Workshop on

THE TECTONIC EVOLUTION OF GREENSTONE BELTS

HOUSTON, TEXAS
JANUARY 16–18,1986

A Lunar and Planetary Institute Workshop

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HOT SPOT ABUNDANCE, RIDGE SUBDUCTION AND THE EVOLUTION OF GREENSTONE BELTS

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A number of plate tectonic hypotheses have been proposed to explain the origin of Archaean and Phanerozoic greenstone/ophiolite terranes. In these models, ophiolites or greenstone belts represent the remnants of one or more of the following: island arcs (1,2), rifted continental margins (3), oceanic crustal sections (1,4), and hot spot volcanic products (1,3,5). If plate tectonics has been active since the creation of the earth, it is logical to suppose that the same types of tectonic processes which form present day ophiolites also formed Archaean greenstone belts. However, the relative importance of the various tectonic processes may well have been different.

The Archaean earth is postulated to have had greater internal heat production and consequently a younger maximum age of the oceanic lithosphere at subduction (6,7). One of the consequences of a greater proportion of subduction of young oceanic lithosphere in the Archaean is that ridge subduction would have been more common (7). The most common type of ridge subduction in the Archaean would have been that where oceanic lithosphere comprised both the overriding and subducting plate. The only present day example of this type of subduction is the subducting ridge in the Woodlark basin. This ridge crest has several geochemical anomalies: basalts with an island arc signature, and a dacite volcano on the ridge crest (8,9). The island arc component of the basalts has two proposed origins: contamination by an older subducting plate due to polarity reversal of the arc (9) and fluid contamination from the base of the subducting plate (10). Plate reorganization and ridge subduction are both postulated to have been more abundant in the Archaean (7). Regardless of the mechanism by which the arc-like component is generated, Archaean oceanic crust emplaced on land would have been much more likely to have an arc-like composition. Similarly, the dacite volcano observed on the Woodlark basin ridge crest could also have counterparts in Archaean greenstone belts.

Other aspects of the Woodlark basin subduction system may also have relevance for Archaean greenstone belts. The New Georgia island arc, which is being formed by subduction of the oceanic crust of the Woodlark basin (Figure 1), is composed of overlapping volcanoes, located 4-70 km above the Benioff zone (11,12). The New Georgia arc is quite different from a 'typical' Phanerozoic arc, e.g. the Marianas arc (Figure 2). In the Marianas, the volcanoes are spaced 50-100 km apart and sit 125-150 km above the Benioff zone (13,14). The island arc volcanics of the New Georgia arc also have some unusual characteristics. One island is a picritic volcano, thought to be the direct result of the ridge subduction process (8). If a higher percentage of Archaean island arcs were like the New Georgia islands, individual volcanoes would possess overlapping edifices and picritic volcanoes would occasionally occur. The overlapping volcanic edifices would increase the thickness of layer 2 (the pillow basalt layer) and would increase the probability of multiple phases of hydrothermal activity. Consequently, the relative abundance of Archaean ore deposits could be due to the greater incidence of New Georgia-like island arcs.

Another probable consequence of greater internal heat production in the Archaean would have been a greater abundance of hot spot activity. For example, in the Phanerozoic, global ridge volume in the Cretaceous is thought to have increased and to have caused the Cretaceous sea-level high. This increase in the global sea floor creation rate may have coincided with an increase in hot spot activity (15). If increases in hot spot activity do coincide with increases in sea floor creation rate, hot spot activity must have been much more abundant in the Archaean. At present, 10% of all sea floor volcanism is estimated to result from hot spot activity (16). In the Archaean, it is likely that an even greater percentage of sea floor magmatism would have been hot spot generated.

Greater hot spot magmatism in the Archaean would have increased the incidence of bouyant subduction. Bouyant subduction can be a result of subduction of young oceanic crust or of older oceanic crust with a thickened crustal section (7). Much of the oceanic crust which subducts bouyantly has no volcanism or reduced volcanism. This reduction in volcanic activity as a result of bouyant subduction is most common if the overlying plate has a thickened crustal section. Consequently, an increase in hot spot activity in the Archaean could have decreased the percentage of subducting plates causing magmatic activity in the overriding plate, particularly when the overriding plate was relatively cold, thick continental lithosphere.

Areas of hot spot magmatism generally have a thickened pillow basalt section and a greater abundance of highly permeable rocks. These thickened pillowed sections can support more intense hydrothermal activity. Increased hydrothermal alteration at hot spots, particularly ridge-centered hot spots, could also have contributed to the relative abundance of Archaean massive sulfide deposits.

In conclusion, it is probable that many of the differences in preserved Archaean and Phanerozoic greenstone belt/ophiolite terranes can be explained as a result of a difference in the relative importance of different plate tectonic processes. This difference is a direct result of the increased internal heat production of the earth in the Archaean.

REFERENCES: (1) Burke, K., J. F. Dewey, and W. S. F. Kidd (1976), The Early History of the Earth, Windley (ed.), 113-129; (2) Miyashiro, A. (1973), Earth Planet. Sci. Lett., 19, 218-224; (3) Clarke (1970), Contrib. Mineral. Petrol., 25, 203-224; (4) Gass, I. G. (1977), Volcanic Processes in Ore Genesis, Geol. Soc. of London, 72-77; (5) Schulz, K. J. (1977), University of Minnesota, Ph.D. Thesis, 349 pp.; (6) Bickle, M. J. (1978), Earth Planet. Sci. Lett., 40, 301-315; (7) Abbott, D. H., and S. E. Hoffman (1984), Tectonics, 3, 429-448; (8) Johnson et al., AAPG CircumPacific Energy Series, in press, 1985; (9) Perfit et al., AAPG CircumPacific Energy Series, in press, 1985; (10) Abbott, D. H., and M. Fisk, in preparation, 1985; (11) Taylor et al., (1982), EOS, 63, 1120-1121; (12) Cooper, P. A., and B. Taylor (1985), Nature, 314, 428-420; (13) Karig, D. (1971), J. Geophys. Res., 76, 2542-2560; (14) Karig, D. (1971), Init. Rpts. DSDP, 6, Fischer, A. G., Heezen, B. C. et al., 681-689; (15) Ribe, N. M., and A. B. Watts (1982), Geophys. J. R. Astr. Soc., 71, 333-362; (16) Kennett, J. P. (1982), Marine Geology, Prentice-Hall, Inc., 813 pp.

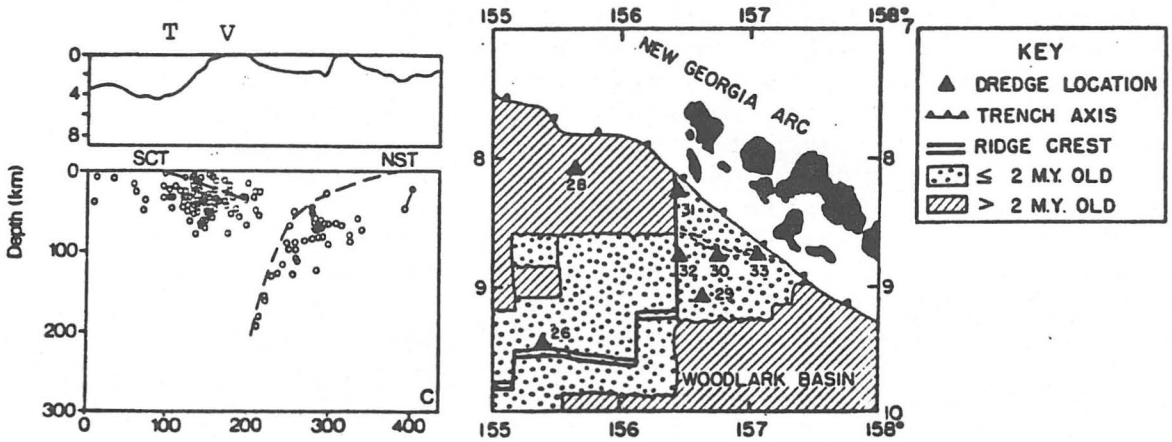

Figure 1. (left) Benioff zone of the New Georgia arc (SCT), after (12); T = Trench, V = Volcanic Line. (right) Volcanoes of the New Georgia arc, after (10).

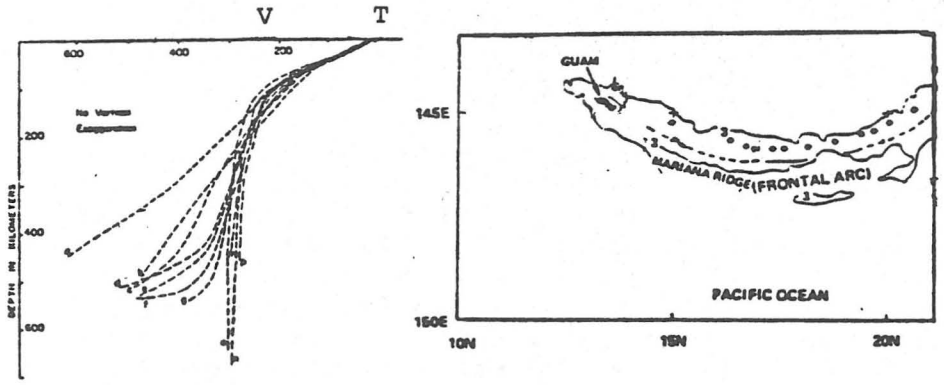

Figure 2. (left) Benioff zone of the Marianas arc, after (13); T and V as in Figure 1. (right) Volcanoes of the Marianas arc are designated by dots, after (14).

LITHOLOGY, AGE AND STRUCTURE OF EARLY PROTEROZOIC GREENSTONE BELTS, WEST AFRICAN SHIELD; Kodjo Attoh, Geology Department, Hope College, Holland, MI 49423

Distribution, Lithologic characteristics and Stratigraphic relations. Distribution of early Proterozoic volcanic rocks in the West African shield is shown in Fig. 1; an approximate boundary between Archean age terrane, to the west, and the Proterozoic terrane to the east, is partly marked by a major fault. Lithologic and chemical data have been compiled for belts (2-9) in the Proterozoic terrane from BRGM reports [1,2] . Available stratigraphic information from geologic maps of these areas indicate that a typical sequence is comprised of predominately mafic lava flows (basalt-andesite) at the base, which are overlain by felsic volcanic rocks including pyroclastic rocks and lavas. This succession, referred to as Lower Birimian, is overlain by Middle and Upper Birimian sedimentary rocks. Lithostratigraphic data from belt (1), located in northeastern Ghana [3], indicate the volcanic succession is 6-8 km thick. The lowest unit in this succession is represented by 2 km of felsic pyroclastic rocks, flows and fine grained sediments. This is followed by 3-4 km of basaltic lava flows which are locally pillowed, the top of the unit is marked by a distinctive manganese formation (MF) consisting of Mn-Fe rich cherts up to 200 m thick. Dacitic lithic tuffs, welded tuffs and andesitic flows up to 2500 m thick overlie the mafic lava flow unit. The youngest volcanic unit consists of mafic tuffs and breccia with a distinctive fragmental texture. Preliminary data indicate that a similar succession occurs in belt (10). The internal plutonic rocks of belt (1) include: (a) hornblende-bearing granodiorite bodies considered to be subvolcanic plutons (σ-plutons); and (b) post-kinematic mica-bearing granitic plutons (pi-plutons). External plutonic rocks include tonalitic and granodioritic rocks which immediately flank the volcanic belt, and paragneisses which occur within the plutonic terrane.

Chemical charcteristics and Ages. Of about 100 chemical analyses reported for belts(2-9) calc-alkaline rocks consitute 55% and tholeiites 45%. Quartz-normative basalt constitutes 99% of the rock type in the tholeiitic suite. In the calc-alkaline suite, 9% of the analyses is basalt, 45% andesite and the rest is dacite and rhyodacite. The ratio of tholeiitic to calc-alkaline rocks based on the stratigraphic thicknesses and chemical analyses of samples from belt (1) is between 57% and 43%. Ultramafic volcanic rocks occur in belt (3), indicated from chemical data from belt (6) and (9) and constitute 1% of all samples analysed. Komatiites have not been reported from the West African Shield, thus the rocks analysed may be classified as high-Mg-basalts. The tholeiitic rocks from belt (1) are enriched in Ti, and depleted in Zr relative to modern ocean floor basalts [4] ,and are depleted in K, Rb, Sr and Ba relative to the calc-alkaline rocks. Within the calc-alkaline suite which include the subvolcanic plutons, the major and trace elements show continuous trends from calc-alkali basalts to rhyolites.

The hornblende-bearing plutons plot in granodiorite and diorite fields of Q-Kf-Pl diagram; whereas the rocks from the pi-plutons have normative and modal mineral compositions of granodiorite, quartz-monzonite and minor quartz synite and monzonite. All the plutonic rocks are strongly HREE depleted [6]. The -plutons (SiO_2=56-66%) show the least depletion with $[La/Yb]_n$ = 13-43. The paragneisses of the external plutonic terrane (SiO_2=70-71%) show the steepest REE pattern with $[La/Yb]_n$ = 33 - 66; while the post-kinematic plutonic rocks (SiO_2=70-75), and La/Yb = 18 - 58, are somewhere in between. Relative to the subvolcanic plutons with (Th=1.9-5.7, and U=0.9 -1.9) the pi-plutons are enriched in Th and U (Th=7.7-10.9 and U=4.5-25ppm). Age of volcanism in the West African Shield is not known; however, K/Ar and Rb/Sr ages have been reported for the rocks which intrude the volcanic rocks and can be used to place minimum age limits. Rb/Sr analyses of mica pi-pluton samples from belts (2-9) yielded the following ages (my): 1870 ± 157 to 2004 ± 42 for whole rock; and 1940± 45 mineral (plagioclase) isochron [5]. K/Ar analyses of amphiboles from belt (1) gave the following ages: (i) an older age of 2223± 283 was obtained from hornblende in the youngest volcanic unit; and (ii) a younger age, 2087 ± 138 was obtained from zoned, titaniferous hornblende in a deformed diorite porphyry intruded into the lowest unit in the volcanic succession. The available data lead to the conclusion that the minimum age for the volcanic activity must be between 2200 and 2100 my. It is significant that Archean ages have not been reported from any of the volcanic belts (1-10).

Structure of an early Proterozoic Volcanic belt in northeastern Ghana. Cleavage in the volcanic belt strikes N40E and dips steeply to the NW and SE. Mesoscopic folds, with locally well developed axial surface cleavage parallel to this foliation, plunge steeply NW and SE. Because the orientation of fold axes and cleavage surfaces do not change with respect to the stratigraphic position, it is concluded that the whole volcanic succession was deformed during a pre-2000 my old orogenic event. Evidence for multiple deformation occurs in the form of NW plunging folds and the folded trace of the axial surface of the major folds. The strong NE-SW orientation of the major structures is such that one has to conclude that the second deformation was not as intense as the first. Foliation in the external plutonic terrane is subparallel to the foliation in adjacent volcanic rocks. Unequivocal evidence for pre-greenstone belt structure was not found in the external plutonic terrane; however, NS structures occur in the paragneisses, which are oblique to NNE-NE structures in the volcanic belt. Gravity anomalies associated with the greenstone belt and the internal plutons have been modelled taking the surrounding plutonic terrane as background. The model predicts that the depth to the bottom of the volcanic succession is 3-4 km. Fig 2 is a structural section of belt(1) based on gravity models especially with regard to allowable geometries of the rock units at depth. The overturned limb of the major anticlinal fold is consistent with available facing indicators.

EARLY PROTEROZOIC GREENSTONE BELTS, WEST AFRICA

Attoh, K.

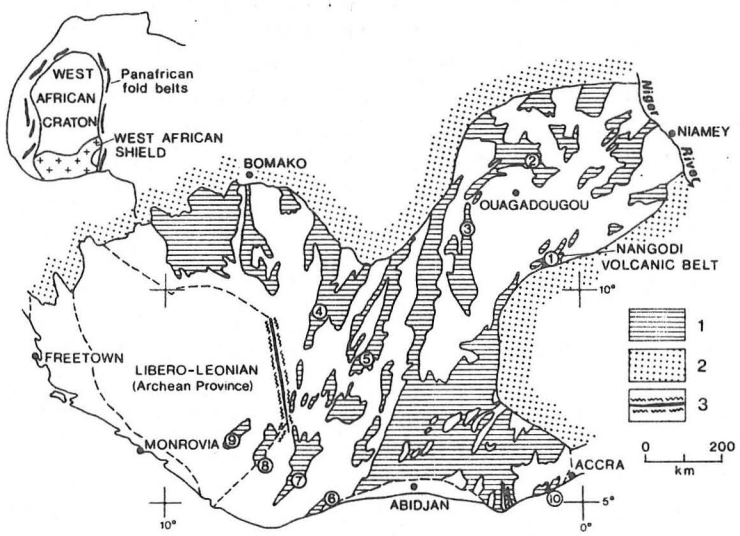

Fig 1. West African Shield showing the distribution of Proterozoic volcanic-sedimentary belts: 1) early Proterozoic volcanic belts, numerical labels referred to in text; 2) late Proterozoic platform sediments; 3) boundary between Archean and Proterozoic shields.

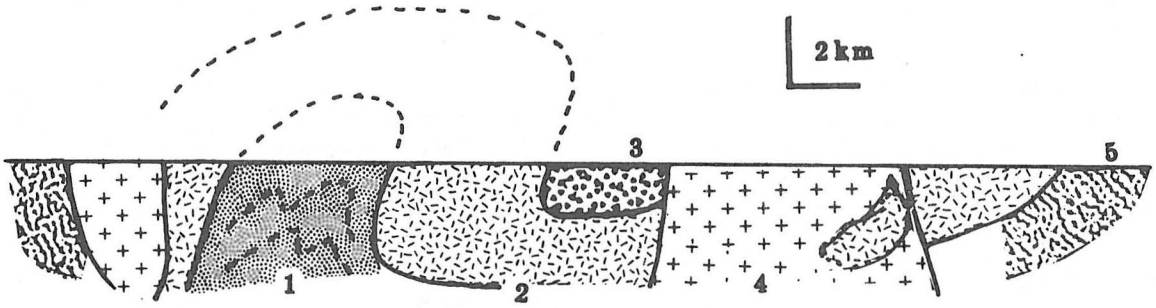

Fig 2. Geologic section across belt (1) in northeastern Ghana: 1) epiclastic sediments and tuffs; 2) mafic lavas (tholeiitic basalts); 3) felsic tuffs and intermediate lavas (calcalkaline); 4) postkinematic granites (pi-granite); 5) granodiorites, tonalites and paragneisses of external plutonic terrane.

References - [1] BRGM: Bureau de Recherches Geologique et Minieres. [2] Alsac, C., (1968, 1969) - BRGM publication G4514; Blanchot, A. et al (1972) Carte Geologique de la partie meridionale de l'Afrigue de l'ouest (BRGM map); Doucelier, J., (1963) BRGM mem 10; Papon, A., (1973) BRGM mem 80. [3] Attoh, K., (1980), Geol. Surv. Can. 80-1C; (1982) Prec. Res. 18, 275-290. [4] Pearce, J. A. and Cann, J. R., (1973) - Earth Plan. Sci. Lett., 19, 290-300. [5] Bessoles, G., (1977) BRGM mem 88. [6] REE analyses done by Schultz, K. J.

CORRELATIONS AND CONTRASTS IN STRUCTURAL HISTORY AND STYLE BETWEEN AN ARCHEAN GREENSTONE BELT AND ADJACENT GNEISS BELT, NE MINNESOTA; R.L. Bauer, Dept. of Geology, University of Missouri, Columbia, MO 65211, P.J. Hudleston, Dept. of Geology and Geophysics, University of Minnesota, Minneapolis, MN 55455, and D.L. Southwick, Minnesota Geological Survey, St. Paul, MN 55114.

Introduction. Low-grade metagraywacke and greenstone of the Vermilion district and amphibolite facies schist and migmatites of the Vermilion Granitic complex (VGC) are separated by a series of east-trending dip-slip and strike-slip faults (1). Structural analysis in the boundary region between these two terranes indicates that they both sustained an early D_1 deformation which lead to recumbent folding. This was followed by a north-south transpression that resulted in the generation of upright F_2 folds and locally well-developed, dextral, D_2 shear zones (2). Despite these correlations, there are distinct differences in structural style and late-stage fold history between the two terranes that we attribute to: 1. differences in the crustal levels of the two terranes during deformation, and 2. effects of late-D_2 plutonism in the VGC.

D_2 deformation produced a series of upright F_2 folds with easterly striking axial planes that are the most prominent fold structures in both terranes. The largest fold of this series is a westerly plunging antiform that straddles the dip-slip fault boundary between the two terranes. Large-scale parasitic folds on this structure are invariably of S symmetry in the southern VGC and occur on the northern limb of the antiform. D_2 dextral shear zones are well represented in the Vermilion district where they are generally parallel to the regional F_2 axial planes. Although distinct ductile shear zones are not observed in the VGC, evidence of a D_2 dextral shear component is locally indicated by asymmetrical pull-aparts and rotated vein segments in the migmatites.

F_1 recumbent folding is inferred from structural facing in the major F_2 antiform that crosses the boundary between the two terranes. Facing is downward on both limbs of the fold which is interpreted to be part of the lower, overturned limb of a large-scale F_1 recumbent fold. A change to upward facing strata further south in the Vermilion district indicates a crossing onto the upper limb of this structure. Finite strain data, determined from clasts in sedimentary/volcaniclastic units in the Vermilion district, can be completely accounted for in terms of the deformation producing the F_2 folds (3). Locally intense F_1 folding in these rocks is therefore attributed to deformation in soft or very poorly lithified sediment. However, biotite schists making up part of the same structure in the VGC display a pronounced S_1 foliation that developed parallel to bedding during the early stages of metamorphism. We have suggested that metamorphic dehydration reactions occurring in the lower strata led to the development of high pore pressures in the upper portion of the sedimentary pile (4 and 5). The combination of high pore pressures and gravitational instability during the F_1

folding resulted in soft-sediment, coherent down-slope movement in the upper strata while the lower strata underwent strain and metamorphic rescrystallization during F_1 folding. Soft-sediment F_1 folding in the Vermilion district could have led to a rather complex distribution of F_1 structures, because the more competent greenstones could not have been soft and therefore may have undergone a much different response to the F_1 folding.

F_3 folding has been observed only in the VGC to the north of the boundary zone with the Vermilion district, near the southwestern contact between the migmatites and the Lac La Croix Granite batholith. Along this margin of the batholith, F_2 folds were reoriented during the emplacement of the pluton and subsequently refolded by F_3 conical folds that formed during the waning stages of the regional north-south transpression that generated the F_2 folds. Such F_3 folds are not observed along the southern margin of the batholith where the F_2 folds are parallel to the batholith boundary and therefore were not reoriented.

In summary, our analysis of the deformation along the boundary between the Vermilion Granitic Complex and the Vermilion district indicates that the two terranes have seen a similar deformation history since the earliest stages of folding in the area. Despite this common history, variations in structural style occur between the two terranes, such as the relative development of D_1 fabrics and D_2 shear zones, and these can be attributed to differences in the crustal levels of the two terranes during the deformation. Similarly, the local development of F_3 folds in the VGC, but not in the Vermilion district, is interpreted to be a result of late-D_2 pluton emplacement which was not significant at the level of exposure of the Vermilion district.

References:

1. Sims, P.K. (1976) Bull. Geol. Soc. Amer., 87, p. 49-62.

2. Hudleston, P.J., Schultz-Ela, D., Bauer, R.L., and Southwick, D.L., (1986) In Workshop on the Tectonic Evolution of Greenstone Belts. Lunar and Planetary Institute, Houston, (this volume).

3. Hudleston, P.J. (1976) Can, J. Earth Sci., 13, p. 587-592.

4. Bauer, R.L. and Hudleston, P.J. (1982) Trans. Am. Geophy. Union, 63, p. 614.

5. Bauer, R.L. (1985) Geology, 13, p. 657-660.

TECTONIC EVOLUTION OF GREENSTONE-GNEISS ASSOCIATION IN DHARWAR CRATON, SOUTH INDIA: PROBLEMS AND PERSPECTIVES FOR FUTURE RESEARCH

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The two fold stratigraphic subdivision of the Archean-Proterozoic greenstone-gneiss association of Dharwar craton into an older "Sargur group" (older than 2.9 Ga.) and a younger "Dharwar Supergroup" (1) serves as an apriori stratigraphic model. The concordannt greenstone (schist)-gneiss (Peninsular gneiss) relationships, ambiguities in stratigraphic correlations of the schist belts assigned to Sargur group and difficulties in deciphering the older gneiss units can be best appreciated if the Sargur group be regarded as a trimodal association of: (i) ultrabasic-mafic metavolcanics (including komatiites), (ii) clastic and nonclastic metasediments and paragneisses and (iii) mainly tonalite/trondhjemite gneisses and migmatites of diverse ages (2) which could be as old as c. 3.4 Ga. or even older. The extensive occurrence of this greenstone-gneiss complex is evident from recent mapping in many areas of central and southern Karnataka State.

The Dharwar Supergroup is deposited unconformably over an ensialic basement comprising the older trimodal association and is further divisible into a lower Bababudan and an upper Chitradurga groups. The volcanic and sedimentary rocks in the Dharwar schist belts display highly variable compositions, lithofacies and stratigraphic thicknesses. The available data is compatible with their deposition in a variably subsiding and progressively evolving basin(s) in an intracratonic or continental margin setting. The Bababudan group is dominated by sediments characteristic of the nearshore intratidal to shallow marine environments and subaerial toshallow marine volcanics (3, 4). The sediment thickness and way-up criteria are suggestive of progressive subsidence of the basin from south to north and concomitant accumulation of sediments derived from both intrabasinal and exterior sources which culminated in the deposition of thick (over 5 kms) sequence of polymict conglomerates and alluvial fan deposits in the rapidly subsiding Kaldurga basin (4). Subsequent sedimentation and volcanism proceeded in essentially deep marine environment as evident from rocks in the interiors of Shimoga and Chitradurga belts. The volcanic character evolved from predominantly tholeiitic (with minor komatiitic occurrences) in the lower units of Bababudan group to calc-alkalic affinities in the upper units of the Dharwar Supergroup. The overall major and trace element compositions of the Dharwar metavolcanics are comparable to Phanerozoic volcanics from continental margin or back-arc settings. While both light REE depleted and enriched types are noted often within the same volcanic formation, an important feature of the metavolcanics is their high Zr/Y character compared to most other Archean volcanic suites in the southern hemisphere suggesting possible trace element heterogeneities in the source regions of Dharwar volcanic rocks (5, 6).

