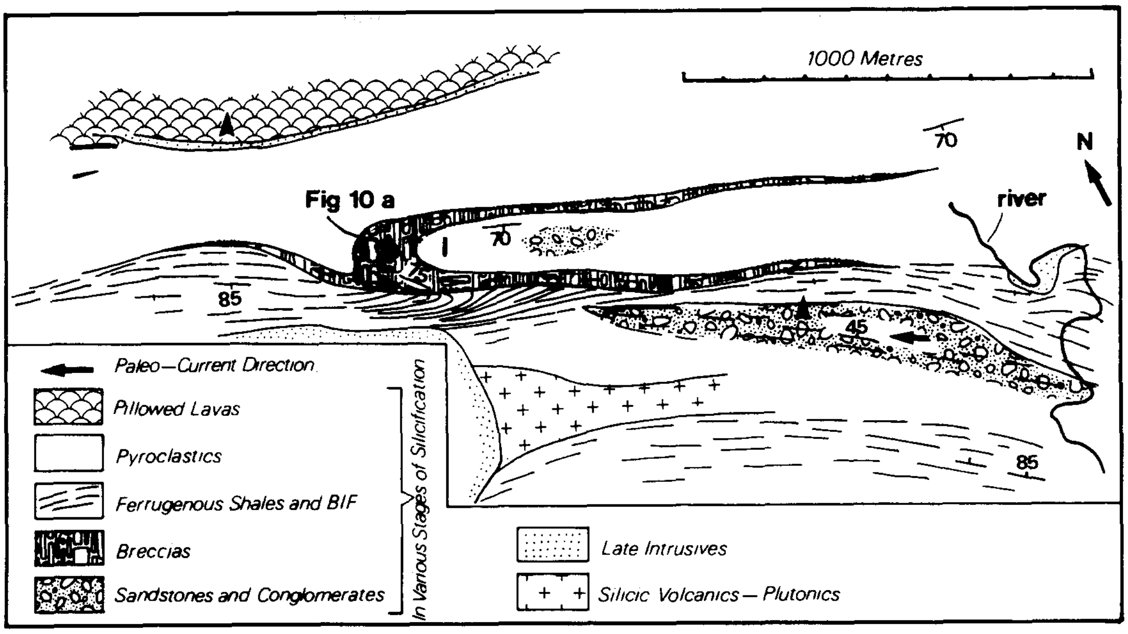


Fig. 9. Geological sketch map of the isoclinal fold around locality I in Fig. 1. There is a gradual eastward transition from a disconformable to a conformable relationship between the breccias which define the fold and the underlying banded cherts, from which the breccia clasts were derived, suggesting that the fold was a "mega slump" structure.

Similar coarse chaotic breccia lenses are common within the shales and cherts underlying the recumbent fold area, (e.g. locality J, Figs. 1 and 4). They range from matrix to clast-supported and may be traced into well-sorted conglomerates. The matrix shows various stages of silicification at different localities (de Wit *et al.* 1981 in review). One lens at locality K contains folded and cleaved banded-chert clasts, indicating a pre- or syn-depositional deformation history (Fig. 10b). These are thus precursory olistostromes, that is materials derived from the front of an advancing allochthonous sheet (Elter & Trevisan 1973).

Within many of the chert units and silicious layers intercalated with the shales, some 5–10 km westward along strike from the recumbent fold area, such slumps and breccias are comprised of finer fragments and are predominantly intraformational. In most places the clasts are slab-like and display imbricate layering (Fig. 11a) which may simulate cross-bedding. These deposits are apparently formed by horizontal translation of rigid chert slabs. Figure 11(b) shows a relatively thick slab apparently 'ploughing' through such a layer of thinner and previously imbricated slabs. The structures suggest continuous deposition of rigid chert slabs at much shallower slope angles than their breccia facies equivalents about 10 km to the east. This is consistent with the increasingly larger scale of slump folding in the east and the concomitant decrease in grading of the accretionary spheroids within the associated cherts.

DISCUSSION

The existence of nappes, olistostromes, regionally inverted stratigraphy and the juxtaposition of sediments from deep- and shallow-water environments, indicates significant early horizontal mass translations including isoclinal over-folding of at least 2 km, within the stratigraphic sections of the upper Onverwacht (Geluk Subgroup) and the Fig Tree Groups. The close association of the nappes with the olistostromes suggests their near contemporaneous emplacements.

Relative timing and directions of mass transportations

Some of the large-scale mass transport evidently took place down an unstable palaeoslope from a subaerial shallow water environment in the south(-east) into a deeper part of a basin with shallower slope to the north (-west) as indicated by various olistostromes and sediment transport directions. It is therefore inferred that gravity tectonics was operative early in the history of the area.

Furthermore, some of the sediments that constitute the nappes and tectonic slices were transported from deep-water into a subaerial environment. By analogy with some modern deposits, the original water depth in the deepest part of the basin could have been as much as 5000 m (Stanistreet *et al.* 1981), but may have been as little as 1500 m, because for the Archaean in general, McKenzie *et al.* (1980) have suggested that the maximum water depth was only about 1500 m.