The greenstone and gneiss formations throughout the craton show evidences of two or three phases of deformation with superposed folding resulting often in complex interference patterns. Both pre-Dharwar and Dharwar formations display broadly similar deformation styles and a remarkable parallelism in their tectonic fabrics differing in the intensities of deformation and grade of regional metamorphism (4, 7). The older sequences show superposition of tight upright or overturned isoclinal and/or recumbent folds of the first and second generations (F1 and F2) and a set of open folds (F3) and metamorphosed to amphibolite or granulite facies while the Dharwar rocks are generally in greenschist facies with large scale recumbent and tight isoclinal folds being uncommon (4). The structural history of the craton is complicated by repeated syn or late tectonic diapirism

and intense shearing, strike-oblique slip movements and thrusting particularly along several of the N-S trending regional shear systems (8).

Apart from the general problems concerning the conceptual approaches to early Archean tectonics and crustal evolution, the stages of the tectonic evolution in the Dharwar craton are poorly constrained by lack of information on many crucial aspects of the geology such as; chrono-stratigraphy of schist belts, timing of the major thermal and tectonic events, schist-gneiss relationships and their relative antiquities in the (older) trimodal association, the nature and evolution of the low grade-high grade transitions in the craton. Thus, while the evolution of the pre-Dharwar greenstone-gneiss association is largely enigmatic, the Dharwar Supergroup appears to be a consequence of wide-spread heating of the continental crust around c. 3.0 Ga., tectonic instability resulting in rifting probably along reactivated pre-existing lineaments, formation of broad basin(s), volcanism and sedimentation concomitant with variable rates of subsidence of the basin(s) in response to basement instability and differential upliftment of the surrounding basement highs (horsts?) across the boundary faults (4). The tectonic evolution of the pre-Dharwar crust and the relative importance of the "thick skin" vis-a-vis "thin skin" tectonics (4, 8) to the Archean/Proterozoic history of the Dharwar craton can be assessed only after more detailed structural data on a regional scale become available in conjunction with precise and reliable data on the primary and metamorphic ages of the schists and gneisses in the craton.

REFERENCES

1. Swami Nath J. et al. (1976) Rec. Geol. Surv. India, 107(2), p. 143-175.
2. Bhaskar Rao Y.J. et al (1983) Memoir. Geol. Soc. India, 4, p. 309-328.
3. Bhaskar Rao Y.J. and Naqvi S.M. (1978) in B.F. Windley and S.M. Naqvi (Edt), Archaean Geochemistry, Elsevier.
4. Chadwick B. et al. (1985) Jour. Geol. Soc. India Parts I and II, 26, p. 769-822.
5. Bhaskar Rao Y.J. and Drury S.A. (1982) Jour. Geol. Soc. India, 23, p. 1-12.
6. Drury S.A. (1983) Geochim. Cosmochim. Acta., 47, p. 317-329.
7. Naha K. et al. (1985) Tectonophysics (in press).
8. Drury S.A. et al. (1984) Jour. Geology., 92, p. 3-20.

CAN TRACE ELEMENT DISTRIBUTIONS RECLAIM TECTONOMAGMATIC FACIES OF BASALTS IN GREENSTONE ASSEMBLAGES?

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During the past two decades many words have been written both for and against the hypothesis that the tectonic setting of a suite of igneous rocks is retained by the chemical variability within the suite. For example, Pearce and Cann (1) argued that diagrams can be constructed from modern/recent basalt subcompositions within the system Ti-Zr-Y-Nb-Sr such that tectonomagmatic settings can be reclaimed. If one accepts their general conclusion, it is tempting to inquire as to how far this hypothesis can be extended into other petrological realms. If chemical variations of metabasalts retain information relating to their genesis (tectonic setting), for example, this would be most helpful in reconstructing the history of basalts from greenstone belts.

Pearce and Cann (1) type diagrams are prepared by selecting a training set for which the tectonic settings of all of the analyses are known and obtaining a projection in which overlap of the fields of the known groups is minimized. IF, the training set is representative of a larger target population of interest, the projection may allow assignment of an "unknown" (an analysis not part of the training set) to one of the recognized groups. As the ratios of the variables are retained when percentages are formed, the search for such fields presumes that there are limits on the ratios of the three variables which identify a particular tectonomagmatic setting. The selection of three components and projection onto the plane of the ternary, however, does ignore potentially useful information and one could argue that a dimension-reducing procedure such as principal components analysis might lead to a more satisfactory and potentially useful display form.

However, a successful analysis of date with any multivariate procedure requires more than an understanding of the procedure itself. Additionally, the form of the data should be such that statistical procedures can be rationally interpreted. The subcomposition Ti-Zr-Y-Sr, for example, is part of a set of percentages and therefore subject to all of the concerns previously expressed by Chayes (2), Butler (3) and others concerning difficulties in interpreting both statistical measures of relationship (such as the correlation coefficient) and empathetic analysis of "patterns and trends" expressed in some compositional sub-space.

Simply stated, a set of composition percentages contains a mix of information from at least two sources:

(1) physical/chemical relations among the variables
(2) a change in the structure of the data as a result of a transformation such as percentage formation.

Statistical procedures typically allow one to recognize a behavior pattern that departs from a hypothesized expected behavior. The difficulty in interpreting percentages arises as a result of the mix of information noted above. For example, given

a statistically significant correlation between Zr and Ti, can one automatically assume that contribution from the mechanical process of forming percentages is negligible? Is it possible, in fact, that the mechanical process is the only one operative for a given measured relationship? Can the investigator separate these two effects in a given situation and assess their influence?

Until recently (Aitchison, 4,5) these questions received a great deal of discussion and warning (Chayes, 2 and Butler, 3) but no defined solution. Aitchison (4,5) presents a set of procedures that ultimately are designed to allow an invstigator to make use of information contained within a set of percentages and these procedures are adequately described in the literature. A training set of average Ti-Zr-Y-Sr analyses of 35 modern basalts (including 24 from Pearce and Cann (1) and 11 drawn from the current literature) with known tectonic settings was drawn from the literature Space is insufficient to tabulate these raw data and details will be published elsewhere; copies of the raw data, however, are available from the author).

Aitchison's tests for basis independence (4) and complete subcompositional independence (5) both reject their respective null hypotheses (6). Thus, the investigator is assured that the mechanical contribution is not dominant and that a physical-chemical interpretation is warrented. Each analysis was normalized to its geometric mean and eigenvalues and associated eigenvectors extracted from the variance-covariance matrix of the resulting log-row-centered data using principal components analytical procedures. The first two eigenvalues account for some 92% of the total variance and a plot of the first two principal component scores is given in Figure 1. The boundaries are empirical and constructed so as to isolate the known tectonomagmatic groups. The distribution of scores successfully delineates (1) the Within Plate Basalts, (2) the Ocean Floor Basalts, and (3) the Arc Basalts. The principal component scores are computed as follows:

$$\text{Score 1} = -0.371*Ti-0.067*Zr-0.399*Y+0.836*Sr$$

$$\text{Score 2} = -0.338*Ti-0.560*Zr+0.740*Y+0.158*Sr$$

where the individual variables are expressed in log-row-centered form. In keeping with Pearce and Cann's suggestions (1), Ti is defined as TiO_2 times 100 and Y is defined as 3Y. As one is dealing with a logarithmic function, multiplication by a constant changes the scale of the resulting projection but not the spatial relationships. Ten sets of analyses from the literature were cast into the space defined in Figure 1. In general, the tectonomagmatic settings predicted from Figure 1 are in excellent agreement with interpretations by the respective authors. Of prime concern in this case, however, is the effect of metamorphism on such subcompositions. Many authors (1) have noted that Sr is easily mobilized during low to intermediate grade metamorphism whereas Ti, Zr and Y remain relatively constant. Three of the 35 analyses are plotted in Figure 2 with additions and subtractions of 10% and 30% total Sr. These sets of points define sets of straight lines which are subparallel to the X-axis. Note that the "trend" of these lines is such that it

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may be possible to differentiate between Within Plate and Plate Margin Basalts (Ocean Floor plus Arc) if the above model for Sr mobilization holds.

Perhaps a combination of detailed knowledge of the geology of a particular greenstone assemblage plus judicious use of diagrams analogous to Figure 1 will enable the investigator to see through effects which heretofore may have masked petrogenetically significant information.

References: (1) Pearce, J.A. & Cann, J.R. (1973), Earth planet. Sci. Lett.,19,290. (2) Chayes, F., (1971) Ratio Correlation, Univ. Chicago Press, Chicago, Ill. (3) Butler, J.C. (1979) Amer. Mineral., 64, 1115. (4) Aitchison, J. (1981) Math. Geol., 13, 175. (5) Aitchison, J. (1984) Math. Geol., 16, 617. (6) Woronow, A. & Butler, J.C., (1986) Comp. and Geos., in press.

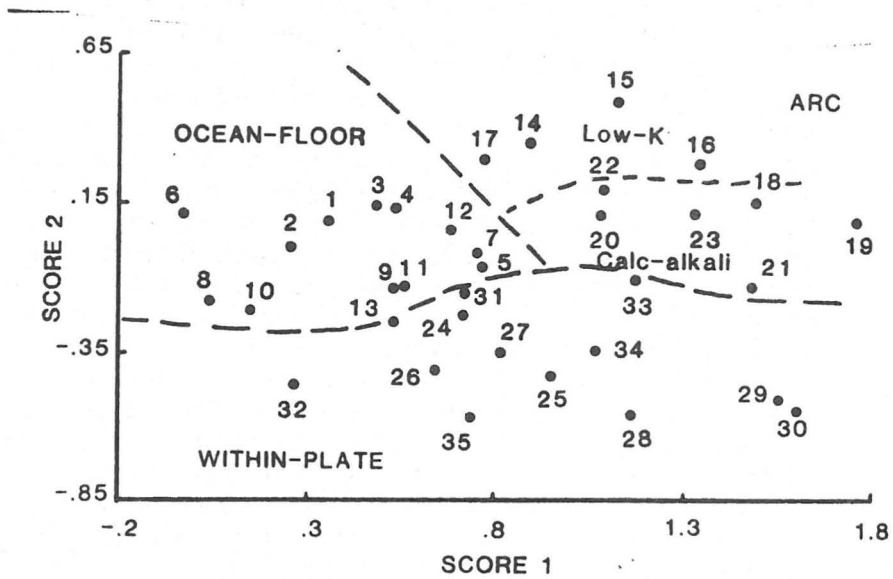


FIGURE 1. A plot of the first two principal components for the training set of 35 basalt analyses. Numbers refer to specific analyses which are available as a separate from the author. Boundary curves are empirical and drawn to isolate the tectono-magmatic facies.

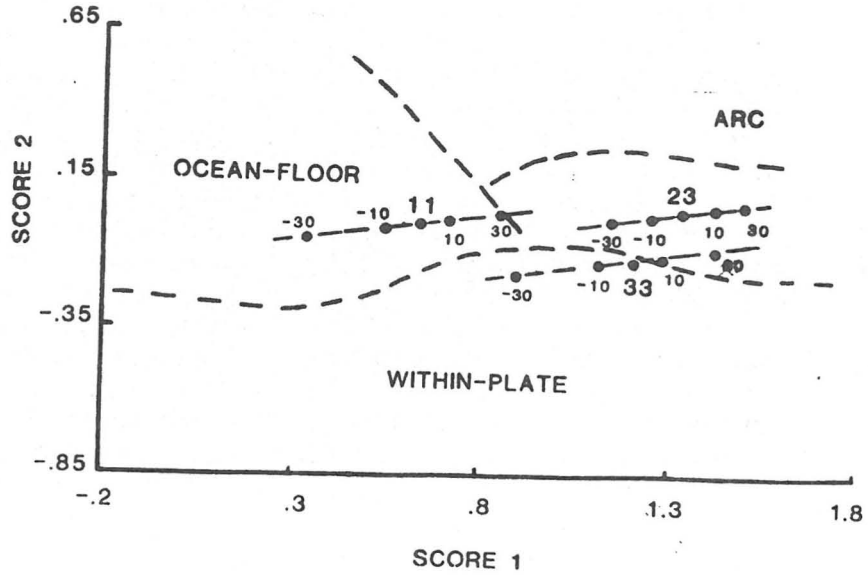


FIGURE 2. Additions and subtractions of 10% and 30% Sr for three of the basalt analyses in the space defined by the first two principal components. Note that the trends are parallel to each other and sub-parallel to the boundary between the Within Plate and Plate Margin facies.

BARBERTON GREENSTONE BELT VOLCANISM: SUCCESSION, STYLE, AND PETROGENESIS; Gary R. Byerly and Donald R. Lowe, Department of Geology, Louisiana State University, Baton Rouge, Louisiana 70803 USA

The Barberton Mountain Land is a small but remarkably well-preserved and accessable early Archean greenstone belt along the eastern margin of the Kaapvaal Craton of southern Africa. Although there is some question about the role of structural repetition of various units, detailed mapping in the southern portion of the belt leads to the conclusion that a substantial thickness is due to original deposition of volcanics and sediments (1). In the area mapped, a minimum thickness of 12km of predominantly mafic and ultramafic volcanics comprise the Komati, Hooggenoeg, and Kromberg Formations of the Onverwacht Group, and at least one km of predominantly pyroclastic and epiclastic sediments derived from dacitic volcanics comprise the Fig Tree Group. Much greater apparent thicknesses of Fig Tree are due to numerous structural repetitions. The essentially non-volcanic Moodies Group lies conformably on top of the Fig Tree. The position or correlation for the Sandspruit and Theespruit Formations relative to the above units is not known. The Barberton greenstone belt formed primarily by ultramafic to mafic volcanism on a shallow marine platform which underwent little or no concurrent extension. Vents for this igneous activity were probably of the non-constructional fissure type. Dacitic volcanism occurred thoughout the sequence in minor amounts. Large, constructional vent complexes were formed, and explosive eruptions widely dispersed pyroclastic debris. Only in the final stages of evolution of the belt did significant thrust-faulting occur, generally after, though perhaps overlapping with, the final stage of dacitic igneous activity.

The volcanic succession in the Barberton greenstone belt is often used as a general model for greenstone belt stratigraphy (2). Previous workers have suggested that volcanism there was cyclic, ultramafic to mafic to felsic, on a scale that ranged from tens of meters to tens of kilometers in stratigraphic thickness, with small cycles superimposed on large cycles. In the grossest sense, the base of the sequence is predominantly komatiitic and the top dacitic, but beyond this the detail of volcanic succession is complex. Thin units of dacitic tuff are recognized within the Komati Formation and komatiitic lavas are interbedded with Fig Tree Group sediments. Simple, small-scale cycles are not present. Sequences previously interpreted as small-scale cycles are now known to represent thick, stratiform alteration zones of mafic and ultramafic lavas to silicic rocks with a remarkably calc-alkaline-like chemistry (3). Systematic increases in Si,K, and Rb accompany decreases in Fe,Mg, and Ni, while Al,Ti, and Zr remain constant from base to top in these cyclic units(4). Throughout these alteration zones the flows typically have mafic volcanic textures and structures, and are usually fine-grained and in places pillowed. Preserved spinels in silicified rocks initially crystallized in mafic or ultramafic lavas. After taking into account the nature of this common style of alteration it appears there are no obvious systematic trends in lava composition in the stratigraphic sequence. Notably, however, the two thick sequences of dacitic volcanics seem to represent prolonged volcanic episodes with no mafic or ultramafic activity.

Komatiitic or basaltic igneous activity seems also to occur with little or no other type of igneous activity in three or four thick sequences.

Styles of igneous activity vary primarily as a function of lava composition. Komatiites throughout the sequence occur as massive flows with typical spinifex textures or as thick flows that often display cumulus-layered bases or as pillowed flows and only rarely as hyaloclastite units. In most sections the flows are quite thin, typically 50cm to 5m, and only rarely up to 50m. They are rarely vesicular, suggesting deep water extrusion, but in several sections interbedded sediments are of shallow-water origin. We have observed no vent complexes for the komatiites, though they are relatively widespread units. The komatiitic unit within the Hooggenoeg can be traced for over 50 km of strike around the Onverwacht anticline. The komatiitic unit beneath the Msauli Chert crops out over a similar distance. Only in the uppermost komatiitic unit is there a local lateral facies; here the lavas interfinger with shallow marine sediments and were of more local extent, though again no vents are recognized. Basaltic igneous activity is characterized by thick to thin lava flows, in places pillowed or massive and only rarely by pillow breccias. Two separate basaltic sequences in the Hooggenoeg crop out for 50 km along the Onverwacht anticline. These lavas are non-scoriaceous, but commonly contain up to 5% vesicles primarily as radial vesicles about the margins of pillows. The lower basalts of the Kromberg occur as a thick sequence of lapilli tuffs, especially thick on the west limb of the Onverwacht anticline (5). These units are locally crosscut by irregular dikes and sills of basalt, and in places contain blocks and bombs of both juvenile and accidental lithologies. They appear to represent near-vent facies and were perhaps similar to modern basaltic cindercone fields. Some lithologies in this unit are moderately scoriaceous. Laterally, these units are represented by interbedded sediments and pillowed to massive lava flows. Dacitic igneous activity is represented on two different scales: by the relatively common tuffaceous units that occur throughout the section, and by very thick sequences of lavas, pyroclastics, and epiclastics at two locations in the sequence (1). Thin, typically a few tens of cm but rarely to a few tens of m, tuffaceous units occur throughout the sequence, and are usually completely altered to a micromosaic of quartz and sericite. Textures are remarkably well preserved, however, and indicate highly pumiceous particles often in the form of accretionary lapilli commonly in graded airfall beds. These units are regionally extensive, greater than 50 km strike length, but associated vent complexes are not found. The two major dacitic units, at the top of the Hooggenoeg and top of the Fig Tree, clearly represent vent complexes. They form complex associations of lava flows or domes, breccias, and tuffs hundreds of m thick. Along strike systematic changes in lithologies can be recognized where sedimentary rocks represent debris being shed off the constructional vent complex. These units do not appear to be laterally interbedded with more mafic lavas.

Petrogenesis of Barberton greenstone belt volcanics is not likely a single, one-stage process. Indeed, the succession of units and common isolation of one compositional group from the others may even

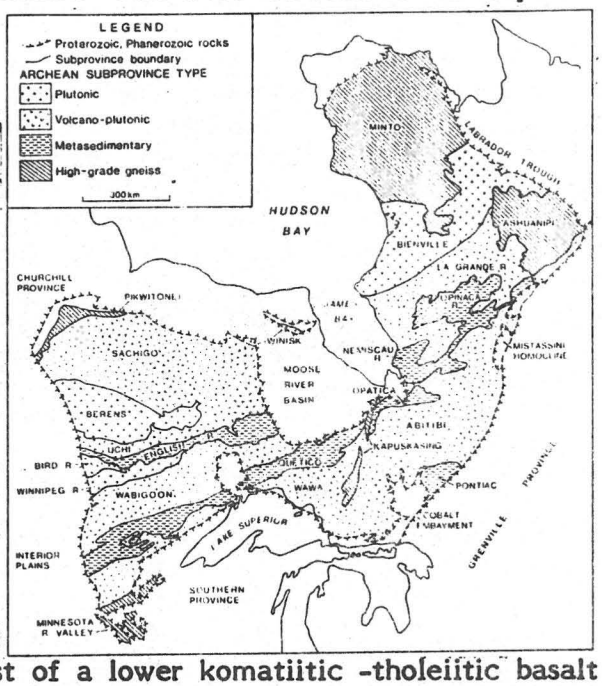
require a separate petrogenesis for komatiites, basalts, and dacites. Komatiites from the top and the base of the sequence are remarkably similar in composition (4,6,7). They are typical of komatiites worldwide except for very low Al/Ti, very high Ti/V, and other ratios that require a very depleted upper mantle source (6). Otherwise, most compositional variation within the komatiitic suite seems consistent with low-pressure fractionation of olivine, later joined by clinopyroxene in komatiitic basalts. Basalts of the Hooggenoeg and Kromberg Formations have typical tholeiitic compositions, including a pronounced iron-enrichment and lack of alumina-enrichment, that can be produced by low-pressure fractionation of plagioclase, clinopyroxene, and olivine. Immobile trace elements and their ratios, such as very low LREE/HREE, also require a depleted upper mantle source. Compositional data are not inconsistent with a single liquid line of descent of komatiites and basalts. While both komatiite and basalt sequences suggest substantial low-pressure fractionation there is not generally an adequate mass of layered intrusives to account for this fractionation in situ. The Barberton sequence contains less than 5% layered intrusives, yet basaltic komatiites and Fe-rich basalts each require 50% or more fractional removal of crystalline phases from their parental melts which must have taken place beneath the present level of the greenstone belt. Dacites are the only intermediate to silicic magmatic rocks found. They range from 60-70% SiO_2, 15-16% Al_2O_3, and have Na_2O/K_2O ratios of about 3 in the freshest samples and are thus trondjhemitic in character. They display extreme fractionation of LREE to Y, and have very high concentrations of highly incompatible elements such as U and Th. Plagioclase and hornblende are the major phenocryst phases in all dacites. Some also contain either quartz or biotite as phenocrysts. Their compositions suggest a source that was mafic in composition and a relatively small degree of partial melting of an assemblage dominated by amphibole. They are not related to associated basalts by any simple, one-stage magmatic process, though could be related to a second stage of igneous activity at the base of a thick, hydrated pile of mafic volcanics.

REFERENCES

(1) Lowe et al,1985,Precamb.Res.,27,165-186.
(2) Anhauesser,1971,G.S.Aust.S.P.#3,57-70.
(3) Byerly et al,1983,LPSC XIV,84-85.
(4) Byerly and Lowe,1985,LPSC,XVI,101-102.
(5) Smith and Erlank,1982,in Komatiites,347-398.
(6) Viljoen et al,1983,in Contr.Geol.Barberton,1-20.
(7) Ransom,1986,M.S. Thesis,LSU.

TECTONIC SETTING AND EVOLUTION OF LATE ARCHEAN GREENSTONE BELTS OF SUPERIOR PROVINCE, CANADA: K.D. Card, Geological Survey of Canada

Late Archean (3.0-2.5 Ga) greenstone belts are a major component of the Superior Province of the Canadian Shield where alternating, metavolcanic - rich and metasedimentary - rich subprovinces form a prominent central striped region bordered in part by high-grade gneiss subprovinces, the Pikiwitonei and Minto in the north, and the Minnesota River Valley in the south. The high-grade gneiss subprovinces are characterized by granulite facies gneiss of plutonic and supracrustal origin, and by abundant plutonic rocks. Minnesota River Valley has rocks older than 3.5 Ga; absolute ages of Pikiwitonei and Minto rocks re unknown but Minto does have north-south structural trends distinctive from the dominant east-west structures of Superior Province.

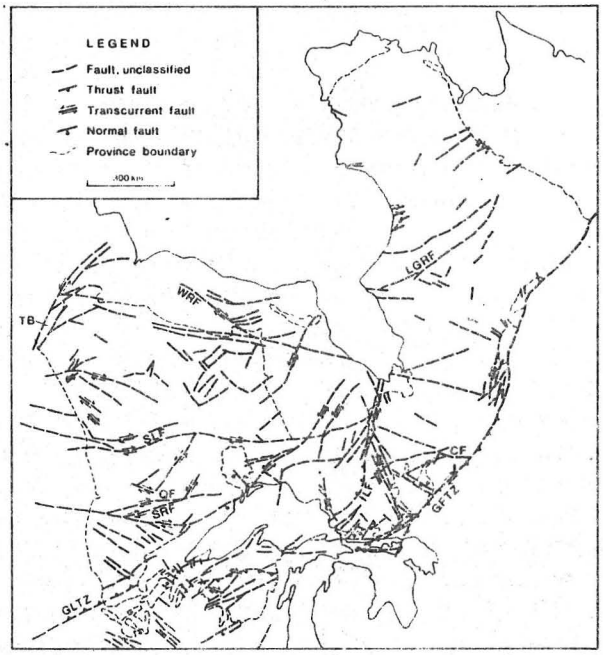


Volcano-plutonic subprovinces of Superior Province consist of generally narrow, sinous greenstone belts bordered and intruded by voluminous plutonic rocks, including tonalitic gneiss, synvolcanic plutons, and younger foliated to massive, generally composite plutons, ranging from quartz dorite to granite and syenite. Supracrustal rocks of the greenstone belts include komatiitic, tholeiitic, calc-alkalic, and rare alkalic volcanics with volcanogenic clastic (wacke, conglomerate) and chemical (iron formation, chert) sediments. Most greenstone belts consist of several lensoid, overlapping piles each on the order of 100 km in maximum dimension and approximately 5 to 10 km thick and commonly comprising several volcanic cycles. Some cycles consist of a lower komatiitic -tholeiitic basalt sequence, a middle tholeiitic basalt -andesite sequence, and an upper calc-alkalic dacite-rhyolite-andesite sequence. Other cyles are bimodal tholeiitic basalt-dacite (rhyolite) sequences. Minor alkalic and shoshonitic volcanics and associated alluvial/fluvial sediments are present in some belts where they unconformably overlie older volcanics and synvolcanic plutons. In term of rock types, sequences, and overall configuration, many Superior Province greenstone belts are closely comparable to modern island arcs.

Superior Province greenstone belts typically have upright folds with curved, bifurcating axial surfaces, steep foliations and lineations, and major domal culminations and depressions, the products of polyphase deformation. Some belts, however, display low angle foliations and faults, overturned sequences, and recumbent and downward - facing structures suggestive of thrust-nappe style tectonics (1,9,10).

The enclosing gneissic and plutonic rocks display domal structural patterns, again the product of polyphase deformation involving recumbent folding and diapirism. Metamorphic grade in the greenstone belts is generally subgreenschist to greenschist in the central parts grading outward to low pressure amphibolite facies in belt margins and surrounding plutonic gneisses.

The contacts between greenstone belts and enclosing plutonic rocks, and between the greenstone-rich subprovinces and adjacent plutonic subprovinces are generally either intrusive or tectonic. An unconformity between greenstones and older granitoid rocks has been demonstrated only at Steeprock, Ontario(5) and although younger volcanics and older plutonic rocks are juxtaposed in a number of places, faults, mylonites, or shear zones invariably intervene. Dextral transcurrent faults trending EW and NW and sinistral faults trending NE form subprovince boundaries in part, as do NE and EW trending thrusts. One notable product of this faulting, the Kapuskasing Structural Zone, exposes granulites considered to represent upthrusted lower crust (7,8). Late alkalic volcanic - fluviatile sediment sequences are spatially related to major transcurrent faults and may represent deposition in pull-apart basins formed by alternating periods of transtension and transpression in strike-slip zones.



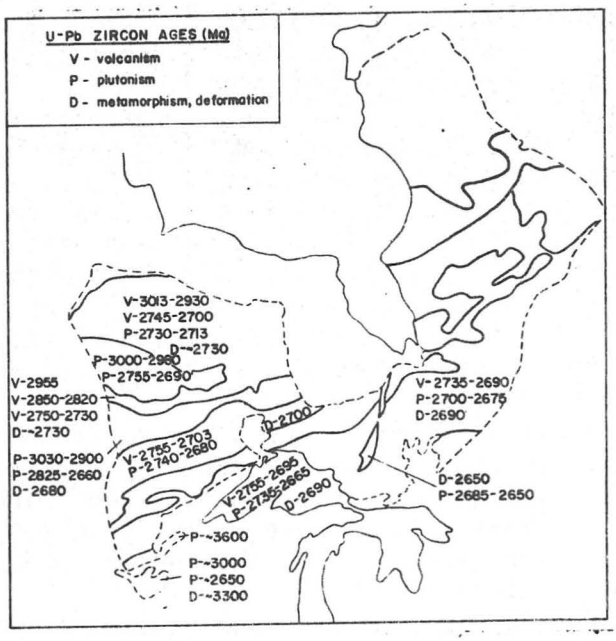
Interpretation of geophysical data shows changes in depth to the Conrad Discontinuity and to the Moho from one subprovince to another, indicating significant structural relief across their faulted boundaries (4). Greenstone belts of Abitibi and Wabigoon subprovinces generally extend to depths of only 5 to 10 km(3) whereas metasedimentary gneisses of English River Subprovince and plutonic rocks of Winnipeg River Subprovince may extend to depths of 10 to 20 km(4). Juxtaposition of high-pressure granulites of the Kapuskasing zone with low-pressure greenschist-amphibolite facies rocks of Abitibi Subprovince implies structural relief of 15 to 20 km across the boundary thrust (7,8).

Metasedimentary subprovinces (English River, Quetico, Pontiac etc.) consist mainly of turbidite wacke and pelite metamorphosed at grades ranging from low greenschist at belt margins to upper amphibolite and locally, low-pressure granulite in belt interiors. Anatectic, s-type granitic rocks are prevalent in the migmatitic, high-grade interiors of the metasedimentary belts.

Most metasedimentary subprovinces have a linear aspect attributable to transcurrent boundary faults and isoclinal folds with subhorizontal to subvertical axes, late structures superimposed on earlier complex, recumbent folds and dome-basin structures. In areas where contacts between metasedimentary and volcano-plutonic subprovinces are unfaulted, there appear to be rapid facies transitions from sedimentary to dominantly volcanic sequences. Preliminary isotopic age data also indicate that the sedimentary and volcanic sequences of some adjacent subprovinces are broadly coeval.

U-Pb zircon dates demonstrate that volcanic, plutonic, deformational, and metamorphic events of relatively brief duration affected large parts of Superior Province and that there are detectable differences in ages of these events from one area to another(6). In the northwest (Sachigo, Berens, Uchi subprovinces) major volcanism and

accompanying plutonism occurred at about 3.0 to 2.9 Ga, 2.85 to 2.80 Ga, and 2.75 to 2.7 Ga. These voclanic episodes were followed by major deformation, metamorphism, and plutonism about 2.73 to 2.7 Ga. In the south (Wabigoon, Wawa, Abitibi subprovinces) volcanism and plutonism occurred mainly between 2.75 and 2.69 Ga, followed by major deformation, metamorphism, and plutonism at about 2.70 to 2.66 Ga. There is evidence for somewhat younger (2.65 to 2.63 Ga) metamorphic-plutonic events, or of later closure of isotopic systems, in the high-grade rocks of the metasedimentary belts and of the Kapuskasing zone.



In summary, Superior Province consists mainly of Late Archean supracrustal and plutonic rocks with Middle Archean gneisses in the south and possibly in the north. The Late Archean supracrustal sequences are possibly mainly of island-arc and inter-arc affinity, although continental rift zone settings have also been postulated(2). Abundant plutonic rocks include early synvolcanic intrusions and later synorogenic and post-orogenic intrusions derived in part from the mantle and in part from crustal melting caused by thermal blanketing of newly-thickened continental crust combined with high mantle heat flux.

The contemporaneity of magmatic and deformational events along the lengths of the belts, coupled with the structural evidence of major compression and transcurrent faulting, is consistent with a subduction-dominated tectonic regime for assembly of the Superior Province orogen. Successive lateral and vertical accretion of volcanic arcs and related sedimentary accumulations, accompanied and followed by voluminous plutonism, resulted in multi-stage crustal thickening and stabilization of the Superior craton prior to emplacement of mafic dyke swarms and Early Proterozoic marginal rifting.

REFERENCES

1. Attoh, K. (1981); Geological Survey of Canada, Paper 81-1B, p. 49-54.
2. Ayres, L.D. and Thurston, P.C. (1985): Geological Association of Canada, Special Paper 28, p. 344-380.
3. Gupta, J.K., Thurston, P.C. and Dusanowskyj, T.H. (1982): Precambrian Research, v. 16, p. 233-255.
4. Hall, D.H. and Brisbin, W.C. (1982): Candian Journal of Earth Sciences, v. 19, p.2049-2059.
5. Joliffe, A.W. (1955): Economic Geology, v. 50, p. 373-398.
6. Krogh, T., Corfu, F. and Davis, D.W. (1985): G.A.C. Spec. Paper 28, p. 358.
7. Percival, J.A. and Card, K.D. (1983): Geology, v. II, p. 323-326.
8. Percival, J.A. and Card, K.D. (1985): G.A.C. Spec. Paper 28, p. 179-192.
9. Poulsen, K.H., Borradaile, G.J., and Kehlenbeck, M.M. (1980): Canadian Journal of Earth Sciences, v. 17, p. 1358-1369.
10. Thurston, P.C. and Breaks, F.W. (1978): Geological Survey of Canada, Paper 78-10, p. 49-62.

HIGH PRECISION U-PB GEOCHRONOLOGY AND IMPLICATIONS FOR THE TECTONIC EVOLUTION OF THE SUPERIOR PROVINCE; D.W. Davis, F. Corfu, and T.E. Krogh, Jack Satterly Geochronology Laboratory, Royal Ontario Museum, 100 Queen's Park, Toronto, Ontario. M5S 2C6

The underlying mechanisms of Archean tectonics and the degree to which modern plate tectonic models are applicable early in earth's history continue to be a subject of considerable debate. A precise knowledge of the timing of geological events is of the utmost importance in studying this problem. The high precision U-Pb method has been applied in recent years to rock units in many areas of the Superior Province. Most of these data have precisions of about ± 2-3 Ma. The resulting detailed chronologies of local igneous development and the regional age relationships furnish tight constraints on any Archean tectonic model.

Superior province terrains can be classified into 3 types:

1) low grade areas dominated by meta-volcanic rocks (greenstone belts).
2) high grade, largely metaplutonic areas with abundant orthogneiss and foliated to massive I-type granitoid bodies.
3) high grade areas with abundant metasediments, paragneiss and S-type plutons.

Most of the U-Pb age determinations have been done on type 1 terrains with very few having been done in type 3 terrains.

A compilation of over 120 ages indicates that the major part of igneous activity took place in the period 2760-2670 Ma, known as the Kenoran event. This event was ubiquitous throughout the Superior Province.

There is, however, abundant evidence for the widespread occurrence of pre-Kenoran volcanoes and sialic crust, especially north of the Wabigoon-English River subprovince boundary. In the Uchi and Sachigo subprovinces there are volcanic periods about 3000-2900 Ma (2,3) and 2850-2800 Ma (2,3) in age which underlie the Kenoran sequence. The Kenoran rocks are in part disconformable on the older sequences. The general absence of angular unconformities along with other evidence such as the presence of mature sandstones in the Sachigo subprovince (4), implies an extended period of crustal stability preceding the Kenoran event. Tonalites 2950-3200 Ma in age are found in the Favourable Lake (5), North Spirit Lake (3) and Winnipeg River Belts (6,7,8), suggesting a pre-Kenoran crust-forming event at about 3000 Ma. Evidence for the existence of an extensive pre-Kenoran continent is especially strong in the Winnipeg River Belt, a type 2 terrain. Recent data obtained from a type 2 terrain in the Wabigoon subprovince also indicates 3000 Ma volcanic and plutonic sequences (9). This indicates that type 2 terrains in many cases include pre-Kenoran crust, and that pre-Kenoran crustal material may be locally present in type 2 areas throughout the Superior Province.

The earliest Kenoran magmatism consisted of eruption of tholeiitic basalt platforms. These are difficult to date but in some areas pre-date 2750 Ma (10). Intermediate-felsic calc-alkaline volcanism occupied a time span from about 2750-2700 Ma and led to the construction of large composite volcanoes. The transition to calc-alkaline volcanism was associated with the emplacement of layered basic intrusions and contemporaneous tonalite-granodiorite plutons, without major deformation. This was followed by the intrusion of high alumina trondhjemite-granodiorite plutons, in some cases accompanied by later calc-alkaline volcanism. This resulted in the development of large intravolcanic batholiths (11). Significant regional

deformation began relatively late, at about 2700 Ma over much of the Superior Province. In some areas, such as the Wabigoon greenstone belt, it significantly post-dated the bulk of calc-alkaline intrusive and volcanic activity (12).

Some greenstone belts underwent at least two periods of deformation. A D1 event affected the calc-alkaline sequences and pre-dated sedimentation and eruption of alkaline volcanic rocks (e.g. Tamiskaming sequence). The Tamiskaming-type sequences were then affected by a later deformational (D2) event. The ages of the late sequences and the D2 event are bracketed between 2695 and 2685 Ma in the southern Wabigoon and Shebandowan subprovinces (13), but may have been 15-20 Ma earlier in the Oxford Lake belt in the northern Superior Province.

The causes of regional deformation are unclear. It may have been partly the result of diapiric remobilization of the intravolcanic batholiths (e.g. Wabigoon greenstone belt), accompanied by regional compression, perhaps due to the intrusion of marginal late granitoid plutons (e.g. Batchawana belt) (14). The presence of nappe structures in some areas such as the Winnipeg River belt (15) and the southern Wabigoon subprovince (16) further complicates the tectonic picture. The final expression of strain was the establishment of large strike-slip faults, which often separate type 1 from type 2 and type 3 terrains.

The deformational event was accompanied by intrusion of late tectonic plutons, most of which have ages in the range 2700-2670 Ma (17). This resulted in cratonization and brought the Kenoran event to an end. Locally, single volcanic centers passed through this cycle from initial volcanism to terminal deformation in time spans as short as 30 Ma (18).

Although there is some indication of a secular younging in the peak of Kenoran igneous activity in a N-S direction, the broad simultaneity and short time spans of crustal events argue against any simple model for growth of the Superior Province by accretion of island arcs (19). Furthermore, there is a strong vertical control on magmatic and metamorphic ages. The oldest Kenoran plutons occur high in the crust while the youngest plutons and metamorphic ages are found at deeper crustal levels in more uplifted and eroded terrains such as the Berens River subprovince, parts of the Winnipeg River belt and the Kapuskasing structural zone (20).

Despite considerable work on the felsic units in type 1 greenstone terrains, there is almost no evidence of inherited zircon components derived from significantly older sialic material. The intravolcanic granitoid rocks and thick felsic volcanic sequences were largely derived by differentiation processes from mafic precursors within the period of Kenoran activity (11). However, greenstone belts evidently did develop adjacent to older sialic blocks. Evidence for this, found in the Wabigoon greenstone belt, includes pre-Kenoran granitoid clasts in a conglomerate marginal to the belt and the existence of marginal unconformities between 3000 Ma tonalite in the Winnipeg River belt and Kenoran volcanic and plutonic sequences (7). Abundant mafic dykes intrude the older units below these unconformities and indicate a tensional stress regime.

The bulk of the evidence presently available argues for a model in which greenstone belts were initiated by rifting of older sialic crust and the formation of narrow ocean basins. The fault controlled nature of many subprovince boundaries as well as the fact that volcanism was at times nearly coeval throughout the Superior Province suggests that rifting may have been concentrated along major early lithospheric breaks.

Evidence for subduction in late Archean tectonic processes is missing. The absence of an effective subduction mechanism would have inhibited ocean spreading. If the intracratonic rifts were not able to open into wide ocean basins they would have been reworked in place, undergoing dominantly vertical tectonic processes. Continued mantle-derived mafic magmatism may have led to thickening and differentiation of the crust to produce the large amounts of calc-alkaline material now present in type 1 terrains.

Any model for tectonic development can only be tentative and subject to the constraints of a constantly expanding data set. Some of the major questions remaining for geochronology are the extent in time and space of pre-Kenoran material and its deformational history and the origin and basement of the metasedimentary belts. These questions can only be resolved by much more extensive work in type 2 and type 3 terrains.

References

1 Kroner A. (ed.) (1981) Precambrian Plate Tectonics, Elsevier, Amsterdam.
2 Nunes P.D. and Thurston P.C. (1980) Can. J. Earth Sci. 17, p. 710-721.
3 Corfu F., Wallace H., Wood J. and Ayres L.D. (1984) (abstract) Geol. Assoc. Canada annual meeting, London, Ont., p. 54.
4 Donaldson J.A. and Ojakangas R.W. (1977) Can. J. Earth Sci. 14, p. 1980-1990.
5 Corfu F. and Ayres L.D. (1984) Contrib. Mineral. Petrol. 88, p. 86-101.
6 Krogh T.E., Harris N.B. and Davis G.L. (1976) Can. J. Earth Sci. 9, p. 1212-1215.
7 Clark G.S., Bald R. and Ayres L.D. (1981) Can. J. Earth Sci. 18, p. 94-102.
8 Corfu F, Beakhouse G.P., Stott G.M. and Sutcliffe R.H. (1985) (abstract) Institute on Lake Superior Geology, 31st annual meeting, Kenora, Ont., p. 17.
9 Davis D.W. and Jackson M.C. (1985) Ontario Geological Survey Summary of Field Work.
10 Davis D.W., Blackburn C.E. and Krogh T.E. (1982) Can. J. Earth Sci. 19, p. 254-266.
11 Davis D.W. and Edwards G.R. (1985) Summary of Research, Ontario Geoscience Research Grant Program 1984-1985.
12 Davis D.W. and Edwards G.R. (1982) Can. J. Earth Sci. 19, p. 1235-1245.
13 Corfu F. and Stott G.M. (1985) (abstract) Institute on Lake Superior Geology, 31st annual meeting, Kenora, Ont., p. 15.
14 Corfu F. and Grunsky E.C. (1984) (abstract) Geol. Assoc. Canada annual meeting, London, Ont., p. 54.
15 Beakhouse G.P., Stott G.M. and Sutcliffe R.H. (1983) Ontario Geological Survey Summary of Field Work, MP116, p. 5-10.
16 Poulsen K.H., Borradaile G.J. and Kehlenbeck, M.M. (1980) Can. J. Earth Sci. 17, p. 1358-1369.
17 Krogh T.E., Davis D.W., Nunes P.D. and Corfu F. (1982) (abstract) Geol. Assoc. Canada annual meeting, Winnipeg, Man., p. 61.
18 Davis D.W. and Edwards G.R. (1985) (abstract) Institute on Lake Superior Geology, 31st annual meeting, Kenora, Ont., p. 20.
19 Langford F.F. and Morin J.A. (1976) Amer. J. Sci. 276, p. 1023-1034.
20 Percival J.A. and Krogh T.E. (1983) Can. J. Earth Sci. 20, p. 830-843.

CRUSTAL STRUCTURE OF THE ARCHEAN GRANITE-GREENSTONE TERRANE IN THE NORTHERN PORTION OF THE KAAPVAAL CRATON

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Recent investigations of the electrical resistivity, gravity and aeromagnetic signatures of the various granite-greenstone units in the northern portion of the Kaapvaal craton have revealed three features of significance: 1) the Archean greenstone belts are shallow features, rarely exceeding 5 km in depth; 2) the high resistivity upper crustal layer typical of the lower grade granite-greenstone terranes is absent in the granulite facies terrane and 3) the aeromagnetic lineation patterns allow the granite-greenstone terrane to be subdivided into geologically recognisable tectono-metamorphic domains on the basis of lineation frequency and direction.

In the Pietersburg, Sutherland and Murchison greenstone belts geoelectrical investigations showed that the greenstone lithologies have a lower resistivity than the surrounding granitic terranes. Positive gravity anomalies over the greenstone belts are related to more dense metamorphosed ultramafic and mafic rocks in the belts compared to surrounding granitic rocks. Numerical modelling of the geophysical data indicates that the greenstone belts are asymmetrical structures, being thicker along the southeastern flanks. The belts are underlain by high resistivity, low density granitic rocks of which two types are distinguished by their average densities: a lower density series (density = 2600 kg m^{-3}) corresponding to 2650 Ma granodioritic plutons and a higher density series (density = 2670 kg m^{-3}) comprising the older gneissic terrane. The younger series is well developed along the southern margins of the greenstone belts and occurs locally along the northern margins. Primary layering and tectonic fabric within the greenstone lithologies are subvertical. Thicknesses measured across layering exceed the depth of

the belts, suggesting no simple rotation of the greenstone lithologies but instead a truncation at shallow depths of structurally repeated (folded and imbricated) greenstone belts. This truncation may be a major recumbent deformation zone, recumbent syntectonic granite or a late intrusive contact.

Deep resistivity soundings indicate significant changes in the regional structure of the crust in the northern portion of the Kaapvaal craton corresponding to changes in metamorphic grade and tectonic style. In the low-grade granite-greenstone terrane the upper 10 km or less of the crust is characterized by high-resistivity rocks (approximately 100 000 ohm.m) overlying a more conductive layer (approximately 5 000 ohm.m) to a depth of about 35 km. Below this possible mantle rocks with a resistivity of about 50 ohm.m occur. Where the granulite facies rocks of the northernmost Kaapvaal craton (southern marginal zone of the Limpopo belt) occur the approximately 100 000 ohm.m layer is absent and rocks with a resistivity of 5 000 ohm.m extend to a maximum depth of 35 km, where they overlie possible mantle rocks. The significance of these variations in the physical properties of the Kaapvaal craton will be addressed.

The aeromagnetic lineation pattern in the study area can be divided into distinct domains on the basis of the lineation frequency and direction. Although these magnetic anomalies are due to mafic and ultramafic dykes they reflect an inherent fabric in the crust. The domain boundaries correspond to known tectonic and/or metamorphic transitions. One such boundary being the orthoamphibole rehydration isograd that marks the transition between the granulite facies terrane of the southern marginal zone of the Limpopo belt (northern Kaapvaal craton) and the lower-grade rocks to the south. It is clear that the lineation pattern does not reflect the different lithological units in the area and the Sutherland and Pietersburg greenstone belts are, for example, not reflected in the aeromagnetic lineation pattern. This suggests that they are an internal component of certain domains and not marking domain sutures.

EXTENSIONAL TECTONICS DURING THE IGNEOUS EMPLACEMENT OF THE MAFIC-ULTRAMAFIC ROCKS OF THE BARBERTON GREENSTONE BELT. M.J. de Wit, Lunar and Planetary Institute, 3303 NASA Road One, Houston, TX 77058 and BPI Geophysics, University of the Witwatersrand, Johannesburg, South Africa.

The simatic rocks (Onverwacht Group) of the Barberton greenstone belt, which occur in at least 3 regional thrust nappes, are part of the Jamestown ophiolite complex[1]. This ophiolite, together with it's thick sedimentary cover (Fig Tree and Moodies Groups) occupies a complex thrust belt. Field studies have identified two types of early faults which are entirely confined to the simatic rocks and are deformed by the later thrusts and associated folds. The first type of fault (F1a) is regional and always occurs in the simatic rocks along and parallel to the lower contacts of the ophiolite-related cherts (Middle Marker and equivalent layers; for their distribution see Fig. 1, de Wit et al., this volume). These faults zones have previously been referred to both as flaser-banded gneisses[2] and as weathering horizons[3]. (F1a) zones consist of anastomosing, cross-cutting and folded extension veins which have internal cross-fibrous growth textures. Vein filling minerals are predominantly calcite, less often quartz. The veins are separated by schistose to proto-mylonitic folia of fuchsite, chlorite, sericite and serpentines (Fig. 1). In general the zones range between 1-30m in thickness. The veins formed by a succession of dilation-diffusion increments[4,5] and subsequently deformed during simple shear to form banded gneisses (Fig. 1; in this poster presentation, polished slabs of these rocks will be displayed). The simatic host rocks close to (F1a) zones, are ubiquitously brecciated and extensively altered (carbonatized and/or silicified) as documented by the major elements, stable and radiogenic isotope compositions (REE are relatively stable). This alteration is related to a extensive hydrothermal-fluid/rock interaction. It has been postulated that the dilatancy-anisotropy of the fault zones was related to a hydraulic fracturing-gliding mechanism in a geothermal environment[6]. Episodic decrease of fluid overpressure due to movement in these zones would cause boiling, calcite precipitation and crack-sealing with a concomittant resistance to movement of the cherty cap-rock[6]. Displacements along these zones are difficult to estimate, but may be in the order of $1\text{-}10^2$ km. The structures indicate that the faults formed close to horizontal, during extensional shear and were therefore low angle normal faults. In many areas, both the faults and their overlying cherts, are cut by subvertical simatic intrusions of the Onverwacht Group (Fig. 2). Thus (F1a) zones overlap in age with the formation of the ophiolite complex. The second type of faults (F1b) are vertical brittle-ductile shear zones, which crosscut the complex at variable angles and cannot always be traced from plutonic to overlying extrusive (pillowed) simatic rocks. (F1b) zones are therefore also apparently of penecontemporaneous origin with the intrusive-extrusive igneous processes (Fig. 2). Thus (F1b) zones may either represent transform fault-type activity or represent root zones (steepened extensions) of (F1a) zones. Both fault types indicate extensive deformation in the rocks of the greenstone belt prior to compressional overthrust tectonics, and at least (F1a) implies regional extensional tectonics and probably block rotation during the formation of the ophiolites.

References

(1) de Wit et al. this volume. (2) de Wit (1982) Journal of Structural Geology, 4, p. 117-136. (3) Lowe et al. (1985) Precambrian Research, 27, p. 165-186. (4) Ramsay, J.G. (1980) The crack-seal mechanism of rock deformation. Nature, 284, p. 135-139. (5) Ramsay, J.G. and Huber, M.I. (1983) Strain analysis. Academic Press. (6) de Wit et al., Economic Geol., 77, p. 1783-1802.

de Wit, M. J.

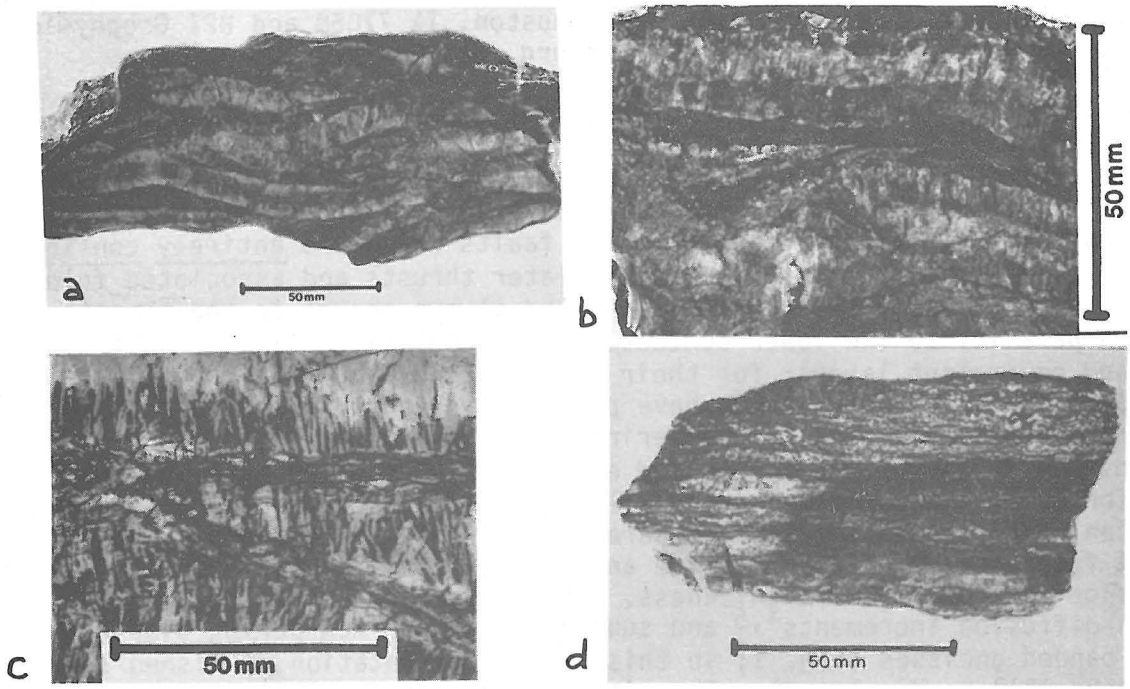


FIG1

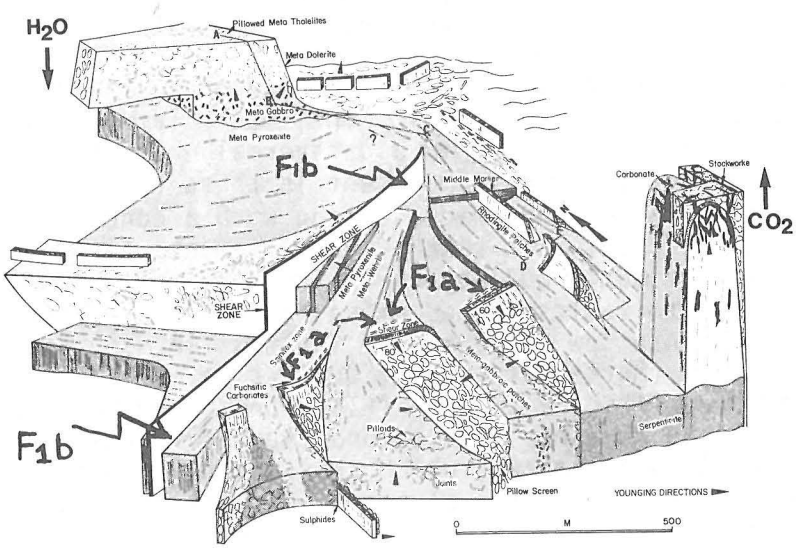


FIG. 2

Figure Captions: Fig. 1 (a) Anastomosing/crosscutting carbonate extension veins (pale-grey) with thin schistose folia (dark grey). Sections up to 30 meters thick entirely composed of this rock-type constitute flaser-banded tectonites[2]. (b) as in (a), showing the cross-fibrous carbonate growth textures in the veins. Different shades of grey are due to variations in concentration of inclusion bands and trails[4]. (c) Internal brecciation and shearing of cross-fibres (vertical) yielding (subhorizontal) protomylonites. (d) gneissose-mylonitic fabric following shearing and flattening of extension veins. (2) Block diagram of area near the Onverwacht bend (see Fig. 1, de Wit this volume for location - as outlined by the box marked Fig. 2) showing the disposition of both (F1a and F1b) fault zones. Note how vertical metawehrlite intrusions (grey) have cross-cut and incorporated screens of middle marker-like cherts underlain by (F1a) gneissose tectonites (shear zones).