Two diametrically opposed source areas for the nappes are possible.

(1) The same basin into which the olistostromes were transported, in which case the nappe emplacement was in an opposite direction to that of the sediment transportation and up a palaeoslope. In the map area, this would imply that the nappes moved from deep water in the north(-west) to the shallower environment of deposition in the south(-east) of the area mapped. An analogous Phanerozoic example is clearly documented from the internal belts of the Appalachians where mass movement directions across the palaeo-continental slope changed because of palaeoslope polarity reversal during oro-

genesis (Stevens 1970, Bird & Dewey 1970, Bursnall & de Wit 1975, Dott 1978, Williams 1979). Similar translation sequences have been described for some of the deep-water derived Tethyan allochthons, which contain detritus from the shallow-water platform sediments onto which they were later tectonically emplaced (Smith *et al.* 1975). In this case there should be a discernible age difference between the two mass transportations. This cannot be verified without better chronological control.

(2) A different basin to the south-east of the recumbent fold area. This would infer that both the nappes and the olistostromes were transported in the same direction, analogous to much of the documented Alpine and late-Appalachian tectonism (Aubouin 1965, Trümpy 1973, Clar 1973, Wunderlich 1973, Bird & Dewey 1970, Dott 1978). In this case the olistostromes represent manifestations of a contemporaneous depository ahead of the nappes, and formed either in response to their 'bulldozing' effect and/or during slumping from their unstable frontal slope, (e.g. Bird 1969, Dott 1978).

The presence of deformed clasts (precursory olistostromes) supports this second tectonic model. The occurrence of chromite detritus in some of the shallow-water sediments (Heinrichs & Reimer 1977) also supports an early emergence of ultramafic rocks to the south-east of the recumbent fold area. Such data has been used both in the Alps (Clar 1973) and in the Appalachians (Stevens 1970, Dewey & Bird 1971, Williams 1979) to document timing and transport directions of the tectonic emplacement of ophiolitic nappes. Significantly, it has recently been shown that the Havelock ultramafic complex, about 10 km to the east of the map area (Fig. 1, inset), is allochthonous and possibly of Alpine-type (C. Barton, personal communication ms in prep. 1981).

Both of these above scenarios are incompatible with a stable Atlantic-type continental margin model for the Barberton Greenstone Belt (Eriksson 1980) and support the conclusions of Lowe and Knauth (1977) that the Fig Tree Group represents an orogenic phase of sedimentation in response to major tectonic uplift and exposure of plutonic rocks, and as such is a flysch deposit.

Similarly, the conglomerates (Moodies Group) which stratigraphically overlie the deformed and metamorphosed rocks (Fig. 1, extreme east) and are also derived from the south(-east) (Eriksson 1977), probably represent a molasse-type deposit, scouring large channels into the underlying rocks (Fig. 1). The stratigraphic section as a whole is therefore regressive; that is thick clays loaded by coarse clastic sediments. At locality M, a cross-bedded carbonate unit youngs towards these conglomerates. Thus the intervening carbonate–fuchsite shear zone must represent a significant tectonic break. The carbonate unit is overlain to the east by serpentinites (see also Visser *et al.* 1956) which can be traced along strike into the Rosentuin Complex (Fig. 1, south-east corner), a serpentinized layered ultramafic suite with a strike length of more than 11 km and a thickness of about 150 m, and which has been compared to the Havelock ultramafic complex (Viljoen & Viljoen 1969c, d). The Rosentuin Complex is bound on both sides by carbonate–breccias and fuchsitic cherts, similar to some of the fuchsitic décollement zones in the present study area. The breccias on top of the body contain ultramafic fragments, magnetite and chromite grains in a cherty matrix and younging directions above the serpentinite body are eastward (Viljoen & Viljoen 1969c). Therefore this region could represent an allochthonous serpentinite sheet structurally overlying the conglomerates. If so, the period of early horizontal mass emplacement spanned a significant period of time, as originally suggested by Visser *et al.* (1956, p. 87).

Tectonic emplacement mechanism

Clasts silicified prior to their incorporation into the stratigraphy are abundant (see also Heinrichs & Reimer, 1977) and testify that the silicification process was penecontemporaneous with sedimentation and thus also of a polyphasal and heterogeneous nature throughout much of the time during which the sediments were deposited (de Wit *et al.* 1981 in review).