A MID-ARCHEAN OPHIOLITE COMPLEX, BARBERTON MOUNTAIN LAND; M.J. de Wit, Lunar and Planetary Institute, 3303 NASA Rd. One, Houston, TX 77058 and BPI Geophysics, University of the Witwatersrand. Roger Hart, School of Oceangraphy OSU, Corvallis, Oregon 97331. Rodger Hart, SC Nuclear Sciences, University of the Witwatersrand, Johannesburg, South Africa.

New field observations and structurally restored geologic sections through the southern part of 3.5-3.6 Ga Barberton greenstone belt (Fig. 1) show that it's mafic to ultramafic rocks form a pseudostratigraphy comparable to that of Phanerozoic ophiolites; we refer to this ancient ophiolite as the Jamestown ophiolite complex[1]. It consists of an (in part sheeted, Fig. 2) intrusive-extrusive mafic-ultramafic section, underlain by a high-temperature tectono-metamorphic residual peridotitic base, and is capped by a chert-shale sequence which it locally intrudes. Geochemical data support an ophiolitic comparison (Fig. 3). Fractionation of high temperature melting PGE's (> 2500°C) in the residual rocks suggest a lower mantle origin for the precursors this crust[2]. An oceanic rather than arc-related crustal section can be inferred from the absence of contemporaneous andesites. This ancient simatic crust was thin (<3 km), contains a large ultramafic component ($\simeq$25%), is pervasively hydrated (> 95%) with H_2O contents ranging between 1-15% and consequently has a low density ($\simeq 2.67$ g/cm^3)[3].

The entire simatic section has also been chemically altered during its formation by hydrothermal interaction with the Archaean hydrosphere (Fig. 4). Only an igneous "ghost" major element geochemistry is preserved. This regionally open-system metasomatism may have increased the MgO content of the igneous rocks by as much as 15%. The most primitive parent liquids, from which the extrusive sequence evolved, may have been "picritic" in character. Rocks with a komatiitic chemistry may have been derived during crystal accumulation from picrite-crystal mushes (predominantly olivine-clinopyroxene) and/or by metasomatism during one or more subsequent episodes of hydration-dehydration (Fig. 5).

The Jamestown ophiolite complex provides the oldest record with evidence for the formation of oceanic lithosphere at constructive tectonic boundaries. Our observations are in agreement with models predicting higher oceanic Archean heat flux per unit ridge length than today, associated with deep mantle diapiric upflow. Because of its low density, this ophiolite resisted subduction during subsequent tectonism; it was obducted to form part of a thrust complex[3].

References

(1) de Wit, M.J., Hart, R. and Hart, R. (1985) The Jamestown ophiolite complex, Barberton Mountain Belt: a reconaissance study of a composite section through 3.5 Ga simatic lithosphere (under review). (2) Tredoux, M., de Wit, M.J., Hart, R., Armstrong, R., Sellschop, J.P.E., (1985) PGE in the 3.6 Bon Accord Ni-Fe-Co-Cr occurrence, Barberton South Africa: implications for crust-mantle-core segregation. 4th Intern. Platinum Symposium, Toronto, August, 1985, abstract; also de Wit et al. (1985), EOS, Nov. 12, p. 1114. (3) de Wit M.J. (1985) What the oldest rocks say, (abstract) Workshop on The Earth as a Planet, LPI and GSA, Orlando, Florida, Oct. 27.

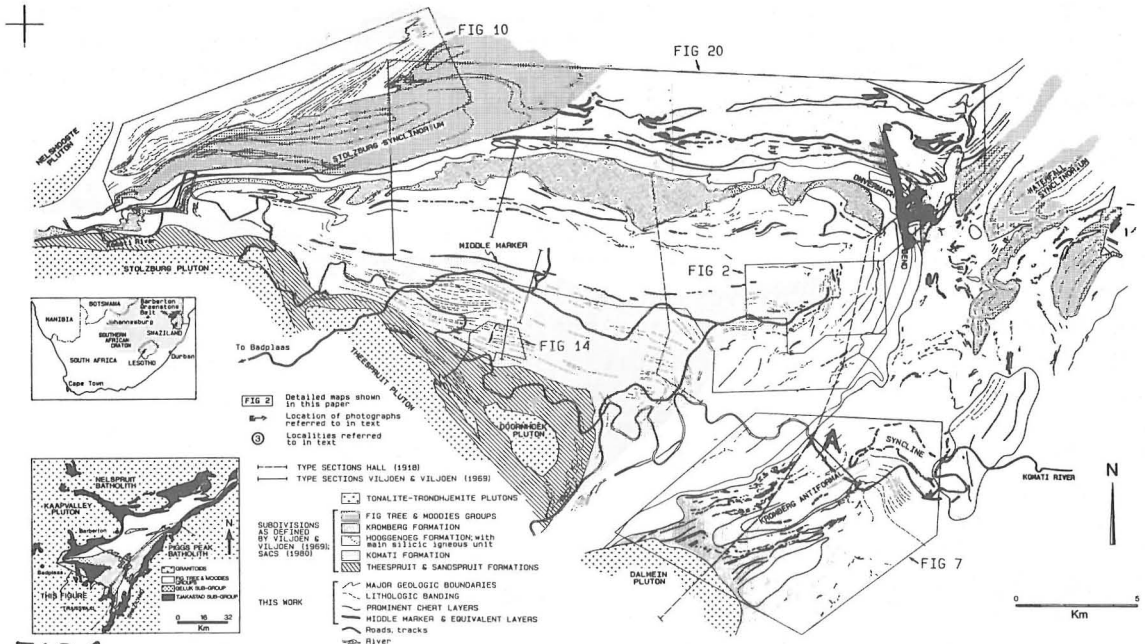


FIG 1

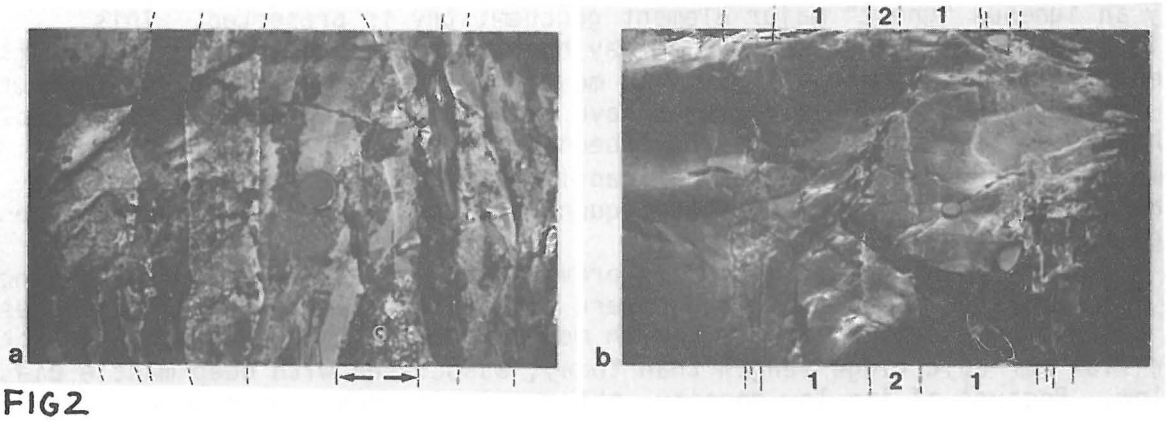


FIG2

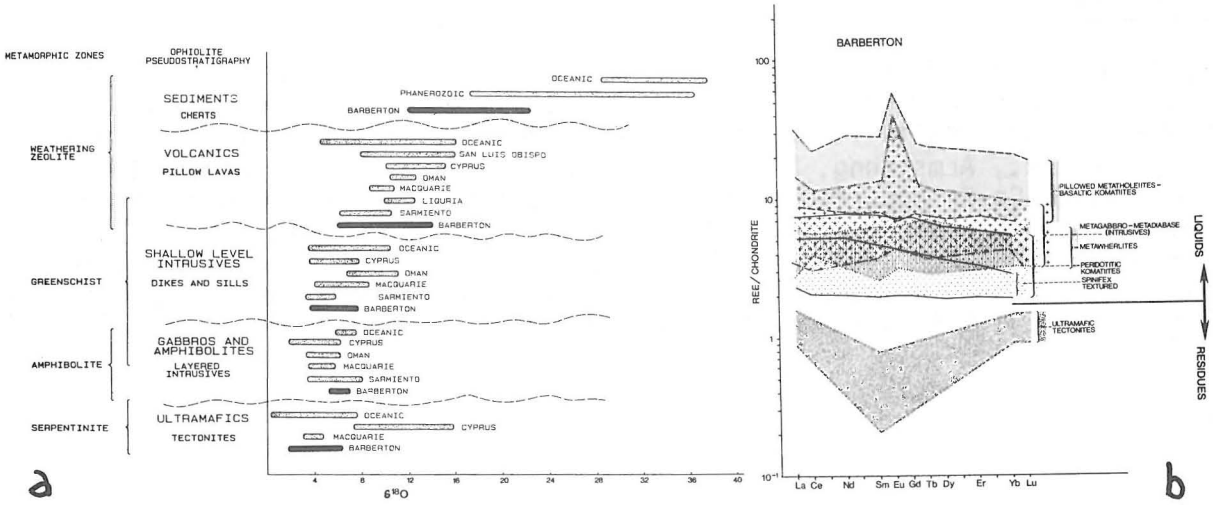


FIG 3

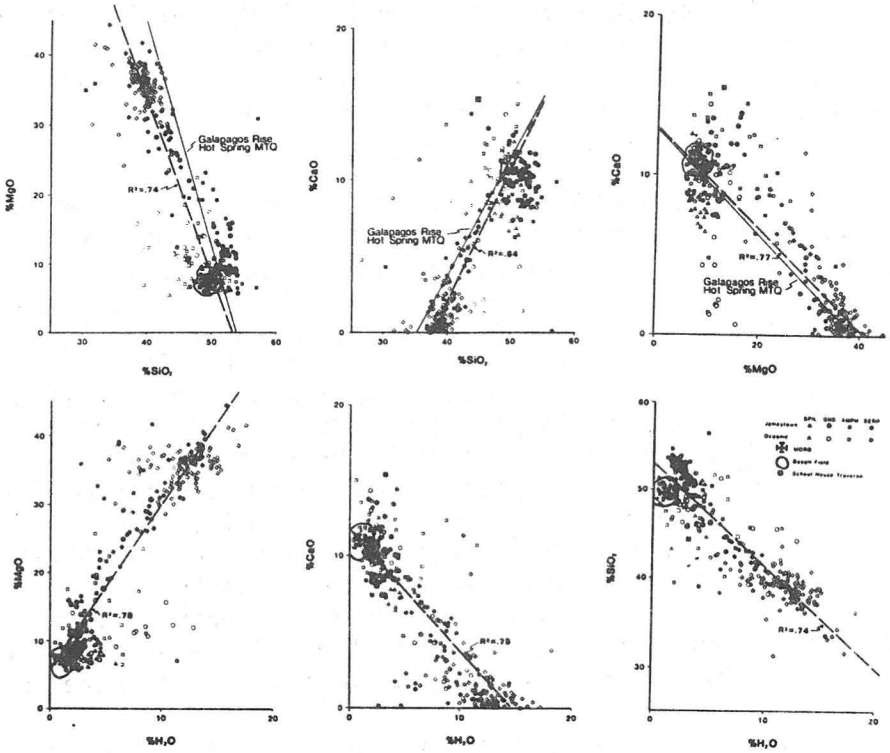


FIG.4

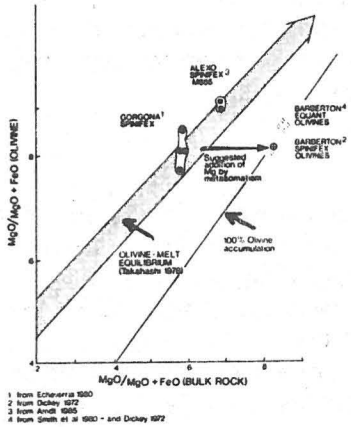


FIG.5

Figure Captions: (1) Simplified geological map of an area in the southern part of the Barberton greenstone belt studied between 1978 and 1985. (2) Vertical sheeted intrusives with pale chilled margins from a 30 meter river outcrop (exposed during 1984 draught) at locality A, Fig. 1. Note the remnant chert xenolith (a; arrow). (b) clearly depicts the splitting in two of an earlier intrusion (1) by a later one (2). (3) Representation of $\delta^{18}O$ (a) of the Barberton rocks (black) plotted in their restored pseudo-stratigraphic sequence compared to Phanerozoic ophiolites and oceanic crust (open symbols) (b) REE data from Barberton; this plot compares favourably with Phanerozoic ophiolites and oceanic crust. (4) Binary correlation plots of MgO, CaO, SiO, and H_2O for rocks of oceanic crust (open symbols) and from the Jamestown ophiolite complex (closed symbols). These plots illustrate the close correlation between the major oxides concentrations and the degree of hydration in these environments. For comparison, the slopes of the chemical flux in the Galapages hot spring fluids are also shown. (5) This figure shows that the bulk rock MgO/MgO + Feo ratio of Barberton Komatiites are enriched in MgO over that of the original melt. The enrichment may be the result of either crystal accumulation or magnesium metasomatism during hydrothermal alteration; there is textural evidence that both mechanism were important. At any rate the plot clearly shows that the MgO composition of the silicate liquids which formed the Barberton Komatiite was between that of Gorgona Island (15-22% MgO) and Alexo (28% MgO), and may have been of picritic composition. All diagrams from reference 1.

FELSIC IGNEOUS ROCKS WITHIN THE BARBERTON GREENSTONE BELT: HIGH CRUSTAL LEVEL EQUIVALENTS OF THE SURROUNDING TONALITE-TRONDHJEMITE TERRAIN, EMPLACED DURING THRUSTING. M.J. de Wit, Lunar and Planetary Institute, 3303 NASA Road One, Houston, TX 77058 and BPI Geophysics, University of the Witwatersrand, Johannesburg. A.H. Wilson, University of Natal, Pietesmaritzburg, South Africa.

Felsic rocks within the 3530 ± 50 myrs[1] simatic rocks of the Onverwacht Group of the Barberton greenstone belt have traditionally been mapped as recurring volcanic units within a continuous stratigraphic succession. In the past, these felsic units have been interpreted to be part of several mafic to felsic volcanic cycles within this sequence. Some of these silicic layers have been shown to be silicified simatic rocks[2,3]. Our field data (Fig. 1) indicates that the genuine felsic igneous rocks are predominantly shallow level intrusives and subsurface felsic domes associated with only minor volcanics and volcanoclastics. A 3.360 ± 1 myrs (U-Pb, zircon)[4] age from the main felsic intrusion indicates that its emplacement post-dated the simatic rocks of this greenstone belt between 120-220 myrs. Our geochemical results also show that the felsic igneous rocks are not directly related to the mafic-ultramafic rocks of the Onverwacht Group. On the contrary the major trace and REE data (Fig. 2) all indicate that these felsic units are high-level equivalents of the widespread, and time-equivalent, trondhjemite-tonalite plutons which either intrude the lower parts of the greenstone belt, or with which they are in tectonic contact.

Structural and stratigraphic analysis indicates that the felsic intrusions were emplaced along thrusts during sedimentation and a prolonged period of horizontal compressional stress exerted on the greenstone belt (Fig. 3). Thus, integrated, the data suggest that the simatic rocks of the Barberton greenstone belt were thrust across an actively stoping plutonic environment and that the greenstone belt is at least partly allochthonous (Fig. 4).

References

(1) Hamilton, P.J. et al. (1979) Sm-Nd dating of the Onverwacht Group volcanics, South Africa. Nature, 279, p. 298-300. (2) de Wit et al. (1982) Archean abiogenic and probable biogenic structures associated with mineralized hydrothermal vent systems and regional metasomatism with implications for greenstone belt studies. Econ. Geol., 77, p. 1783-1802. (3) Hanor, J.S. and Duchac, K.C. (1985) Geochemical consequences of the metasomatic conversion of an early Archean Komatiite sequence into chert. GSA Abstracts, Geol. Annual Meeting, Orlando, p. 603. (4) Van Niekerk, C.B. and Burger, A.J. (1969) A note on the mininum age of the acid lava of the onverwacht series of the Swaziland system. Trans. Geol. Soc. South Africa, 72, p. 9-21.

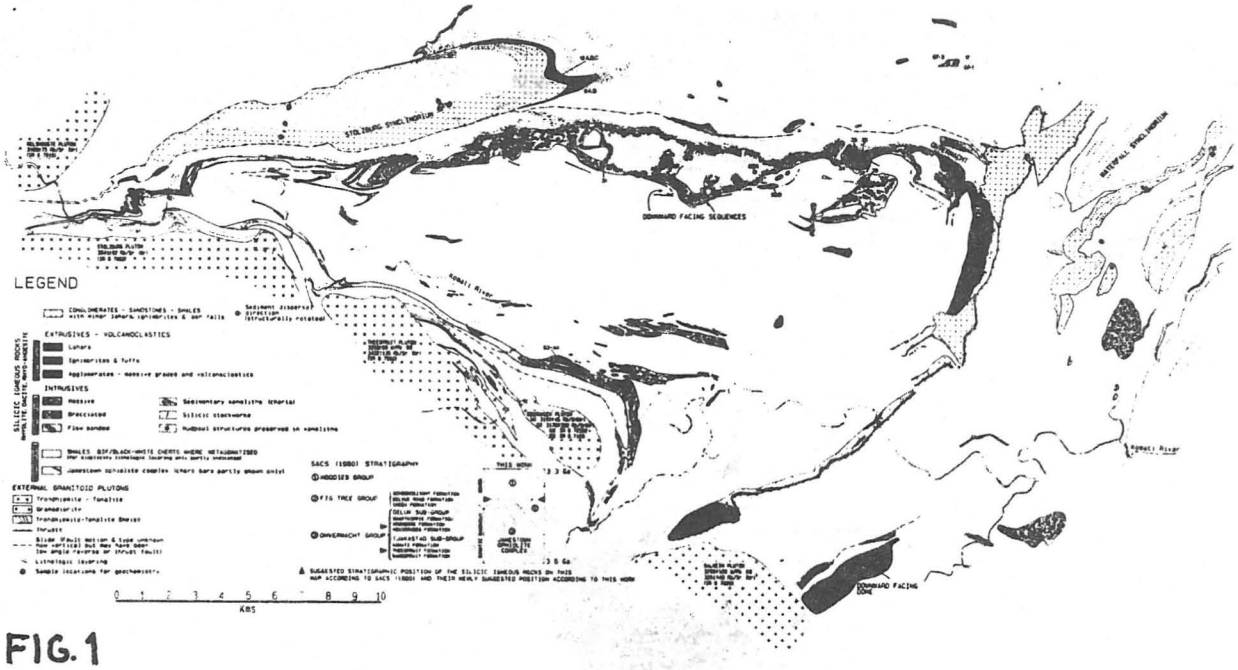


FIG. 1

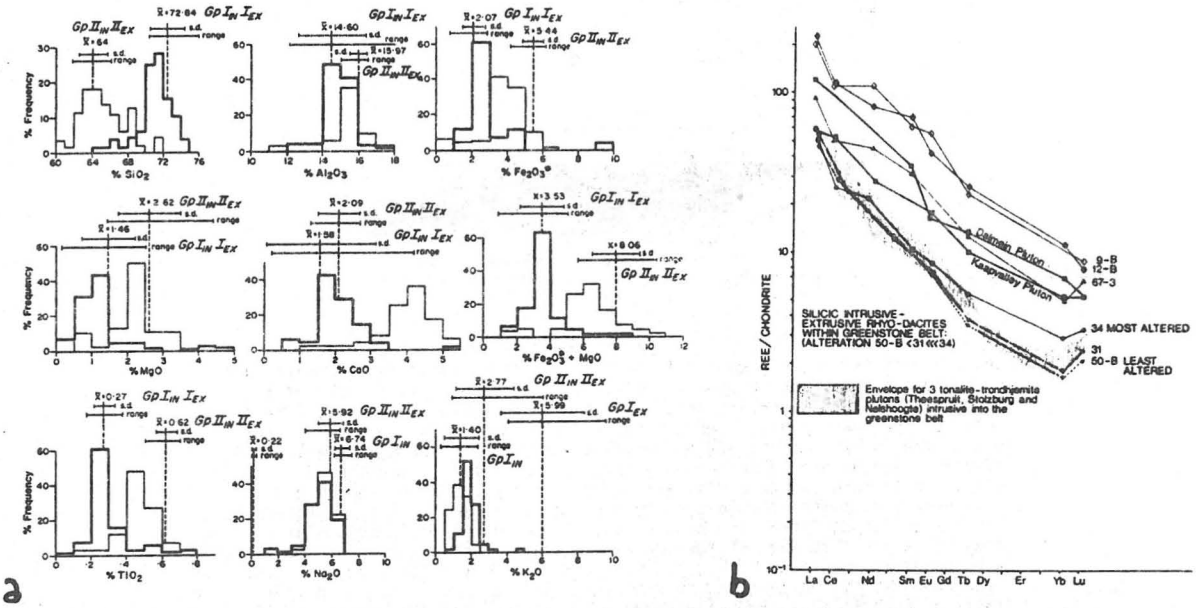


FIG. 2

Figure Captions: (1) Simplified geological map of the southern part of the Barberton greenstone belt, showing location of main silicic (felsic) rocks. (2) (a) Statistical analysis of major element data from the felsic igneous rocks within the study area, compared to those of the surrounding tonalite and trondhjemites. The felsic igneous rocks are clearly divided into two groups (I and II) in which both extrusive (ex) and intrusive (in) samples are represented. The two groups are geochemically similar to the trondhjemites (thin frequency boxes) and tonalites (bold frequency boxes) (b) Chondrite normalized REE patterns of intrusive and extrusive representatives of both groups of felsic igneous rocks from within the greenstone belt, compared to the granitoid plutons surrounding the greenstone belt.

FELSIC IGNEOUS ROCKS EMPLACED DURING THRUSTING

de Wit, M. J. and Wilson, A. H.

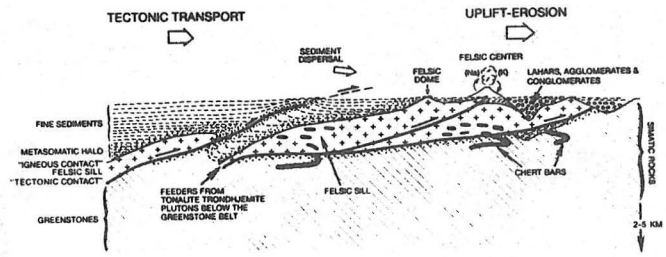


FIG.3

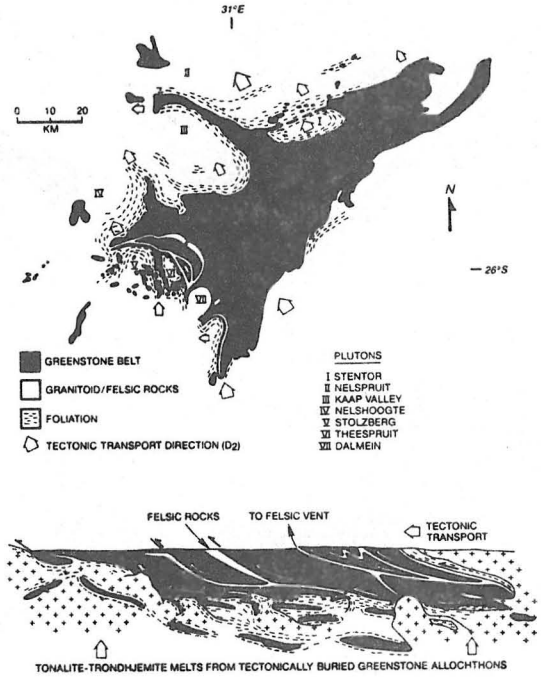


FIG 4

(3) Schematic representation of the tectonic-intrusive emplacement of the felsic igneous rocks as composite sills close to the interface between the mafic-ultramafic (simatic) rocks of the Onverwacht rocks (diagonal lines) and the overlying Fig Tree-like silts-shales. Note how the lower contacts of the sills are predominantly tectonic (thrusts) whilst the upper contacts are predominantly preserved igneous contacts. (4) Plan and section of the Barberton greenstone belt (black) and the surrounding granitoid terrain (white). The map shows the generalized D_2 tectonic transport directions, the felsic igneous rocks internal to the greenstone belt, and the gneissose fabric in the surrounding tonalite-trondhjemite plutons. Note that large scale stoping of the greenstone belt by the surrounding and intruding granitoids is suggested by the outcrop pattern of the felsic igneous rocks (eg. compare this pattern to the shape and outline of the Stentor pluton). The section schematically shows the lower parts (3-5 km) of the greenstone belt thrust over the granitoid terrain whilst the latter syntectonically intrudes and engulves the greenstone belt: this process is thought to have formed recumbent-like mantle-gneiss folds (probably sheath-like in 3-dimentions). Regional disruption and stoping of the greenstone belt occurs during intrusion of Na-rich felsic phases from the plutons of the granitoid terrain into the greenstone belts, along thrusts generated during the tectonic emplacement of the entire greenstone belt. The section represents a restoration prior to subsequent horizontal flattening which later deformed and rotated the rock units and their contacts into a pseudo-synformal structure. All diagrams from de Wit, Wilson and Armstrong (1985 under review).

HEAT FLOW AND HEAT GENERATION IN GREENSTONE BELTS; Malcolm J. Drury, Earth Physics Branch, Energy, Mines and Resources Canada, 1 Observatory Crescent, Ottawa, Ontario, K1A 0Y3 Canada

Heat flow has been measured in Precambrian shields in both greenstone belts and crystalline terrains. Values are generally low, reflecting the great age and tectonic stability of the shields; they range typically between 30 and 50 mW/m^2, although extreme values of 18 and 79 mW/m^2 have been reported (1,2). For large areas of the earth's surface that are presumed to have been subjected to a common thermotectonic event, plots of heat flow against heat generation appear to be linear (3,4), although there may be considerable scatter in the data. The relationship is expressed as:

$$Q = Q_o + D\ A_o \qquad [1]$$

in which Q is the observed heat flow, A_o is the measured heat generation at the surface, Q_o is the "reduced" heat flow from the lower crust and mantle, and D, which has the dimension of length, represents a scale depth for the distribution of radiogenic elements. Most authors have not used data from greenstone belts in attempting to define the relationship within shields, considering them unrepresentative and preferring to use data from relatively homogeneous crystalline rocks, e.g. (5).

The heat generated by radioactive decay is expected to be less in basic than in acidic rocks because of their different chemistry. Hence we would expect heat flow in greenstone belts to be lower than that in adjoining crystalline areas if the greenstones are thick, but to be similar if the belts are merely superficial. Table 1 is a compilation of data from seven Precambrian shields. Only those data specifically identified as being from greenstone belts, or those for which geological descriptions are unambiguous, are used in column 2. There is the possiblity that some of the data identified as being from crystalline areas are in fact from greenstone belts.

Table 1. Compilation of heat flow data for Precambrian shields, listed according to geological setting. The ratio in column 4 is that of the mean heat flow in the greenstone belts to that in crystalline areas of the shield.

Shield	Mean and 1 s.d. heat flow (mW/m^2)			
	All sites	Greenstones	Crystalline	Ratio
Canadian*[a]	42±8 (22)	39±5 (8)	43±10 (14)	0.91
Canadian*[b]	43±10 (10)	40±9 (6)	48±11 (4)	0.83
Baltic[c]	40±6 (26)	41±6 (4)	40±6 (22)	1.03
W. African[d]	36±12 (19)	35 (1)	38±11 (18)	0.92
Indian[e]	64±15 (6)	44 (1)	68±12 (5)	0.65
Australian[f]	40±8 (16)	38±8 (8)	42±8 (8)	0.90
Brazilian[g]	52±11 (12)	51±18 (2)	52±10 (10)	0.98

[a] Superior province, reference 6 with additional data not yet published; [b] Churchill province, 1; [c] 5,7,8,9,10,11; [d] 12; [e] 2; [f] 13, 14, 15; [g] 16. * - heat flow values adjusted for glaciation effects.

Although it appears from column 4 of Table 1 that mean heat flow in greenstone belts is indeed lower than that in crystalline areas of the shields, there is, in all shields except one, considerable overlap of the two values. The exception is the Indian Shield, but there is only one value from a greenstone belt for that. Further, in most cases no statistical signficance can be inferred as there are fewer data for greenstone belts than for crystalline areas. Taking the mean values, the heat flow from crystalline areas is apparently approximately 10% higher than that from greenstone belts.

Not all heat flow data used for compiling Table 1 had associated heat generation data. The most complete set is for the Canadian shield (Superior and Churchill provinces). Linear least squares regression for those data yields:

	Q_o (mW/m^2)	D km	r	Q^* (mW/m^2)	A^* ($\mu W/m^3$)	n
Greenstone belts	33±4	7±6	0.45	37±7	0.51±0.46	7
Crystalline areas	26±6	12±4	0.67	40±9	1.16±0.51	11

where n is the number of data pairs, r is the correlation coefficient, Q^* is the mean heat flow and A^* is the mean heat generation of borehole samples. The correlations are low and statistically the differences between the parameters for the two crustal types are insignificant. However, assuming that radiogenic elements are distributed uniformly with depth to D, the value of D for the greenstones suggests that they are approximately 7 km thick, a value compatible with those cited by Condie (17). The data also suggest that the heat flow - heat generation relationship for the greenstones could be written as

$$Q = Q_o = (D_c - D_g)A_c + D_g A_g \quad [2]$$

in which subscripts g and c refer to greenstone and crystalline crust and Q_o is the reduced heat flow for the crystalline crust. This can be seen by inserting appropriate values for greenstones and crystalline terrain into equation [2]. It implies that the greenstones are underlain by normal crystalline crust, including 5 km of upper crust, but that they are not allochthonous, replacing 7 km of that crust rather than simply overlying it.

2 Rao R.U.M., Rao G.V. and Narain H. (1976) Earth Plan. Sci. Lett., 30, p. 57-64.

3 Roy R.F., Blackwell D.D. and Birch F. (1968) Earth Plan. Sci. Lett., 5, p. 1-12.

4 Lachenbruch A.E. (1968) J. Geophys. Res. 73, p. 6977-6989.

5 Swanberg C.A., Chessman M.D., Simmons G., Smithson S.B., Gronlie G. and Heier K.S. (1974) Tectonophysics, 23, p. 31-48.

6 Jessop A.M. and Lewis T.J. (1978) Tectonophysics, 50, p. 55-77

7 Puranen M., Jarviamki P., Hamaleinen U. and Lehtinen S. (1968) Geoexploration, 6, p. 151-162

8 Lubimova E.A., Karus E.V., Firsov F.V., Starikova G.N., Vlasov V.K., Lyusova L.N. and Koperbakh E.B. (1972) Geothermics, 1, p. 81-89

9 Parasnis D.S. (1975) Geophys. J. Roy. Astr. Soc., 43, p. 531-554

10 Parasnis D.S. (1982) Geophys. J. Roy. Astr. Soc., 71, p. 545-554

11 Gordienko V.V., Zavgorodnya O.V., Moiseenko U.I. and Smyslov A.A. (1985) Geophys. Res. Lett., 12, p. 549-552

12 Brigaud F., Lucazeau F., Ly, S. and Sauvage, J.F. (1985) Geophys. Res. Lett., 12, p. 549-552

13 Howard L.E. and Sass J.H. (1964) J. Geophys. Res., 69, p. 1617-1625

14 Hyndman R.D., Lambert I.B., Heier K.S., Jaeger J.C. and Ringwood A.E. (1968) Phys. Earth Plan. Int., 1, p. 129-135

15 Jaeger J.C. (1970) Earth Plan. Sci. Lett., 8, p. 285-292

16 Viterello I., Hamza V.M. and Pollack H.N. (1980) J. Geophys. Res., 85, p. 3778-3788.

17 Condie, K. (1981) Developments in Precambrian geology 3, Elsevier Scientific Publishing Co., New York.

KOMATIITE GENESIS IN THE ARCHAEN MANTLE, WITH IMPLICATIONS FOR THE TECTONICS OF ARCHAEN GREENSTONE BELTS; D. Elthon, Lunar and Planetary Institute, 3303 NASA Rd.1, Houston, TX 77058 and Department of Geosciences, University of Houston — University Park, Houston, TX 77004.

The presence of ultramafic lavas (komatiites) associated with Archaen greenstone belts has been suggested to indicate very high increments (50-80%) of partial melting of the Archean mantle [e.g., 1-3]. Such extensive melting of the Earth's mantle during the Archaen might have profound effects on the early tectonic and chemical evolution of the planet [e.g., 4 & 5], although problems associated with keeping the komatiite liquid in equilibrium with the residual mantle at such high increments of melting has cast doubt upon aspects of extensive melting [e.g.,6 & 7]. Two important aspects of the origin of komatiites are discussed below.

I. WHAT IS THE NATURE OF PRIMARY KOMATIITE LIQUIDS?

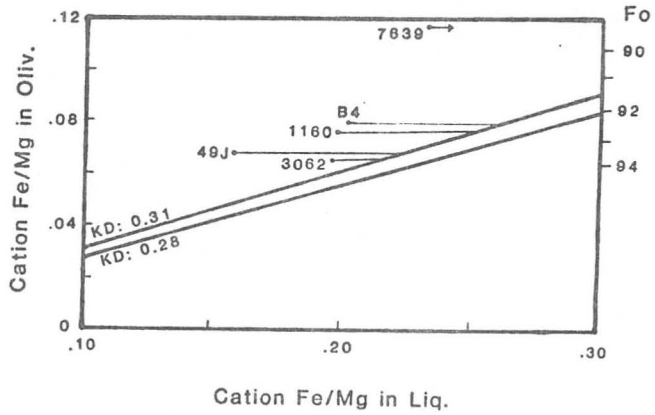
One of the most fundamental aspects of understanding the tectonic and geochemical mode of origin for komatiites is the problem of komatiite primary magmas. The identification of primary komatiite magmas is complicated by the extensive metamorphism that these rocks have typically undergone and by olivine (+ minor spinel) crystallization at low pressures (~1 atm). The crystallization of olivine rapidly depletes a komatiite liquid in MgO, such that the most likely candidates for primary magmas are those with the highest MgO contents.

Previous efforts to evaluate primary komatiitic liquids have proposed that they might contain as much as 33% MgO [2] or 30% MgO [8]. These studies have relied principally on comparison of the compositions of olivines crystallized in high-pressure experimental studies of komatiites with relict olivines found in komatiites (as high as ~Fo94).

The Fe-Mg exchange between olivine and basaltic-komatiitic liquids has recently been summarized by [9], in which they present equations for calculation of olivine-liquid equilibria over a wide temperature (1074-1600° C.) and pressure (1 atm to 25 kbar) range. The KD values for a wide range of komatiites (>20% MgO) were calculated using this equation and range from 0.28 to 0.31 at temperatures of 1450-1650°C at 1 atm. This olivine-liquid equilibrium is shown in Fig. 1 along with the compositions of the most magnesian olivines in komatiites (olivine and komatiite compositions from [8] and references therein). The 1 atm KD values have been used here because the present author considers it most likely that the olivines in komatiites have crystallized at very low pressures (~1 atm); previous investigators [2 & 4] have used KD values from high pressure experiments, which are substantially higher [10 & 11].

The data shown in Fig. 1 (horizontal lines connect the olivine compositions with the liquid from which they could have crystallized) indicate that the komatiite olivines probably have crystallized from liquids with Fe/Mg >0.230. This Fe/Mg (0.230) corresponds to 22-25% MgO in the komatiites, depending upon the FeO content of the liquid. This data indicates that the most magnesian olivines in komatiites could have crystallized from liquids with 22-25% MgO, in contrast to previous estimates of 30-33% MgO. These liquids will have liquidus temperatures of ~1500°C at 1 atm pressure. More MgO-rich komatiites have probably become enriched in MgO as a consequence of olivine accumulation and/or Mg metasomatism.

Fig. 1
Cation Fe/Mg in olivine versus cation Fe/Mg in the liquid. The horizontal lines connect the composition of olivines in komatiites with liquids in equilibrium with these olivines.



II. WHAT PERCENTAGE OF MELTING IS REQUIRED TO PRODUCE KOMATIITES?

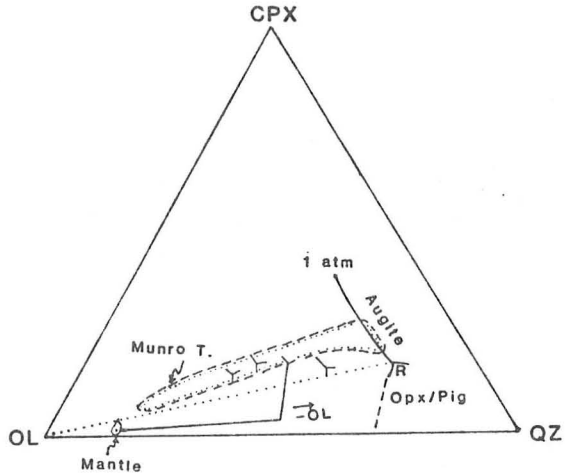
As noted above, it is generally assumed that a very high increment of melting (50-80%) is required in order to generate komatiites from the Earth's mantle. Experimental studies of the melting of reasonable mantle compositions have shown that very magnesium-rich magmas may be produced at high increments of melting [e.g., 11-13]. As outlined below, however, these MgO-rich magmas produced by very large increments (40-80%) of melting within the mantle are NOT komatiites.

A pseudo-liquidus phase diagram for evaluating the petrogenesis of komatiites is shown in Fig. 2. At low pressures (~1 atm), some magmas (those above the Ol-R join) will crystallize augite as the first pyroxene and others (those below the OL-R join) will crystallize pigeonite or orthopyroxene first. Field and petrographic studies of komatiites have shown that they crystallize augite as the first pyroxene in virtually all instances. Most terrestrial magmas also crystallize augite as the first pyroxene; boninites are an obvious exception. Also shown in Fig. 2 is the field for the compositions of komatiites from Munro Township, which crystallize augite as the first pyroxene [14].

A partially schematic melting path for melting of the mantle is shown in Fig. 2 for melting at 15 kbars. At small to moderate increments of melting (<~30%), the primary liquids will lie above the OL-R join, but will lie below the OL-R join at larger increments of melting. The extent of melting required to produce primary magmas below the OL-R join will vary as a function of the composition of the mantle and the pressure of melting, but it is clear that high increments of melting that might produce dunite or OPX-poor harzburgite residues will produce primary magmas that will lie below the OL-R join and will evolve to crystallize orthopyroxene and/or pigeonite before augite. The extent of melting most likely to produce komatiitic magmas is more like 20-25% rather than the 50-80% previously proposed. Although not discussed by the previous authors, this feature is further apparent in the data of [12 & 13].

Spinels from komatiites have Cr/(Cr+Al) from ~0.70 to 0.80 [15], which would suggest a slightly higher increment of partial melting of the mantle that occurs in the present-day suboceanic mantle [16], rather than the much higher increments proposed in previous studies.

Fig. 2
Pseudo-liquidus phase diagram projected from plagioclase[17]. Pseudo-invariant points at 1 atm (R), 10, 15, 20 and 25 kbars are shown.



In summary, it is suggested that the extent of partial melting that produces komatiite primary magmas is ~20-25% and that these magmas have 22-25% MgO or less. This substantially lower estimate for the extent of melting and eruption temperatures will certainly influence those tectonic characteristics of greenstone belts associated with the dynamics of mantle upwelling and convection.

REFERENCES:

[1] Viljoen, R.P. and Viljoen, M.J. (1969) Geol. Soc. S. Afr., Spec. Publ. 2, 275-296.
[2] Green, D.H., Nichols, I.A., Viljoen, M.J. and Viljoen, R.P. (1975) Geology 3, 11-15.
[3] Bickle, M.J., Hawkesworth, C.J., Martin, A., Nisbet, E.G. and O'Nions, R.K. (1976) Nature 263, 577-580.
[4] Bickle, M.J. (1978) Earth Planet. Sci. Lett. 40, 301-315.
[5] Nisbet, E.G. and Walker, D. (1982) Earth Planet. Sci. Lett. 60, 105-113.
[6] Arndt, N.T. (1977) Contrib. Mineral. Petrol. 64, 205-221.
[7] Waff, H.S. (1980) J. Geophys. Res. 85, 1815-1825.
[8] Bickle, M.J. (1982) in: Komatiites N.T.Arndt & E.G.Nisbet (eds). pp.479-494.
[9] Ford, C.E., Russell, D.G., Craven, J.A. and Fisk, M.R. (1983) J. Petrol. 24, 256-265.
[10] Grover, J.E., Lindsley, D.H. and Bence, A.E. (1980) Proc. Lunar Sci. Conf. 11, 179-196.
[11] Mysen, B.O. and Kushiro, I. (1977) Am. Mineral. 62, 843-865.
[12] Jaques, A.L. and Green, D.H. (1980) Contrib. Mineral. Petrol. 73, 287-310.
[13] Takahashi, E. and Kushiro, I. (1983) Am. Mineral. 68, 859-879.
[14] Arndt, N.T., Naldrett, A.J. and Pyke, D.R. (1977) J. Petrol. 18, 319-369.
[15] Cameron, W.E. and Nisbet, E.G. (1982) in: Komatiites N.T.Arndt & E.G. Nisbet (eds.) pp. 29-50.
[16] Dick, H.J.B. and Bullen, T. (1984) Contrib. Mineral. Petrol. 86, 54-76.
[17] Elthon, D. and Scarfe, C.M. (1984) Am. Mineral. 69, 1-15.

THE YILGARN CRATON WESTERN AUSTRALIA : A TECTONIC SYNTHESIS : I

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The Yilgarn Craton in Western Australia is one of the larger contiguous preserved Archaean crustal fragments, with an area of about 650,000 square kilometres. Of this, by area, about 70% is granitoid and 30% greenstone. The Craton is defined by the Darling Fault on its western margin, by Proterozoic deformation belts on its southern and northwestern margins, and by unconformable younger sediments on its eastern and northeastern margins.

A regional geotectonic synthesis at a scale of 1:500,000 is being prepared. This is based largely upon the 1:250,000 scale mapping of the Geological Survey of Western Australia together with interpretation using geophysical data, mainly airborne magnetic surveys.

On a regional basis the granitoids are classified as pre-, syn- and post-tectonic (1) with respect to greenstone belt deformation. The post-tectonic granitoids yield Rb-Sr isochrons of about 2.6 b.y., close to Rb-Sr ages for the greenstones themselves which are up to about 2.8. b.y. old (2), although data for the latter is sparse.

Contacts between earlier granitoids and greenstones which are not obscured by the post-tectonic granitoids are most commonly tectonic contacts, intensely deformed and with mylonitic fabrics. The general concensus however is that there is a pre-tectonic, pre-greenstone sialic gneiss preserved in places (1,3).

Existing models for the evolution of the belts involve 3 large basinal structures ("broad elongate downwarps"), of which the Eastern one (the Noreseman-Wiluna Belt) is considered to be a rift fault-bounded graben (1). The postulated basins are separated by large tabular belts of discordant post-tectonic granite when viewed regionally. This may be a 'red herring'. It is possible that, for example, the entire greenstone package preserved on the Craton was part of one basin, or numerous combinations and parts of basins. There is no compelling diagnostic evidence collated to date to postulate on the original disposition, geometry and relationships between belts.

This synthesis is a preliminary attempt at addressing this problem, by attempting to decipher the broad tectonic-stratigraphic sequences preserved and thereby to reconstruct, as far as is possible, the original nature of the greenstones. There is structural evidence to suggest that the deformation histories of the greenstones and some of their surrounding and occluded granitoids involves early fold-nappe tectonics in places, and possibly thrust nappes, as well as late large-scale imbrication or slicing. During early deformation of the belts, massif-style nappe tectonics may have occurred in places, on scales not dissimilar to those seen in young fold belts.

It is intended, with future work, to test these postulates and to examine whether the tectonic history of the Yilgarn Craton is indicative of the loss of considerable greenstone (back to the womb?) and perversely (sic), its local preservation by obduction and stacking. How well can we reconstruct the deformed granitoids and greenstones, in their undamaged state?

REFERENCES

1. Gee, R.D., Baxter, J.L., Wilde, S.A. and Williams, I.R. (1981) Spec. Publs geol. Soc. Aust., 7, p. 43-56.

2. de Laeter, J.R., Libby, W.G. and Trendall, A.F. (1981) Spec. Publs geol. Soc. Aust., 7, p. 147-157.

3. Archibald, N.J. and Bettenay, L.F. (1977) Earth planet. Sci. Lett., 33, p. 370-378 .

TECTONICS OF SOME AMAZONIAN GREENSTONE BELTS; Allan K. Gibbs, Institute for the Study of the Continents, and Dept. Geol. Sci., Cornell University, Ithaca, New York, 14853

Greenstone belts exposed amid gneisses, granitoid rocks, and less abundant granulites along the northern and eastern margins of the Amazonian Craton yield Trans-Amazonican metamorphic ages of 2.0-2.1 Ga. (Regional geology: 1-13). Early Proterozoic belts in the northern region probably originated as ensimatic island arc complexes. The Archean Carajas belt in the southeeastern craton probably formed in an extensional basin on older continental basement. That basement contains older Archean belts with pillow basalts and komatiites. Belts of ultramafic rocks warrant investigation as possible ophiolites.

NORTHERN BELTS - Volcanic rocks of the northern belts were erupted in the Early Proterozoic (2.3-2.1 Ga)(14-17). The contiguous belts of Guyana (18,19) and Venezuela (20,21) closely resemble those of Suriname (7-9,22) and French Guiana (1,4,16,23), though the two regions are separated by the Central Guiana Granulite Belt. Typical sections consist of a lower flow and pillowed low-K basalt-gabbro unit, overlain by interbedded mafic, intermediate, and felsic volcanics of both tholeiitic and calc-alkaline suites; overlain by and interstratified with volcaniclastic greywackes, pelites, and chemical sedimentary rocks. Basalts with pronounced iron-enrichment and others with high magnesium contents are both present, as are both tholeiitic and calc-alkaline andesites and felsic volcanics (18,19,22,24). Generally conformable tuffaceous and epiclastic conglomerates, greywackes, lithic arenites, and shales appear petrographically and geochemically to have been derived from the associated volcanic rocks, without significant contributions from continental sources (18,25). The relative abundances and types of volcanic and sedimentary rocks vary: felsic volcanics are irregularly distributed, and magnesian basalts and possible komatiites are particularly common in central French Guiana (22). Ultramafic, mafic, and anorthositic intrusive complexes may be genetically associated with some of the volcanic rocks (1,18,23). Some belts are overlain by quartz-rich epiclastic sedimentary rocks that were folded and metamorphosed with the belts but appear to be unconformable (1,13).

The northern belts have randomly-branching synclinal map patterns. Prominent metamorphic foliations generally correlate with the regional folds, with foliations locally crenulated or destroyed by younger shear deformation, which elongated (WNW-ESE) both the belts and associated granitoid rocks. Metamorphic grades range from amphibolite on the belts' peripheries to lower greenschist and zeolite in the interiors. Diverse local mineral assemblages indicate high, intermediate, and low-pressure metamorphic series. Anatectic, two-mica granites intrude metapelitic schists along the northern periphery.

No evidence has been reported of basement-cover relations between the northern belts and adjacent gneisses. Field observations and geochemical similarities suggest that the greenstones pass into the intervening gneisses by increase in metamorphic grade (14,15,17,26-28). The associated granulites also appear to represent Early Proterozoic, rather than Archean crust (16,27,29). Sm-Nd and Rb-Sr isotopic systematics indicate that little if any older continental crust was involved in this greenstone-belt volcanism.

The northern belts are thought to have been originally contiguous with the Birimmian belts of west Africa. Mature sedimentary rocks overlying the greenstone belts have much in common with the Tarkwaian of West Africa.

EASTERN BELTS - Belts of the east-central craton (30,31) have not been adequately dated. Most lithostratigraphic sections have not yet been resolved, in part due to intense deformation and common medium grade metamorphism. Prominent banded iron formations, ultramafic schists, and current-bedded, fuchsite-bearing quartz arenites and conglomerates are present: these lithologies are uncommon in the northern belts. Small enclaves of iron formations and chromite-bearing ultramafic rocks occur in south and central Suriname, and might correlate with the east-central belts.

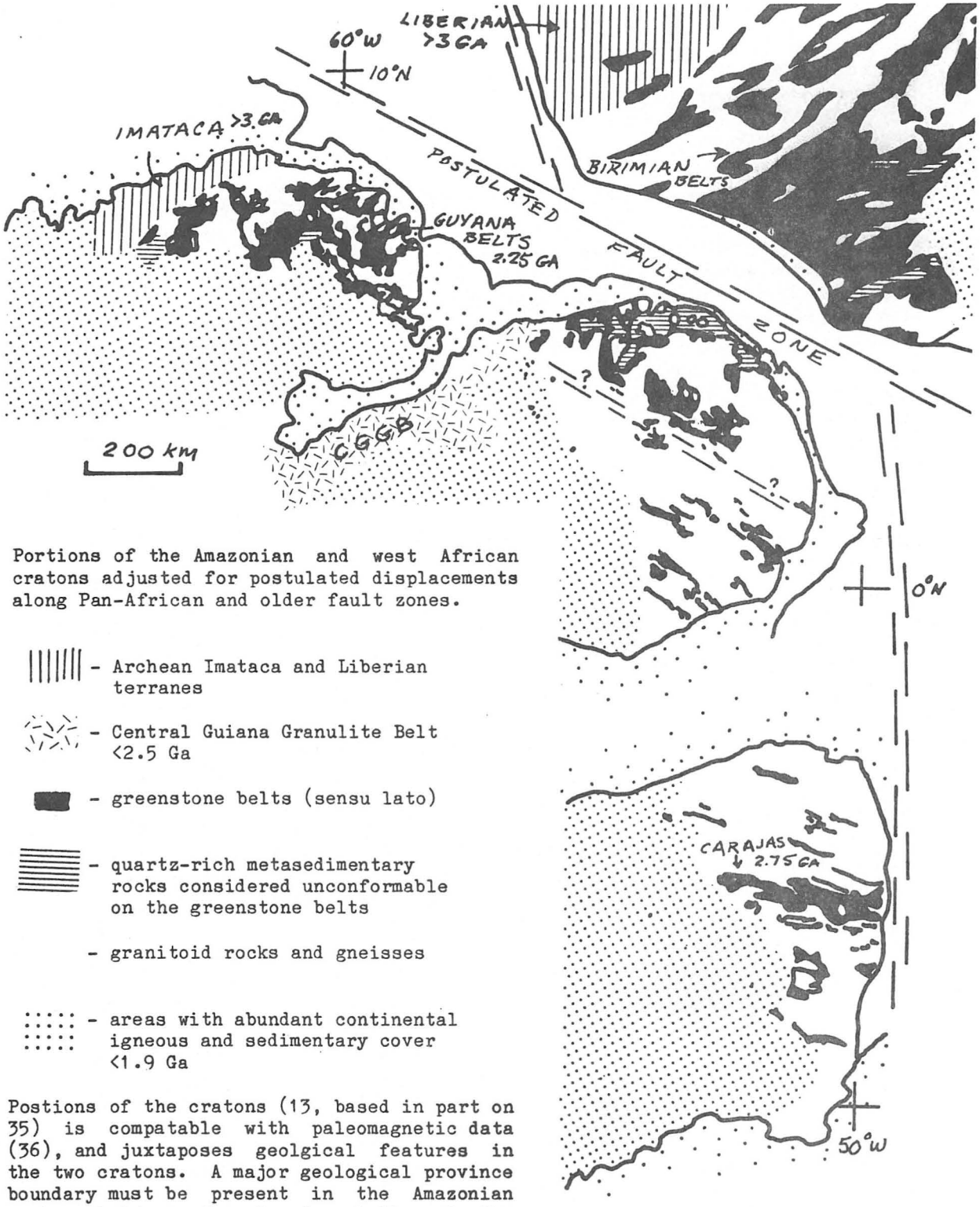
Archean greenstone belts with pillow basalts and komatiites, and belts of serpentinite occur amid granitoid rocks and gneisses in the southeastern craton, apparently forming a basement to the Serra dos Carajas belt (32). The latter has a dominantly mafic bimodal volcanic suite, roughly 4-6 km thick and dated at 2.75 Ga, overlain by 100-300 m of iron formation, and a 1-2 km thick fine clastic and chemical sedimentary complex (33,34). The mafic rocks are unlike typical Archean basalts and basaltic andesites, but have chemical and isotopic evidence of contamination with older continental crust, like many basalts of modern contiental extensional settings.