The fuchsitic shear zones are almost ubiquitously silicified and/or carbonated and contain secondary quartz and carbonate extension veins in various stages of deformation, indicating significant solution–precipitation of these materials and an extended deformational history of these zones. De Wit *et al.* (op cit.) have interpreted these zones to have served as channelways for hydrothermal fluid circulation, in which the flow regimes paralleled the stratigraphy on a kilometre scale. Vertical flow across the stratigraphy through hydrothermal channels was at least an order of magnitude smaller. In the overlying sediments, mud-pool and diapiric structures have been found (Figs. 12 and 13). These have been shown to relate to paleo-hydrothermal hotsprings within a geothermal 'hot spot' where vapour-dominated systems alternated with liquid-dominated systems during successive heat pulses related to igneous intrusions. Thus it seems reasonable to assume that high fluid pressures operated within these zones. De Wit *et al.* (op. cit.) have further shown that associated silica precipitation fronts migrated across bedding planes both up and down the adjacent stratigraphic succession. Coarse clastic beds were particularly prone to silicification, especially close to their interfaces with shales, meta-pillow lavas and serpentinites. Parts of the succession therefore became progressively impermeable, and it is inferred that this would have increasingly retarded dissipation of fluid pressures. Specifically zones trapped between approaching upward and downward migrating silica fronts within the sedimentary pile might have been subjected to progressively higher fluid-pressures. Collectively, such conditions meet all the requirements to induce extension failure parallel to stratigraphic layering by hydraulic fracturing *and* to maintain high fluid pressures to overcome frictional and cohesion forces, which could trigger and maintain momentum for gravitational overthrusting (Hubbert & Rubey 1959; Hsü 1969; Kehle 1970; Roberts 1972; Voight 1974).

Chapman (1974) extended the fluid-pressure hypothesis of overthrust faulting and fold nappe formation to stratigraphically controlled clay diapirism resulting from differential compaction in the upper 2 to 3 km of a young regressive sedimentary sequence. Significantly Chapman (1974) predicted the presence of mud-volcanism at the trailing edge, or 'pull-apart gap', (Hubbert & Rubey 1959, Voight 1974), of thrust sheets driven by gravity sliding.

Chapman's hypothesis is thus compatible with the field observations in this study area, and in addition the latter suggest that the trailing edges of the nappes were areas of relatively high, upwelling, heat flow. Therefore, heat, fluids and gravity were probably the dominating factors driving the horizontal translations, for which 'hovercraft tectonics' (Hughes 1970) is thought to be a vivid descriptive term. A similar interrelationship between fluid transport, deformation, metamorphism and heat flow has been documented for deeper crustal level shear zones (Beach 1976) and used as evidence to indicate large scale overthrusting (Beach & Fyfe 1972).

Scale of tectonic transport

From the deduced sedimentary environments, a vertical differential of between 1000 and 5000 m can be inferred for the relative elevations of the sites of deposition. Assuming a slope of between 2 and 10° to trigger and sustain the gravitational gliding (e.g. Pierce 1973), individual slabs may have travelled between 5 and 150 km. For an angle of 1°, justified by the field observations which appear to simulate the experimental and theoretical boundary conditions of Hubbert & Rubey (1959—"the beer can experiment", p. 161), this range is between 60 and 290 km. The graded units of accretionary spheroids are closely associated with meta-pillow lavas and serpentinites which have been interpreted as Archaean oceanic crust (Anhaeusser 1973, de Wit & Stern 1980). Thus, should the calculated Archaean oceanic depths of about 1500 m prove to be realistic (McKenzie *et al.* 1980), the translations may have been up to about 86 km. Assuming a Newtonian behaviour in the décollement layer, such translations may have occurred within about 1 Ma (Kehle 1970).

Figure 1 shows that most of the chert horizons are discontinuous, but that the longest coherent sections are in excess of 5 km. Thus, apparently individual slabs of at least 25 km^2 were involved during this 'hovercraft tectonics'. Lower temperatures away from the areas of hydrothermal upwelling and in deeper waters of the basin would have enhanced mineral precipitation in the zones of fluid transport. Thus the gliding slabs would have had an inherent natural 'braking system', because such cementation by silica and/or carbonates would have gradually increased the cohesive forces between them.

Stratigraphic and structural implications

The structures described emphasise the need to re-evaluate the stratigraphy within the Barberton Greenstone Belt, especially because it is commonly used as a type example of Archaean greenstone belt stratigraphy for basin modelling and for understanding the Earth's early geotectonic and geodynamic histories. Precisely how these structures should be palinspastically restored is not yet known.

Elsewhere on the Kaapvaal craton (inset Fig. 1) within the probably similar-aged Sutherland and Murchinson schist belts, early recumbent folds have also been documented (Fripp *et al.* 1980). To the north, on the Zimbabwean craton (inset Fig. 1), Stowe (1974) has shown that the Selukwe schist belt, which is about 3.4 Ga old, is downward facing in part. Within the younger (< 3.0 Ga) greenstone belts of the Zimbabwean, Canadian and Australian cratons, large scale stratigraphic inversions and probable nappes have now also been documented (Coward *et al.* 1976, Poulson *et al.* 1980, and Platt 1980, respectively). Thus this type of tangential tectonics may have been widespread and commonplace on the oldest preserved shield areas, although they may not necessarily be the manifestations of the same geotectonic environments (Bridgewater *et al.* 1974).

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