ACKNOWLEDGMENTS - Research in the Amazonian Craton has been substantially supported by the Guyana Geological Survey, the U.S. National Science Foundation (under Grants No. EAR-7617026, 8207422, and 8410379), and Utah International, Ltd. Generous institutional cooperation of DOCEGEO, the United Nations Revolving Fund for Mineral Exploration, COGEMA; the BRGM, the Suriname Geology and Mining Service, and ICOMI are also gratefully acknowledged.

REFERENCES [1] Choubert, B. (1974) Mem. B.R.G.M., 81, 213 pp. [2] Martin Bellizzia, C. (1974) Mem. Novena Conf. Geol. Inter-Guayanas, Venez. Min. Minas Hidroc. Bol. Geol. Publ. Esp. 6, pp. 251-305. [3] Mendoza, V., (1974) Segundo Congr. Latinoamer. Geol. Caracas, V. III, pp. 2237-2270. [4] B.R.G.M., (1979) Inventaire minier du departement de la Guyane. [5] Choudhuri, A. (1980) Precamb. Res., 13: 363-374. [6] Gibbs A.K., and Barron, C.N. (1983) Episodes 6, No. 2, 7-14. [7] De Vletter, D.R. (1984), Contrib. Geol. Suriname 8: 11-30. [8] Bosma, W., Kroonenberg, S.B., van Lissa, R.V., Maas, K., and de Roever, E.W.F. (1983) Geol. Mijnb. 62:241-254. [9] Bosma, W., Kroonenberg, S.B., Van Lissa, R.V., Maas, K., and de Roever, E.W.F. (1984) Cont. Geol. Suriname, 8: 31-82. [10] Santos, J.O.S., (and Santos and S.O.C. Loguercio) (1984) in Schobbenhaus et al. (eds.) Geologia do Brasil, DNPM, Brasilia, p. 59-91, 93-127. [11] Amaral, G. (1984) in F.F.M. de Almeida and Y. Hasui, O Precambriano do Brasil, Edgard Blucher, Sao Paulo, p. 6-35. [12] Gibbs, A.K., and Wirth, K.R. (1985) Workshop on Early Crustal Evolution, Lunar Planet. Sci. Inst. p. 87-91. [13] Gibbs, A.K. (1985 submitted) 7th Geol. Soc. Africa Conf. Botswana, November, 1985, 21 p. [14] Gibbs, A.K., and Olszewski, W.J., Jr. (1982) Precamb. Res. 17: 199-214. [15] Klar, G. (1979) Ph.D. Thesis, Case Western Reserve Univ., 163 pp. [16] Gruau, G., Martin, H, Leveque, B., Capdevila, R. and Marot, A. (1985) Precamb. Res. 30: 63-80. [17] Teixeira, W., Ojima, S.K., and Kawashita, K. (1984) D.N.P.M., II Symp. Amazonico, Manaus, April 8-12, 1984, Anais, p. 75-86. [18] Gibbs, A.K. (1980) Ph.D. Thesis, Harvard Univ., 385 pp. [19] Gibbs, A.K., O'Day, P.A., and Renner, R. (1984) Anais II Amazonian Symp. Manaus, Brazil, April 8-15, 1984. D.N.P.M., Brazil p. 39-52. [20] Menendez, A., Benaim, N., and Espejo, A. (1974) Mem. IX Conf. Geol. Interguayanas, Ciudad Guayana, Venezuela, 1972, pp. 339-342.. [21] Benaim, N. (1974) IX Conf. Geol. Inter-Guayanas, Publ. Espec. Min. Min. Hidroc. Venezuela, no. 6, pp. 198-206. [22] Marot, A., et al. (1984) 10 Reunion Ann. Sci. Terre, Bordeaux, 1984, Soc. Geol. Fr. Edit. Paris. [23] Veenstra, E. (1983) Contrib. Geol. Suriname, 7, 134 p. [24] Renner, R. (1985) M.Sc. Diss. Cornell Univ., 252 p. [25] O'Day, P.A. (1984) M.Sc. Diss., Cornell Univ., 154 p. [26] Cannon, R.T. (1964) Geol Surv. British Guiana Bull 35, 83 p. [27] Priem, H.N.A., et al. (1978) in Zartman, R.E., (ed.), U.S. Geol. Survey Open File Report 78-701, pp. 341-343. [28] Cordani, U.G., Teixeira, W., and Basei, M.A.S. (1975) Anais 10th Inter-Guianas Geol. Conf., Belem, Brazil. Bol., no. 2, pp. 53-54 (abstr). [29] Ben Othman, D., Polve, M., and Allegre, C.J. (1984) Nature, 307: 510-515. [30] Jorge Joao, X., and Marinho, P.A. (1982) Anais I Simp. Geol. Amazonia, Belem, v. 2, pp. 207-228. [31] Jorge Joao, X., et al. (1979) Geologia, 10, Secao Geol. Basica N. 7, DNPM, Brazil. 125 p. [32] Hirata, W.K., Rigon, J.C., Kadekaru, K., Cordeiro, A.A.C., and Meireles, E.A. (1982) Anais Simp. Geol. Amazonia I, 100-110. [33] Gibbs, A.K., Wirth, K.R., Hirata, W.K., and Olszewski, Wm. J. Jr. (1985 submitted) II Simposio Geol. Amazonia, Soc. Bras. Geol. Nucl. Norte, Dec. 1-9, 1985, Belem. 17 p. [34] Wirth, K.R., Gibbs, A.K., and Olszewski, Wm. J. Jr. (1985 submitted) II Simp. Geol. Amazonia, Soc. Bras. Geol. Nucl. Norte, Dec. 1-9, 1985, Belem. 12 p. [35] Cohen, H. (1985) Unpubl. Geology B.A. diss., Cornell Univ. 50 p. [36] Onstott, T.C., Hargraves, R.B., York, D., and Hall, C. (1984) Geol. Soc. Amer. Bull 95:1045-1054. [37] Cordani, U.G., Tassinari, C.C.G., and Kawashita, K. (1984) Ciencias da Terra, 9:6-11.

Institute for the Study of the Continents, Contribution No. 38

AMAZONIAN GREENSTONE BELTS
Gibbs, A. K.



Portions of the Amazonian and west African cratons adjusted for postulated displacements along Pan-African and older fault zones.

- Archean Imataca and Liberian terranes
- Central Guiana Granulite Belt <2.5 Ga
- greenstone belts (sensu lato)
- quartz-rich metasedimentary rocks considered unconformable on the greenstone belts
- granitoid rocks and gneisses
- areas with abundant continental igneous and sedimentary cover <1.9 Ga

Postions of the cratons (13, based in part on 35) is compatable with paleomagnetic data (36), and juxtaposes geolgical features in the two cratons. A major geological province boundary must be present in the Amazonian craton between the Carajas belt and the northern belts (17,37,12): one possible position is shown.

Spatial greenstone-gneiss relationships: evidence from mafic-ultramafic xenolith distribution patterns

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ABSTRACT

The distribution patterns of mafic-ultramafic xenoliths within Archaean orthogneiss terrain furnish an essential key for the elucidation of granite-greenstone relations. A complete gradation in scale exists between synclines, large-scale outliers and outcrop-scale xenoliths of mafic and ultramafic metavolcanic rocks. Accordingly, most greenstone belts constitute "mega-xenoliths" rather than primary basin structures. Transition along strike and across strike between stratigraphically low greenstone sequences and xenolith chains demonstrate their contemporaneity, as shown for example in Fig. 1 where the relationships between the Holenarsipur greenstone belts and associated xenoliths in southern India are portrayed. Regional to mesoscopic-scale characteristics of xenolith swarms and their relations with early greenstone units are well expressed in parts of the Pilbara Block, Western Australia. Xenolith distribution patterns in dome-arcuate syncline gneiss-greenstone terrains define subsidiary gneiss domes within the batholiths. These terrains represent least deformed cratonic "islands" within an otherwise penetratively foliated deformed gneiss-greenstone crust. The oval gneiss domes are thought to have developed originally by magmatic diapirism - evidenced by intrusive relations and contact aureoles - followed by late-stage solid state uprise related to isostatic adjustments. The late vertical movements were associated with development of major shear zones along tectonized boundary zones of batholiths, where interdigitated deformed gneiss-amphibolite schist intercalations were derived by the attenuation of xenolith-rich orthogneiss. The deformation process involved interthrusting and refolding of the interleaved plutonic and supracrustal units. The exposure of high grade metamorphic sectors is related to uplift of deep seated zones of the batholiths along reactivated faulted boundaries. Transitions from granite-greenstone terrains into gneiss-granulite suites involve a decrease in the abundance of supracrustal enclaves and an increased strain rate. Whereas early greenstone sequences are invariably intruded by tonalitic/trondhjemitic/granodioritic gneisses, stratigraphically higher successions may locally overlap older gneiss terrains and their entrained xenoliths unconformably. The contiguity

of xenolith patterns suggests their derivation as relics of regional mafic-ultramafic volcanic crustal units and places limits on horizontal movements between individual crustal blocks.

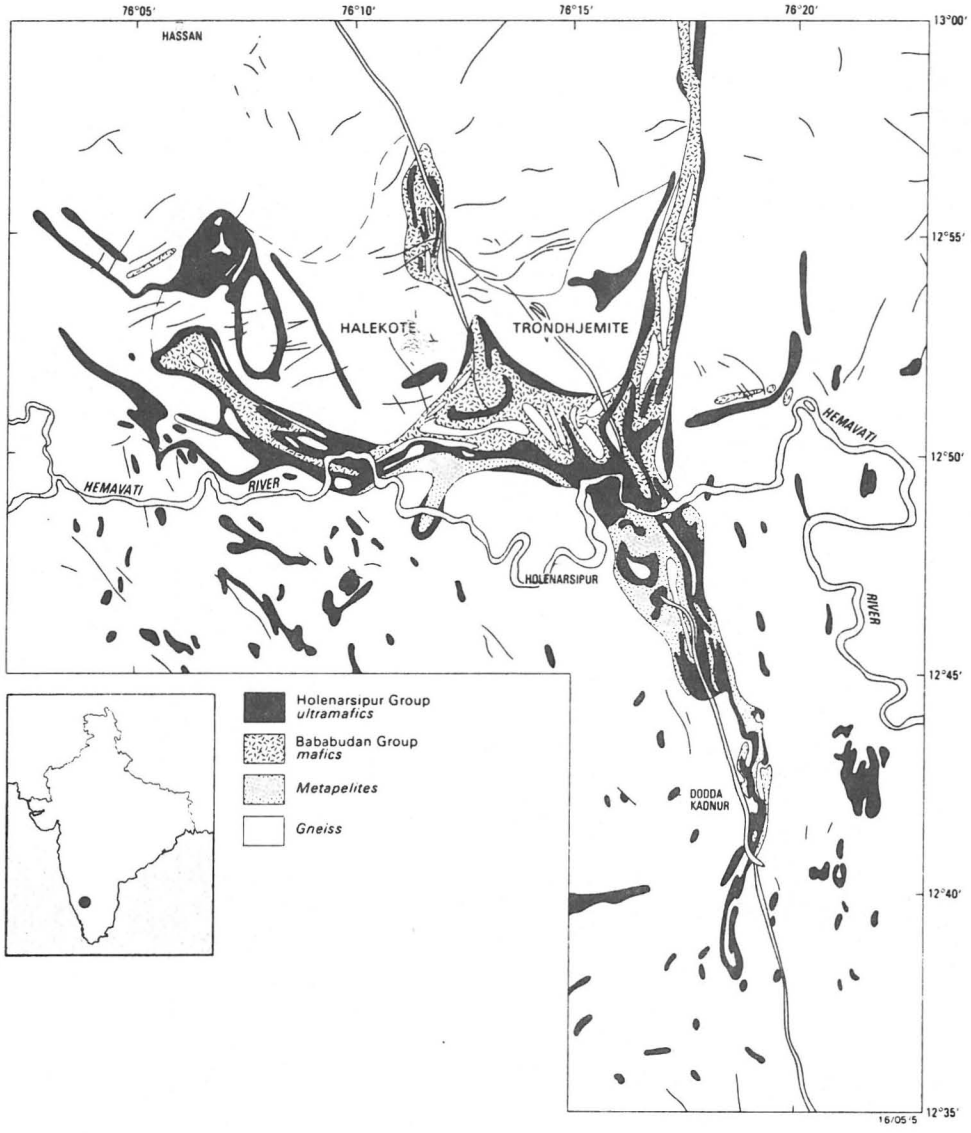


Fig. 1 - A geological sketch map of the Holenarsipur greenstone belt, Karnataka (after Naqvi, 1981, J. Geol. Soc. India, 22:458-469)

Alternative models of granite-greenstone relations are portrayed in Fig. 2. Model 1 applies to late greenstone belts overlapping sial whereas model 2 to early belts or stratigraphically basal volcanic units believed to be derived from simatic crust. Major detachments along gneiss-greenstone boundaries and local overfolding and thrusting suggest horizontal tectonic translations. These are overprinted by the dominantly vertical tectonic movements related to the diapriric (magmatic and post-magmatic) uprise of the tonalite/trondhjemite plutons. The contiguous temporal-spatial grid outlined by the xenolith swarms constrains major lateral movements of individual blocks relative to each other, placing limits on plate tectonics interpretations.

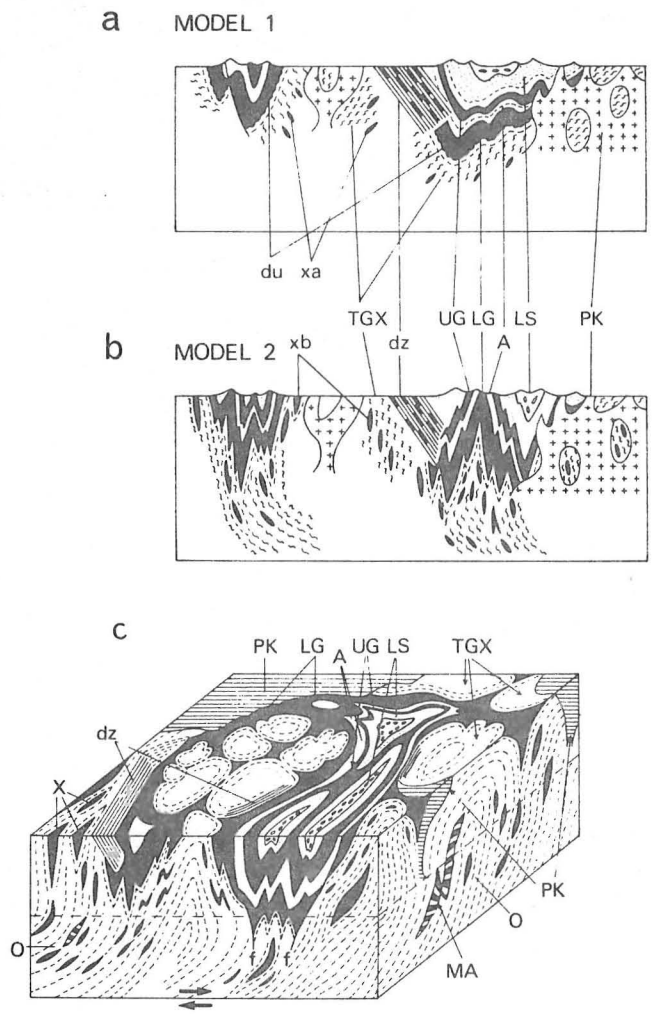


Fig. 2 - Alternative models of gneiss-greenstone relationships.
a - model 1 - gneiss-greenstone basement-cover relations, involving deformed unconformities (du).
b - model 2 - gneiss-greenstone relations involving primary and deformed intrusive contacts.
c - model 2 portrayed in block diagram, showing transition from granite-greenstone to gneiss-granulite terrain with crustal depth.
LG - lower greenstone
UG - upper greenstone
A - acid volcanics & sediments; TGX - Na-gneiss with xenoliths;
PK - late granites;
dz - deformed zone;
LS - late sediments;
O - orthogneiss;
MA - mafic and anorthositic inclusions;
x, xa, xb - xenoliths

DISMEMBERED ARCHEAN OPHIOLITE IN THE SOUTHEASTERN WIND RIVER MOUNTAINS, WYOMING -- REMAINS OF ARCHEAN OCEANIC CRUST. G.D. Harper, Dept. Geological Sciences, State University of New York, Albany, NY 12222

Archean mafic and ultramafic rocks occur in the southeastern Wind River Mountains near Atlantic City, Wyoming (Figure) and are interpreted to represent a dismembered ophiolite suite. The ophiolitic rocks occur in a thin belt intruded by the 2.6 Ga Louis Lake Batholith on the northwest (1, 2). On the southeast they are in fault contact with the Miners Delight Formation comprised primarily of metagraywackes with minor calc-alkaline volcanics.

The ophiolitic and associated metasedimentary rocks (Goldman Meadows Formation) have been multiply deformed and metamorphosed. The most prominant structures are a pronounced steeply plunging stretching lineation and steeply dipping foliation. Pillow lavas are stretched parallel to the lineation and typically have aspect ratios of 10:3:1. Bedding in banded iron formation shows polyphase folding with fold axes parallel to the stretching lineation; sheath folds are locally well developed. The intrusive contact of the Louis Lake batholith with the ophiolitic rocks has been extensively modified by deformation; the batholith becomes progessively more deformed as the contact is approached, and at the contact the batholith is strongly lineated and mylonitic. The contact between the ophiolitic rocks and the Miners Delight Formation is a major fault zone (Roundtop Fault) containing amphibolite-facies mylonites overprinted by greenschist-facies brittle cataclasites (3). These structural data indicate that the ophiolitic and associated metasedimentary rocks have been deformed by simple shear when the Miners Delight was emplaced over the Louis Lake batholith and its ophiolitic wall rocks.

The ophiolitic rocks include ultramafics, metagabbros, metadiabases, and pillow lavas. Relict structures and textures are often well preserved. However, an ophiolite "stratigraphy" is not present; the ophiolitic rocks consist of tectonic slices, from northwest to southeast, of (1) metadiabase, (2) metagabbro and ultramafics, (3) pelitic schists, quartzite, and banded iron formation (Goldman Meadows Formation), and (4) greenschist and amphibolite (Roundtop Mountain Greenstone) locally containing pillows and massive flows or sills. In addition, a thin sliver of pillow lavas occurs between the metadiabase and ultramafic rocks at one locality, but is separated from the metadiabase by a strongly foliated talc-actinolite-chlorite schist.

The ultramafic rocks are largely serpentinites, but some have amphibole-chlorite assemblages and one clinopyroxenite was found. Many of the ultramafic rocks and associated metagabbros have well-preserved relict cumulus textures, and igneous layering is visible in a few outcrops. The ultramafic rocks and associated metagabbros are only weakly deformed, in contrast to the highly deformed mafic and metasedimentary rocks.

Metadiabase occurs in a wide belt along the margin of the Louis Lake batholith, and much of it occurs as large xenoliths within the margin of the batholith. The metadiabase unit locally contains numerous parallel dikes, some of which show one-way chilling. Medium to coarse-grained

metagabbro occurs locally within the metadiabase; some of the metagabbro occurs as thin screens between fine-grained metadiabase dikes. These features suggest that the metadiabase unit represents a deformed sheeted dike complex.

The Roundtop Mountain Greenstone contains common pillow structures with well-preserved chilled rims. Massive lavas or sills comprise a significant portion of the formation, and gray phyllites occur rarely.

Rare isolated outcrops of black and foliated "basaltic komatiites," consisting primarily of actinolite and chlorite, occur in both the Roundtop Mountain Greenstone and metadiabase. However, they are chemically very different from the pillow lavas and metadiabases and possibly represent younger alkaline dikes.

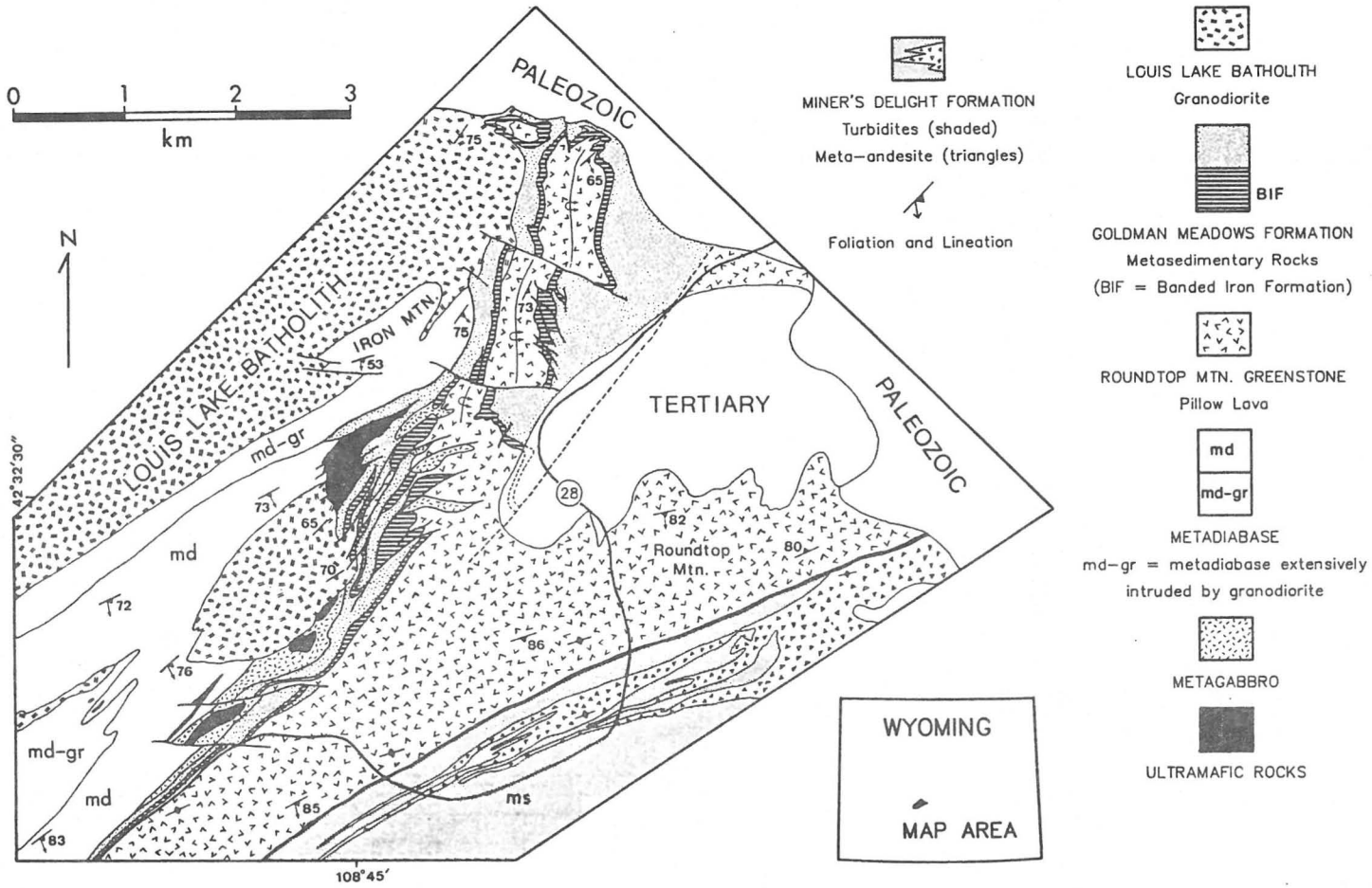
Metasedimentary rocks of the Goldman Meadows Formation overlying(?) the Roundtop Mountain Greenstone consist of pelitic schist, quartzite, and banded iron formation (1). The banded iron formation possibly formed by precipitation from hydrothermal vents in a manner similar to modern metalliferous sediments formed at spreading centers (4). Mafic sills and dikes (amphibolites) intrude the metasedimentary rocks, and are themselves deformed and metamorphosed.

Geochemical analyses were made of the metadiabase and pillow lavas to determine whether they are genetically related (5). "Immobile" trace element compositions (Ti, V, Cr, Ni, Zr, Y, Nb) are very similar in both units, consistent with the interpretation that they comprise different parts of a dismembered ophiolite. These rocks are similar to modern enriched mid-ocean ridge basalts.

The ophiolitic rocks are interpreted as the remains of Archean oceanic crust, probably formed at either a mid-ocean ridge or back-arc basin. All the units of a complete ophiolite are present except for upper mantle peridotities. The absence of upper mantle rocks may be the result of detatchment within the crust, rather than within the upper mantle, during emplacement. This could have been the result of a steeper geothermal gradient in the Archean oceanic lithosphere, or may have resulted from a thicker oceanic crust in the Archean (6).

REFERENCES

(1) Bayley, R. W., Proctor, P. D., and Condie, K. C. (1973). U.S. Geol. Surv. Prof. Paper 793, 39 pp.
(2) Stuckless, J.S., Hedge, C.E., Worl, R.G., Simmons, K.R., Nkomo, I.T., and Wenner, D.G. (1985). Geol. Soc. Am. Bull., 96, p. 850-860.
(3) Hull, J. (1984). Geol. Soc. Am. Abs. with Prog., 16, p. 600.
(4) Bonatti, E. (1981). In: Emiliani, C. (Ed.), The Oceanic Lithosphere, New York, Wiley and Sons, p. 639-686.
(5) Harper, G.D. (1985). Ofioliti (in press).
(6) Sleep, N. H., and Windley, B. F. (1982). Jour. Geol., 90, p. 363-379.


0 1 2 3
km
N
PALEOZOIC
LOUIS LAKE BATHOLITH
IRON MTN.
TERTIARY
PALEOZOIC
Roundtop Mtn.
28
md-gr
md
md-gr
md
ms
42°32'30"
108°45'
WYOMING
MAP AREA
MINER'S DELIGHT FORMATION
Turbidites (shaded)
Meta-andesite (triangles)
Foliation and Lineation
LOUIS LAKE BATHOLITH
Granodiorite
BIF
GOLDMAN MEADOWS FORMATION
Metasedimentary Rocks
(BIF = Banded Iron Formation)
ROUNDTOP MTN. GREENSTONE
Pillow Lava
md
md-gr
METADIABASE
md-gr = metadiabase extensively intruded by granodiorite
METAGABBRO
ULTRAMAFIC ROCKS

PRELIMINARY REPORT ON THE GEOLOGY AND GOLD MINERALIZATION OF THE SOUTH PASS GRANITE-GREENSTONE TERRAIN, WIND RIVER MOUNTAINS, WESTERN WYOMING (USA); W.D. Hausel, Geological Survey of Wyoming, Laramie, Wyoming 82071

The South Pass granite-greenstone terrain lies near the southern tip of the Wind River Mountains of western Wyoming. This Archean supracrustal pile has been Wyoming's most prolific source of gold and iron ore. From 1962 to 1983, more than 90 million tons of iron ore were recovered from oxide-facies banded iron formation, and an estimated 325,000 ounces of gold were mined from metagreywacke-hosted shears and associated placers (1).

Precambrian rocks at South Pass are unconformably overlain by Paleozoic sediments along the northeast flank, and a Tertiary pediment buries Archean supracrustals on the west and south. To the northwest, the supracrustals terminate against granodiorite of the Louis Lake batholith; to the east, the supracrustals terminate against granite of the Granite Mountains batholith. The Louis Lake granodiorite is approximately 2,630 $\pm$ 20 m.y. old (2), and the Granite Mountains granite averages 2,600 m.y. old (3).

The geometry of the greenstone belt is best expressed as a synform that has been modified by complex faulting and folding. Metamorphism is amphibolite grade surrounding a small island of greenschist facies rocks.

The youngest of the Archean supracrustal successions is the Miners Delight Formation. This unit yielded a Rb-Sr isochron of 2,800 m.y. (2). A sample of galena from the Snowbird Mine within the Mienrs Delight Formation yielded a model age averaging 2,750 m.y. (4). The Snowbird mineralization appears to be syngenetic and is hosted by metavolcanics of calc-alkaline affinity.

Based on regional mapping by Bayley and others (5) and by the author (in progress), four mappable supracrustal units are present. The uppermost unit, the Miners Delight Formation is greater than 1,600 m thick and consists of metagreywacke, metavolcanics, metaconglomerate, graphitic schist, and tremolite-actinolite schist. Underlying, and in fault contact with turbidites in the Miners Delight Formation, are metatholeiites of the Roundtop Mountain Formation. These metatholeiites are amphibolites, greenstones, and pillow metabasalts. The geometry of the pillows, which has been used for determining the tops and bottoms of units (5, 6) has only produced ambiguous conclusions due to the intense deformation.

The Roundtop Mountain greenstones are underlain(?) by quartzite, metapelite, and banded iron formation of the Goldman Meadows Formation. This unit, in turn, is underlain(?) by mafic and ultramafic schists tentatively named the Diamond Springs ultramafics. This ultramafic unit consists of amphibolite, serpentinite, metaperidotite, and tremolite-talc-chlorite schist. Harper (6) interprets this unit to represent a dismembered ophiolite sequence.

Mining districts occur on both limbs of the South Pass synform. While the South Pass - Atlantic City District occurs along the northwestern limb, the Lewiston District is found on the eastern limb (7). Gold mineralization in the South Pass - Atlantic City District is found chiefly in shear zones in

metagreywacke adjacent to metagabbro sills and dikes. Wall-rock studies of the auriferous shears, show Si and K have been enriched and Ca and Mg have been leached. Mineralogically, these chemical changes are expressed as weak phyllic alteration of the wall rock. Analyses for native gold from the Diana Mine show high Au/Ag and low Au/Cu ratios (8). The gold analyses and wall-rock alteration are characteristic of a hypothermal vein.

The Lewiston District on the eastern flank of the synform includes strike-trending, metagreywacke - hosted, auriferous shears along the limb of a major fold (9). A few major lodes are localized where the strike shears intersect cross-cutting shears. Wall rocks show distinct chloritic and hematitic alteration as well as weak phyllic alteration.

(1) Hausel, W.D., 1980, Gold districts of Wyoming: Geological Survey of Wyoming Report of Investigations 23, 71 p.

(2) Stuckless, J.S., Hedge, C.E., Worl, R.G., Simmons, K.R., Nkomo, I.T., and Wenner, D.B., 1985, Isotopic studies of the late Archean plutonic rocks of the Wind River Range, Wyoming: Geological Society of America Bulletin, v. 96, p. 850-860.

(3) Stuckless, J.S., and Peterman, Z.E., 1977, A summary of the geology, geochronology, and geochemistry of Archean rocks of the Granite Mountains, Wyoming: Wyoming Geological Association Earth Science Bulletin, v. 10, no. 3, p. 3-10.

(4) Cannon, R.S., Jr., Bayley, R.W., Stern, T.W., and Pierce, A.P., 1966, Ancient rocks and ores in south-central Wyoming [abs.]: Geological Society of America Rocky Mountain Section 18th Annual Meeting Program, p. 27.

(5) Bayley, R.W., Proctor, P.D., and Condie, K.C., 1973, Geology of the South Pass area, Fremont County, Wyoming: U.S. Geological Survey Professional Paper 793, 39 p.

(6) Harper, G.D., in press, Dismembered Archean ophiolite, Wind River Mountains, Wyoming (USA), *in* Desmons, J., *ed.*, Ophiolites through time: Ophioliti.

(7) Harris, R.E., Hausel, W.D., and Meyer, J.E., 1985, Metallic and industrial minerals map of Wyoming: Geological Survey of Wyoming Map Series MS-14, scale 1:500,000.

(8) Antweiler, J.C., and Campbell, W.L., 1977, Application of gold compositional analyses to mineral exploration in the United States: Journal of Geochemical Exploration v. 8, p. 17-29.

(9) Hausel, W.D., 1984, Preliminary geologic map of the Lewiston gold mining district (Radium Springs Quadrangle), South Pass, Wyoming: Geological Survey of Wyoming unpublished mineral report # MR 84-7, scale 1:24,000.

THE KOLAR SCHIST BELT: A POSSIBLE ARCHEAN SUTURE ZONE

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The Kolar Schist Belt in the Karnataka craton, south India, is a 4 to 20 km by 80 km long, N-S trending Archean supracrustal belt dominated by mafic metavolcanics. The schist belt is surrounded on both sides by granodioritic gneisses collectively known as the Peninsular Gneiss. Our work has shown that the Kolar Schist Belt and the surrounding gneisses include major discontinuities in age, structural style, and composition. These discontinuities are defined by the schist belt itself.

The results reported here are based on our Rb-Sr, Sm-Nd, and Pb-Pb whole rock isotope data; U-Pb dating of zircon and sphene; major and trace element (including REE) analyses; and field observations.

The schist belt is broadly synformal, but is complexly refolded into basin and dome structures (D. Mukhopadyay, personal communication). The first period involved N-S trending isoclinal recumbant folds during E-W compression. These folds were refolded into tight, upright folds along E-W trending axes. This sequence is broadly similar to those seen in other schist belts in the western part of the Karnataka craton.

Contacts between the Peninsular Gneiss and the margins of the belt have long been thought to represent an erosional unconformity. However, our recent field work indicates that the rocks at the contacts are physically interleaved by left lateral shearing. Due to this shearing the adjoining gneisses have been converted to quartz-muscovite schists, which were previously interpreted to be metasedimentary rocks.

The gneisses east of the schist belt are relatively homogeneous, granodioritic gneisses which were folded prior to intrusion of minor felsic bodies. Folds have not yet been defined in these gneisses, but a strong foliation was developed which strikes NNE and dips steeply to the west, suggesting horizontal compression.

The gneisses west of the schist belt show a much more complex, earlier history than that of the eastern gneisses. The granodioritic Dod Gneiss is the earliest unit on the western side of the schist belt. This rock was subjected to a period of deformation shown by an early foliation seen in some less-strained exposures. Subsequently, the Dod Gneiss was intruded by the leucocratic, granodioritic Dosa Gneiss and the granodioritic Patna Granite.

Following the intrusion of the Dosa Gneiss, the terrane to the west of the schist belt was subjected to a period of horizontal compression producing tight to isoclinal, W overturned folds with gently N or S plunging axes. The strong NNE axial planar foliation produced by this deformation is cut by the later N-S shears along the western margin of the schist belt.

The gneisses on the east and west side of the belt have been dated using U-Pb ages for small populations of abraded zircons and abraded single zircons as well as sphene. These zircons commonly give concordant ages, in which case the small populations of zircons (ca. 100 micrograms) have analytical uncertainties of less than 1 Ma. and the single zircons have uncertainties of about 5 Ma.

Gneisses east of the belt were intruded at 2529±1 Ma based on U-Pb ages for zircon. This age is consistent with the Rb-Sr and Pb/Pb whole rock isochron ages. The isochrons have a mantle-like initial ratio for Sr (87/86=0.7013) and mu=8 for the Pb data. These values suggest that the gneisses were not derived from a much older continental crust. U-Pb ages for metamorphic sphene are 2520 ±1 Ma suggesting that the gneisses were metamorphosed to at least amphibolite grade at that time.

West of the belt, based on U-Pb ages for zircon, the Dod Gneiss was emplaced at 2610±5 Ma, the Dosa Gneiss was intruded at 2550±10 Ma and the Patna Granite at 2551±1 Ma. The time of metamorphism based on the U-Pb ages for sphene from the Dod Gneiss is 2551±1 Ma. Rb-Sr and Pb/Pb whole rock data suggest that the gneisses were variably contaminated by an older basement. U-Pb ages for some of the single zircon cores from the Dod Gneiss and later aplitic dikes indicate a zircon component was inherited from this basement, which has a minimum age of 3200 Ma. The basement, which has not yet been clearly identified in the field, seems to include quite evolved felsic rocks.

In the Kolar Schist Belt there are two suites of komatiitic and tholeiitic amphibolites. Both the komatiitic and tholeiitic amphibolites on the eastern side are light REE enriched, and almost all of the komatiitic and tholeiitic amphibolites in the west-central part of the belt are lightest REE depleted. The preservation of rare pillow structures and the association of the amphibolites with iron formation suggest that the amphibolites were formed under submarine conditions. The grade of metamorphism is amphibolite facies.

Rajamani et al. (1) concluded that the komatiitic amphibolites from both the east and west central part of the belt were derived by 10 to 25% melting at depths greater than 80 km and at temperatures greater than 1500°C in a mantle with an FeO/MgO ratio greater than that of pyrolite. Other models proposed for the generation of komatiites generally require larger percentages of melting to generate the high MgO abundances.

Rajamani et al. (1 and in preparation) suggest that the tholeiites appear to have been derived by melting at shallower levels than the komatiites and derived from sources which were highly variable in their FeO/MgO ratios, generally with FeO/MgO ratios much greater than that for the sources for the komatiites. The key arguments are that: the tholeiites are very iron-enriched

compared to the field for potential melts of pyrolite at pressures less than 25 kb on an olivine saturation surface; and while the incompatible elements show similar ratios in the komatiites and tholeiites for each suite, the expected correlations between major and trace elements for differentiation from komatiites or melting of sources similar to those of komatiites are not found.

Sm-Nd data for komatiites from both sides of the belt lie with large variations about a 2900 Ma isochron. It is not clear why the data lie about a 2900 Ma isochron. Is this the age of these amphibolites? If this is so, they are much older than the igneous felsic rocks on either side of the belt which are 2500 to 2600 Ma. Or, is this the time when the sources became variably light REE enriched and depleted? Some of the variation in the Sm/Nd ratios is clearly a function of melting processes in which garnet was left in the residue. Perhaps the variability in the data about the reference line reflects a number of reasons such as: variable times of light REE depletion and enrichment of their mantle sources; as well as the possible effects of crustal contamination or metamorphic alteration.

Even though the ages of the units making up the Kolar Schist Belt are poorly constrained, the sources of the amphibolites so far analyzed had long-term histories of LREE depletion (epsilon Nd of +2 to +8 for an age of 2900 m.y.) relative to other Archean mafic rocks which commonly have epsilon Nd equal to about +2.0 $\pm$ 2.0.

The Kolar Schist Belt represents a N-S trending discontinuity in the structures, lithologies, and emplacement and metamorphic ages of late Archean gneisses. The suggestion of a much older basement on the west side of the belt is not seen on the east. Within the schist belt amphibolites from each side have distinctly different chemical characteristics, suggesting different sources at similar mantle depths. These amphibolites were probably not part of a single volcanic sequence, but may have formed about the same time in two completely different settings. Could the amphibolites with depleted light REE patterns represent Archean ocean floor volcanics which are derived from a mantle source with a long term depletion of the light REE? Why are the amphibolites giving an age which may be older than the exposed gneisses immediately on either side of the belt? These results suggest that it is necessary to seriously consider whether the Kolar Schist Belt may be a suture between two late Archean continental terranes.

REFERENCES

(1) Rajamani V., Shivkumar K., Hanson G. N., and Shirey S. B. (1985) Geochemistry and petrogenesis of amphibolites, Kolar Schist Belt, South India: Evidence for komatiitic magma derived by low percentages of melting of the mantle J. Petrol. 26, p. 92-123.

EVIDENCE FOR SPREADING IN THE LOWER KAM GROUP OF THE YELLOWKNIFE GREENSTONE BELT: IMPLICATIONS FOR ARCHEAN BASIN EVOLUTION IN THE SLAVE PROVINCE. H. Helmstaedt and W.A. Padgham, Dept. of Geological Sciences, Queen's University, Kingston, Canada K7L 3N6, and Geology Division, Northern Affairs Program, P.O. Box 1500, Yellowknife, N.W.T., Canada X1A 2R3

The Yellowknife greenstone belt is located in the southwestern part of the Slave Structural Province, a Late Archean (2.7-2.5 Ga) granite-greenstone terrane in the northwestern part of the Canadian Shield. Supracrustal rocks within this province, collectively referred to as Yellowknife Supergroup (Henderson, 1970), differ from the supracrustal successions of the Superior Province and other older Archean terranes by the absence of komatiites and the high proportion of metasedimentary to metavolcanic rocks. The Yellowknife belt was first mapped by Jolliffe (1942, 1946) on the scale of one inch to one mile, and the gold-producing area around Yellowknife was remapped on a more detailed scale (1:12 000) by Henderson and Brown (1966). As the belt became the best-known example of the basalt-dominated supracrustal belts in the western Slave Province (Padgham, 1985), the stratigraphic framework established here (Henderson, 1970), formed the basis for the development of models for Archean basin evolution (McGlynn and Henderson, 1972; Henderson, 1981). Under a recent mapping program of the Geology Division of the Northern Affairs Department in Yellowknife, detailed mapping was extended, and a 1:10 000 map series for the entire belt is currently under preparation. This work resulted in a number of revisions and refinements in the established stratigraphy (Helmstaedt and Padgham, 1986) and provides the basis for a reassessment of current models of greenstone belt evolution in the Slave Province.

The major portion of the Yellowknife greenstone belt is underlain by the predominantly mafic rocks of the Kam Group which consists of a northeasterly-striking, homoclinal sequence of flows and tuffs that dip steeply and face uniformly to the southeast (Fig. 1). Numerous dikes, sills and irregular bodies of gabbro and locally anorthosite appear to form an integral part of the volcanic sequence. The Kam Group has been subdivided into four formations (Fig. 2) with a combined thickness of approximately 11km. The lower contact is obscured by the intrusion of a composite batholith (Western Granodiorite, Fig. 1) that cuts across the strike of the flows. At the base of the exposed section, near the northern end of the belt, a narrow band of felsic volcanic rocks and banded iron- formation is in conformable contact with overlying pillowed flows above which a mafic extrusive- intrusive complex is developed (Fig. 2) whose pseudostratigraphy resembles that of certain Phanerozoic ophiolites. Near the southwestern end of the belt, the upper part of the Kam Group (Yellowknife Bay Formation) overlaps a sequence of older volcanic and sedimentary rocks belonging to the Octopus Formation (Fig. 1). In the northern part of the belt, the upper formations of the Kam Group are truncated by an unconformity beneath conglomerates and sandstones of the Jackson Lake Formation. Farther to the south, where the top of the Kam is preserved locally, it is overlain by calc-alkaline rocks of the Banting Group that, in turn, are overlain by turbidites of the Walsh and Burwash Formations. All rocks of the Yellowknife Supergroup are deformed and metamorphosed, with metamorphic grade increasing from greenschist to amphibolite facies towards the granitoid intrusions. Inspite of the metamorphic overprint, however, primary structures and intrusive relationships are well preserved.

The mafic intrusive-extrusive complex of the Chan Formation (Fig. 2) grades from a lower part, dominated by gabbro, through a multiple dike complex into massive and pillowed flows with thin beds of interflow sediments. At the base of the section is a sheet-like body of massive, medium- to coarse-grained, locally layered gabbro that was intruded into a sequence of pillowed flows, remnants of which are preserved at three levels. The upper boundary of this body is a relatively sharp transition into the dike complex which consists of numerous, fine- to medium-grained metadiabase dikes and septa and irregular bodies of relatively coarse gabbro between which screens of pillowed flows can be recognized. The dikes, which are locally sheeted, show symmetric and asymmetric chilled margins and range in width from less than one to over 10m. Some dikes grade into pillows, suggesting that they were intruded close to the seafloor and may have acted as feeder system to the growing volcanic pile (de Wit and Stern, 1978). Most of the irregular gabbros are multiple intrusions with abundant chilled margins and extremely complex contact relationships. Igneous layering is generally absent at this level, but an up to 100m thick, sheet-like body of gabbroic anorthosite was recognized (Fig. 2). It is surrounded entirely by gabbro that has chilled margins against the anorthosite. Though massive and pillowed flows predominate above the dike complex, sills and irregular bodies of gabbro, many of them multiple intrusions, are common in the upper parts of the Chan Formation. The top half of the Kam Group continues to be dominated by pillowed and massive mafic flows, but contains numerous intercalations of felsic tuffs and tuffaceous sediments. Some of the flows and many of the interflow tuffs and sediments are continuous along strike for more than 10 km and allow stratigraphic correlation across Proterozoic transcurrent faults (Fig. 1). Synvolcanic mafic intrusions in this part of the section consist of numerous sills some of which are connected to dike swarms. The entire section was intruded also by several post-volcanic dike swarms.

The Yellowknife greenstone belt has been interpreted as the western margin of an Archean turbidite-filled basin bordered in the east by the Cameron River and Beaulieu River volcanic belts (Henderson, 1981; Lambert, 1982). This model implies that rifting was entirely ensialic and did not proceed beyond the graben stage. Volcanism is assumed to have been restricted to the boundary faults, and the basin was floored by a down-faulted granitic basement. On the other hand, the enormous thickness of submarine volcanic rocks and the presence of a spreading complex at the base of the Kam Group suggest that volcanic rocks were much more widespread than indicated by their present distribution. Rather than resembling volcanic sequences in intracratonic graben structures, the Kam Group and its tectonic setting within the Yellowknife greenstone belt have greater affinities to the Rocas Verdes of southern Chile (deWit and Stern, 1981), Mesozoic ophiolites, that were formed in an arc-related marginal basin setting. The similarities of these ophiolites with some Archean volcanic sequences was previously recognized by Tarney et al. (1976) and served as basis for their marginal-basin model of greenstone belts. The discovery of a multiple and sheeted dike complex in the Kam Group confirms that features typical of Phanerozoic ophiolites are indeed preserved in some greenstone belts and provides further field evidence in support of such a model.

FIGURE CAPTIONS: (1). Geological map of the Yellowknife greenstone belt. Modified from published maps of the Geological Survey of Canada and Northern Affairs Program, Yellowknife. (2). Generalized section of the Kam Group.

REFERENCES:

de Wit, M.J. and Stern, C.R. (1978) Jour. Volcanology and Geothermal Res., 4, p. 55-80.

de Wit, M.J. and Stern, C.R. (1981) Tectonophysics, 72, p. 229-260.

Helmstaedt, H. and Padgham, W.A. (1986) Canadian Jour. Earth Sciences (in press).

Henderson, J.B. (1970) Geol. Surv. Canada, Paper 70-26.

Henderson, J.B. (1981) in Kroner, A., ed., Plate Tectonics in the Precambrian. Elsevier, Amsterdam, p. 213-235.

Henderson, J.F. and Brown, I.C. (1966) Geol. Surv. Canada, Bull. 141.

Jolliffe, A.W. (1942) Geol. Surv. Canada, Map 868A.

Jolliffe, A.W. (1946) Geol. Surv. Canada, Map 709A.

Lambert, M.B. (1982) Geol. Surv. Canada, Paper 82-1A, p. 165-167.

McGlynn, J.C. and Henderson, J.B. (1972) in Geol. Assoc. Canada, Special Paper 11, p. 506-526.

Padgham, W.A. (1985) in Geol. Assoc. Canada, Special Paper 28, p. 133-151.

Tarney, J., Dalziel, I.W.D., and de Wit, M.J. (1976) in Windley, B.F., ed., The Early History of the Earth. Wiley, London, p. 131-146.

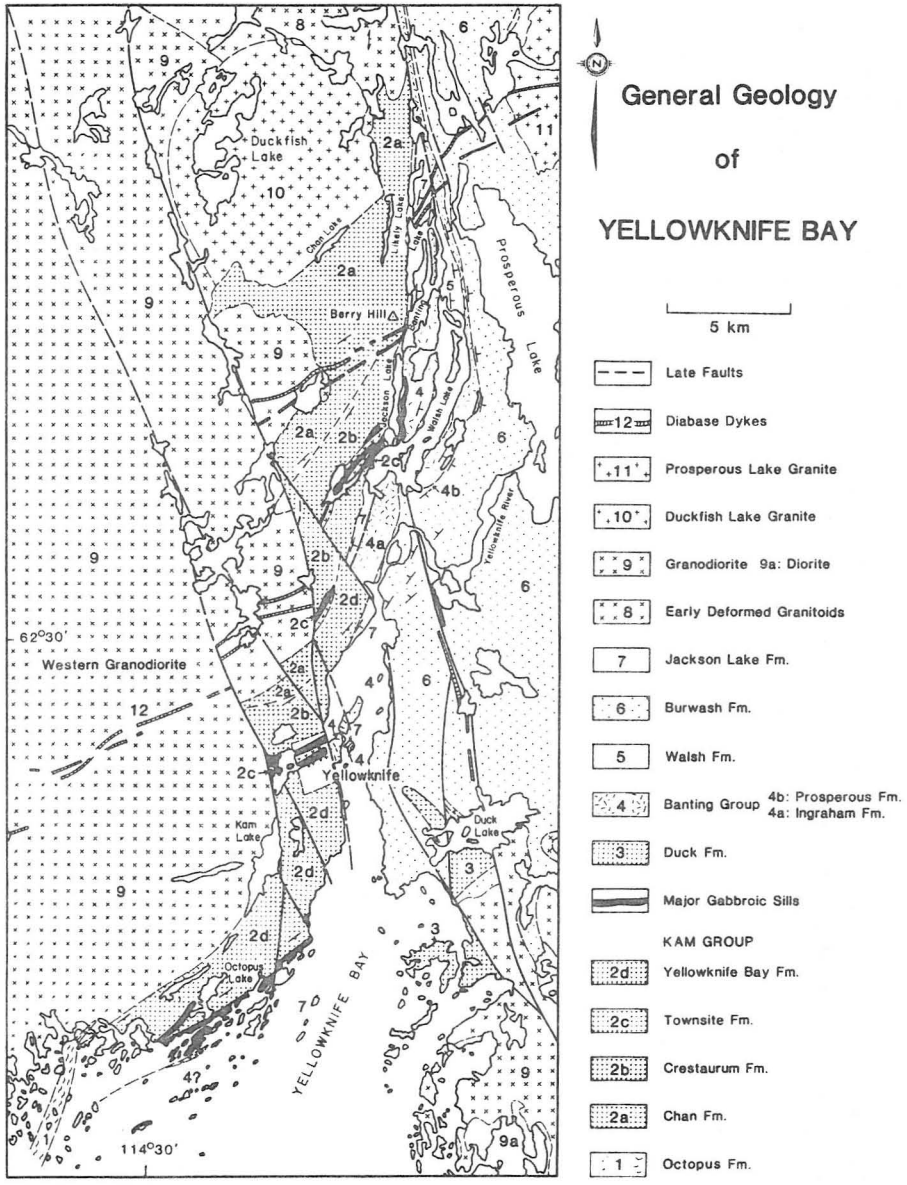

FIGURE 1

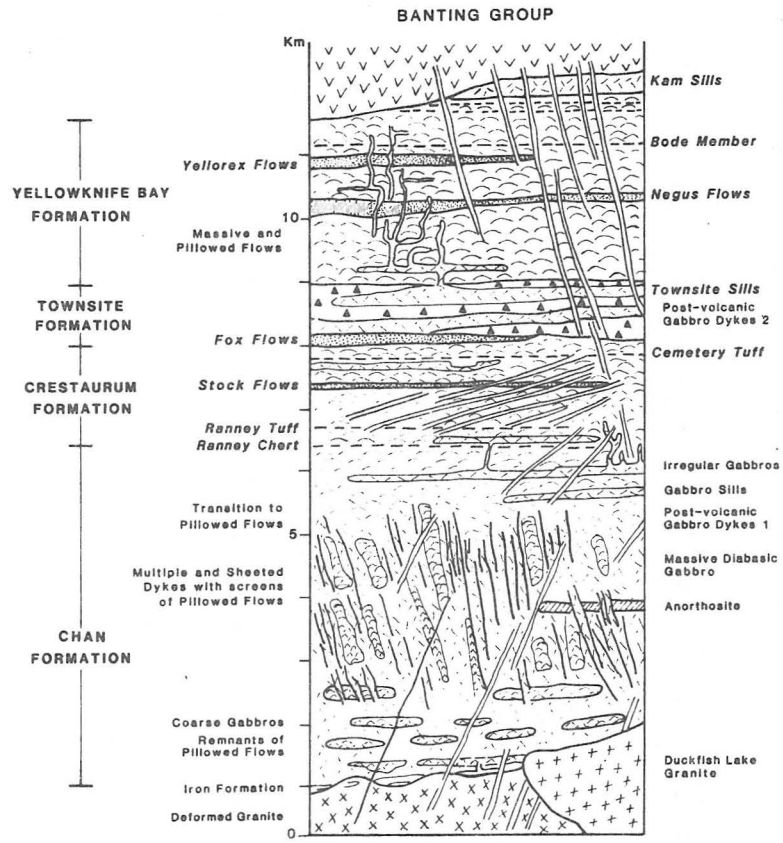

FIGURE 2

ARCHAEAN WRENCH-FAULT TECTONICS IN THE ABITIBI GREENSTONE BELT OF CANADA; C. Hubert and J.N. Ludden, Department of Geology, Universite de Montreal, Montreal, Canada, H3C 3J7.

Structural studies in the southern sector of the Abitibi greenstone belt of Canada have defined a deformation style associated with a wrench-fault system (1). The fundamental features of this tectonic regime are the following:

i) the formation of lozenge - shaped blocks of terrane which are bounded either by fault - zones or by highly strained zones of ductile deformation. In these blocks there is a pronounced gradient in degree of deformation from well preserved cores to highly deformed and sometimes mylonitized margins;

ii) sedimentary accumulations occur along the margins of the blocks in a series of narrow basins bounded by shear - zones;

iii) blocks of different lithologies and structural and metamorphic histories have been juxtaposed;

The deformation history is summarized below and shown in simplified form in Figure 1. The first deformation phase was simple shearing associated with WSW - ESE sinistral wrench faulting which resulted in NW - SE fold traces and transected schitosities. Progressive deformation affected blocks of terrane in a tectonic regime in which volcanism, shearing, deformation and uplift and erosion were synchronous; terranes composed dominantly of felsic volcanics were juxtaposed with blocks of ultramafic volcanic and sedimentary accumulations.

The first deformation phase was followed by N - S compression resulting in the development of major E - W thrust-shears. This deformation resulted in the formation of an E - W fold trace and crenulation clevage. The superposition of the two deformation episodes resulted in the generation of NE - SW and NW - SE complementary faults defining " S " and " Z " sigmoidal forms in highly strained E - W shear - zones (Figure 2).

U - Pb zircon ages, compiled in Ludden et al., (2) indicate that the volcanic accumulations in the Porcupine, Rouyn-Noranda and Val D' Or areas of the southern Abitibi belt define an axis of volcanism of tholeiitic lineage that was at its peak at approximately 2700 m.y.. These volcanic rocks superimpose an older volcano-plutonic terrane which is characterized in the NE - Abitibi belt and can be correlated towards the SW across the Kapuskasing front to the Wawa subprovince (2,3). This axis of volcanism is approximately 2850 - 2720 m.y. in age and is dominated by calcalkaline volcanic and plutonic rocks.

ARCHAEAN WRENCH-FAULT TECTONICS

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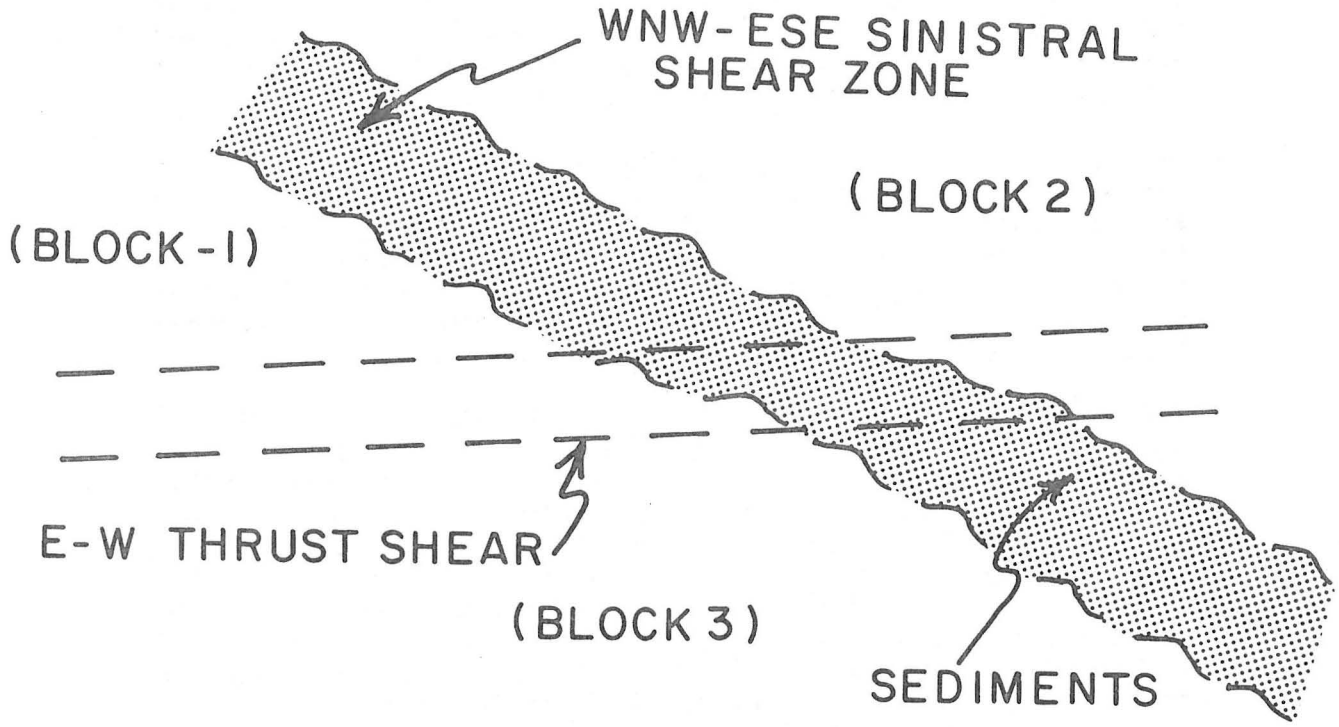
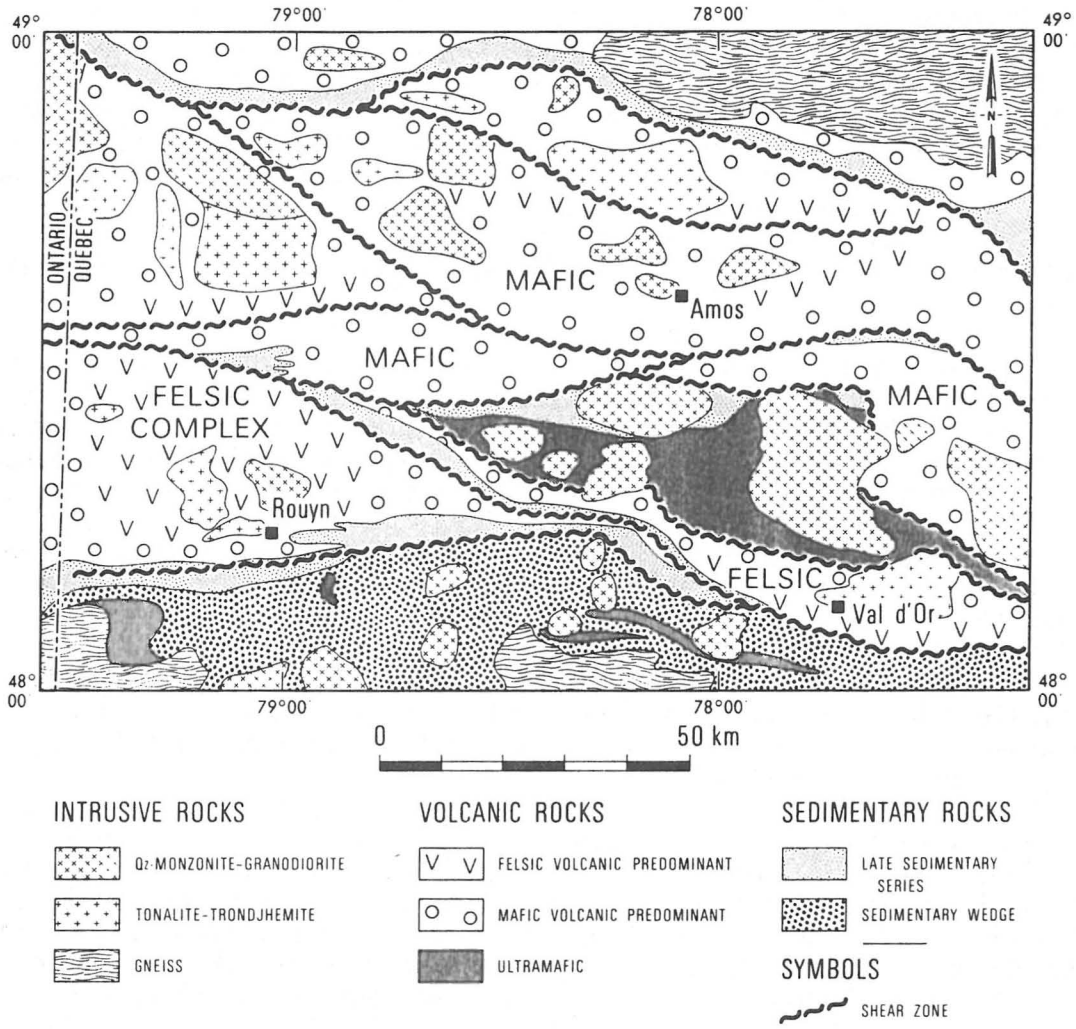
A tectonic model is proposed in which the southern Abitibi belt formed in a series of rift basins which dissected an earlier formed volcanic arc. Comparisons can be made with Phanerozoic areas such as, the Hokuroko basin of Japan, the Taupo volcanic zone of New Zealand and the Sumatra and Nicaragua volcanic arcs. In addition the identification of the major E - W thrust shears make it possible to speculate that the southern Abitibi belt comprises a collage of blocks of terrane which have been accreted against a more stable continental margin or micro-continent. If this interpretation is correct analogies can be made with the SW margin of the U.S.A. in which recently formed blocks of volcanic terrane are being accreted against the western margin of the U.S.A..

FIGURE 1 : Deformation History of the Southern Abitibi Belt.



ARCHAEAN WRENCH-FAULT TECTONICS
C. Hubert and J.N. Ludden

FIGURE 2 : Schematic representation of " lozenge - shaped " blocks of terrane bounded by shear-zones and thrust-shears in the Southern Abitibi belt



REFERENCES:

(1) Hubert C., Gelinas L., Trudel P., 1984, Can J. Earth Sci., 22, 240-255.
(2) Ludden J.N., Hubert, C., Gariepy, C., 1986 , Geol. Mag., In Press.
(3) Percival J.A. and Krogh T.E., 1983, Can J. Earth Sci., 20, 830-843.

TRANSPRESSION AS THE MAIN DEFORMATIONAL EVENT IN AN ARCHEAN GREENSTONE BELT, NORTHEASTERN MINNESOTA; P.J. Hudleston and D. Schultz-Ela, Department of Geology and Geophysics, University of Minnesota, Minneapolis, MN 55455; R.L. Bauer, Department of Geology, University of Missouri, Columbia, MO 65211; D.L. Southwick, Minnesota Geological Survey, 2642 University Ave., St. Paul, MN 55114.

Deformed and metamorphosed sedimentary and volcanic rocks of the Vermilion district constitute an Archean greenstone belt trending east-west between higher grade rocks of the Vermilion Granitic Complex to the north and the Giants Range batholith to the south. Metamorphic grade is low throughout, being lowest in the center of the belt (chlorite zone of the greenschist facies) (1). All the measured strain, a cleavage or schistosity, and a mineral lineation in this belt are attributed to the 'main' phase of deformation (D_2) (2) that followed an earlier nappe-forming event (D_1)(3, 4), which left little evidence of penetrative fabric (2).

Previous work assumed that the D_2 deformation resulted from north-south compression across the district, presumably related to diapiric intrusion of the batholithic bodies to the north and south (1). A number of lines of evidence now lead us to believe that a significant component of this deformation resulted from dextral shear across the whole region. Thus the Vermilion fault, a late-stage largely strike-slip structure (1) that bounds the Vermilion district to the north, may simply be the latest, most brittle expression of a shear regime that was much more widespread in space and time. Features that are indicative of shear include ductile shear zones with sigmoidal foliation patterns, highly schistose zones with the development of shear bands, feldspar clasts or pyrite cubes with asymmetric pressure shadows, and the fact that the asymmetry of the F_2 folds is predominantly Z for at least 15 km south of the Vermilion fault.

The presence of a large component of simple shear may help explain additional structural features in a simpler way than otherwise possible. Just south of the Vermilion fault the cleavage locally becomes folded and a new spaced cleavage develops in a similar orientation to the old cleavage away from the folds. Rather than interpreting this as evidence for an additional episode of deformation, we consider it to be due to a single process of continuous shear: a foliation develops and after a large strain local perturbations result in folding of the old foliation and the development of a new one axial planar to the folds.

The same type of perturbation can lead to the juxtaposition of ENE-trending zones of constrictional and flattening strains (5), a distinctive feature of the rocks of the Vermilion district otherwise hard to account for. The maximum extension directions (X) of all samples showing constrictional strain, plunge east at angles between 30 and 65°. X in samples showing flattening strain plunges east or west, but near the Vermilion fault all plunges are west or more steeply east than they are in constrictional samples. The maximum shortening direction (Z) plunges consistently less than 25° to the north or south.

The strain variations require a model which can satisfy compatibility constraints and space considerations. The area of consistent constrictional strains in the south may represent one regional component of the strain. Spatial correspondence of flattening strains with the Vermilion fault suggests that a simple shear component was added in that area. A modified model of transpression may explain how E-plunging X axes are reoriented to become W-plunging by a concomitant inhomogeneous progressive simple shear. Less than vertical plunge of the X axes may necessitate some component of oblique motion on the fault.

In a general way the strain patterns observed in the Vermilion district can be reasonably explained by a history of N-S shortening accompanied by inhomogeneous dextral simple shear. The variations of strain may be a consequence of variations in the relative intensities of shortening and shear, large perturbations of the shear, or the influences of other structures. There may be an analogy with the strain partitioning that occurs in small scale ductile shear zones at large strains.

For transpression to have occurred, the Vermilion district would have to have been a region of relatively soft crust caught beween two more rigid (either thicker or cooler) blocks to north and south. We do not yet know to what extent the high-grade terranes to north and south were also affected by transpression deformation and therefore the configuration of the more rigid block.

References

(1) Sims, P.K. (1976) Bull. Geol. Soc. Am., 87, p. 379-389.

(2) Hudleston, P.J. (1976) Can. J. Earth Sci., 13, p. 579-592.

(3) Bauer, R.L. (1985) Geology, 13, p. 657-660.

(4) Bauer, R.L., Hudleston P.J. and Southwick d.L. (1986) In Workshop on the Tectonic Evolution of Greenstone Belts. Lunar and Planetary Institute, Houston (this volume).

(5) Ela, D. and Hudleston P.J. (1985) In 31st Institute on Lake Superior Geology (abstract).

A CONTINUOUS RECORD OF TECTONIC EVOLUTION FROM 3.5 Ga TO 2.6 Ga IN SWAZILAND AND NORTHERN NATAL

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The ∿ 3.5 Ga-old bimodal suite underlying an extensive area in south-western Swaziland comprises the oldest-dated sialic rocks in the Kaapvaal structural province(1). The suite consists of leucocratic, layered tonalitic-trondhjemitic gneisses and amphibolites characterized by the effects of repeated high strains(2). This suite is considered to represent a sialic basement on which metavolcanic and metasedimentary rocks, now preserved as scattered 'greenstone' remnants, accumulated. Direct evidence to confirm this temporal relationship is lacking, but structural data from the Dwalile, Assegaai and Commondale areas indicate that (i) the bimodal gneisses experienced a complex structural history prior to the first recognizable deformation in the supracrustal rocks (i.e. D_1 in the supracrustals is equivalent to Dn + 1 in the gneisses) and (ii) scattered remnants of the Dwalile rocks infolded with the bimodal suite structurally overlie the gneisses and are preserved in synformal keels (2)(3). Significant proportions of metaquartzites and metapelites are present in the Assegaai 'greenstone' sequence, the presence of which implies the existence of felsic crust in the source area from which these sediments were derived, a conclusion that is consistent with the structural data.

Ultramafic and pillowed mafic rocks of komatiitic and tholeiitic affinity are present in all four 'greenstone' remnants, but each contains distinctive lithologies. The Assegaai sequence is characterized by the abundance of clastic and chemical sediments that are a minor component of the Commondale and Nondweni remnants. In the former there is a prominent sub-volcanic intrusion composed of multiple layers of massive serpentinite (in which relict cumulate olivine is present locally) alternating with spinifex-textured (olivine and pyroxene) layers. There is a consistent relationship in the thicknesses of the individual layers, i.e. where the serpentinite layers range from 10 to 40 m in thickness the spinifex-textured layers are 1 to 3 m thick. At Nondweni the sequence is dominated by pillowed tholeiites interlayered with high-magnesium basalts and basaltic komatiites (up to 22% MgO). The latter show well developed pyroxene spinifex but peridotite komatiites and units with olivine spinifex are entirely absent. Silicification of the volcanics considered to be contemporaneous with extrusion is not uncommon. Within the volcanic sequence are numerous graded air-fall tuffs and flows of rhyolite compositions. A zone with biogenic or stromatolitic structures is also preserved.

These subtle lithologic differences may reflect different levels of exposure and/or ages of accumulation. The Nondweni greenstones show a consistent northwesterly younging direction in rocks which are not highly strained and which are separated by poorly exposed areas of high strain, suggestive of tectonic interslicing. In contrast the Assegaai and Commondale rocks show evidence of early reclined folds, which may be a reflection of deeper infolding. Preliminary geochronologic data indicate

that the Dwalile 'greenstones' are of similar age to the Barberton greenstones(1). Pb-Pb isotopic data from a single komatiitic flow at Nondweni define an age of 3.15 Ga that is consistent with an Rb-Sr age of ∿3.1 Ga for an associated rhyolite(4). However, Sm-Nd data define an age of 3.6 Ga for komatiitic, tholeiitic and rhyolitic flows. Possible explanations are that either the Pb-Pb and Rb-Sr systems were reset at ∿3.1 Ga subsequent to extrusion at ∿3.6 Ga, or, on eruption 3.1 Ga ago, the extrusions interacted with ∿ 3.5 Ga-old felsic crust leading to a range of initial Nd isotopic compositions of the mafic rocks and the generation of rhyolites by remelting of that crust(4).

Subsequent to the D1 event (Table 1), mantle-derived tonalitic plutons (Tsawela and Braunschweig) and the meta-anorthositic Mponono layered intrusive sheet were emplaced into the bimodal gneisses and Dwalile greenstones. All these rocks were strongly and repeatedly deformed under amphibolite-facies conditions (Table 1).

Sheet-like granitoid batholiths were intruded at ∿3.2 and ∿3.0 Ga, the locus of emplacement migrating northwards with decreasing age. The ∿3.2 Ga-old multiphase sodic granitoid intrusion screens the Assegaai and Commondale greenstone remnants from their underlying gneissic basement. Intrusion occurred in the interval between D1 and D2 in the Assegaai and Commondale areas. A chemically and mineralogically similar granite also intrudes the Nondweni 'greenstones' but neither its age nor structural style have yet been studied.

At a high structural level, a second sheet-like, but more potassic granite, the vast multiphase Lochiel batholith, was intruded at ∿3.0 Ga north of Dwalile. Following this period of widespread emplacement of granitic magmas emergence above sea-level of stable continental crust took place. Subaerial weathering of this dominantly granitoid terrane was accompanied in the north by the development of braided stream systems draining southeast off the flank of the NE-trending Lochiel batholith(5) into the Pongola basin(6). Minor contemporaneous volcanism accompanied the fluvial sedimentation and heralded a period of subaerial extrusion of lavas (the 2.94 Ga-old Nsuze Group), that range in composition from basalt to rhyolite and attain a thickness of ∿ 8.5 km SE of Piet Retief(7). No ultramafic nor high-MgO flow units are present and the sequence is characterized by the simultaneous extrusion of mafic and acidic lavas. Typically porphyritic andesites are also present.

The Nsuze Group is preserved in a series of inliers in the south where its thickness decreases in part due to truncation by the upper (Mozaan) group of the Pongola Supergroup or by the Palaeozoic Natal Group. Volcanic rocks are less abundant in the southern inliers. Shallow water subtidal and tidal-flat sediments including stromatolitic carbonate sands are prominent in the Wit Mfolozi inlier. A heterolithic unit 1.5 km thick dominated by pyroclastic rocks interlayered with shallow marine sediments forms the base of the Nsuze Group south of Babanango. This unit is truncated towards the east by a 4.0 km thick sequence of tidalite sediments with interlayers of basaltic andesite lavas. Transport directions in the inliers are from the north and northwest.

Sedimentation in the Mozaan group was largely controlled by the interaction of a braided alluvial plain and a macrotidal basin(8). Mozaan sediments are not preserved south of the Wit Mfolozi inlier either as a result of removal by erosion or of non-deposition.

The Mozaan Group is typically deformed into gently dipping, doubly plunging synclinal structures resulting from interference of NW and NE-trending axial traces. Adjacent to the southern margin of the Kaapvaal Province, tight E-trending folds with vertical axial surfaces are dominant reflecting a response to deformation related to the development of the Natal thrust zone at ∿ 1.1 Ga. The Nsuze Group is highly strained adjacent to the Swaziland border apparently related to a 20 km wide belt of NW-trending folds and faults with left-lateral movement within which the dyke-like, mafic Usushwana Intrusive Suite was emplaced at ∿ 2.87 Ga(9).

The significance of the Pongola Supergroup lies in the fact that it demonstrates the co-existence of stable continental crust in southeastern Africa and metastable crustal conditions in southern central Africa dominated by extrusion and intrusion of voluminous komatiitic and tholeiitic magmas.

Emplacement of large volumes of granitic magmas principally into Pongola rocks terminated Archaean evolution. Multiple gneiss domes separated by screens of Mozaan sediments of high metamorphic grade developed in southern Swaziland adjacent to the belt of NW-striking, highly strained Nsuze rocks. Subsequently a thin sheet (300 to 1000 m thick) of potassic granite was emplaced at the unconformity between the Mozaan Group and its gneissic granitoid basement. The final pulses of granite plutonism resulted in the emplacement of sharply transgressive, typically coarse-grained, porphyritic plutons ranging in size from 40 km^2 to 650 km^2 about which narrow contact aureoles are developed in the Mozaan sediments. Rb-Sr isotopic data have yielded only whole-rock errorchrons for these rocks(10).

The concentration of post-Pongola granitoids within the core of the Pongola depository suggests that depression of the depositional basin promoted partial melting of the lower crust, which would be consistent with the proposed model for the genesis of the granitic melts based on geochemical data(11). The post-Pongola granites differ in their setting from other Archaean granites in southern Africa(5).

1. Carlson R.W. Hunter D.R. and Barker F. (1985) Abstracts, Spring Meeting, Amer. Geophys. Union.
2. Jackson M.P.A. (1984). In Precambrian Tectonics Illustrated E. Schweizerbart'sche Verlagsbachhandlung, Stuttgart, p 1-18.
3. Talbot C.J. Hunter D.R. and Allen A.R. (in press) Deformation of the Assegaai supracrustals and adjoining granitoid rocks.
4. Wilson A.H. Carlson R.W. Versfeld J.A. and Hunter D.R. (1985). Abstracts. Fall Meeting, Amer. Geophys. Union.
5. Hunter D.R. (1974) Precambrian Res., 1, p 259-326.
6. Watchorn M.B. and Armstrong N.V. (1980). Trans. Geol. Soc. S. Afr., 83, p 231-238.
7. Armstrong N.V. Hunter D.R. and Wilson A.H. (1982). Precambrian Res., 19, p 75-107.

8. Watchorn M.B. (1980) Precambrian Res., 13, p 27-42.
9. Hegner E. Kröner A and Hoffman A.W. (1984). Earth Planet.-Sci. Lett., 70, p 267-279
10. Barton Jr. J.M. Hunter D.R. Jackson M.P.A. and Wilson A.C. Trans. Geol. Soc. S. Afr., 86, p 71-80.
11. Condie K.C. and Hunter D.R. (1976). Earth Planet. Sci. Lett., 29, p 389-400.

TABLE I SUMMARY OF GEOLOGICAL AND STRUCTURAL EVENTS PRE-PONGOLA

SOUTHWESTERN SWAZILAND (Jackson, 1984)			ASSEGAAI (Talbot, et al., in press)			COMMONDALE (Smith and Hunter, in prep.)		
EVENTS			EVENTS			EVENTS		
AGE	GEOLOGICAL	STRUCTURAL	AGE	GEOLOGICAL	STRUCTURAL	AGE	GEOLOGICAL	STRUCTURAL
~3.5	Bimodal layered gneisses. Dwalile metamorphic suite (ultramafic, mafic volcanics; minor clastic and chemical sedimentation)	D_1 :Flat-lying F folds	?	Bimodal layered gneisses. Assegaai supracrustal suite (ultramafic, mafic volcanics and sub-volcanic intrusions; clastic and chemical sedimentation important)	D_1 (early): Sub-horizontal and listric zones of cataclasis; D_1 (late): Transition to ductile flow; generation of F_1 eastward verging recumbent folds.	?	Bimodal layered gneisses. Commondale supracrustal suite (ultramafic, mafic volcanics and sub-volcanic intrusions; minor clastic and chemical sedimentation)	D_1 : generation of recumbent F_1 folds; thrusts parallel to axial surfaces of F_1 folds (?)
~3.3	Intrusion of Tsawela tonalite;Mafic dykes		~3.2 (?)	Intrusion of Anhalt sodic granite sheet		?	Intrusion of Braunschweig tonalite	
?	Intrusion of anorthositic Mponono Intrusive Suite	D_2 : Intense F_2 folding on flat-lying XY planes of bulk strain				?	Intrusion of Bazane sodic granite sheet; hornblende granodiorite and related leucotonalite intrusions	
?	Mafic dykes	D_3 : Local ductile shear zones D_4: Tight to isoclinal F_4 folds; vertical to gently dipping NE trending axial surfaces D_5: Gently dipping ductile shear zones, displacement upwards to NW			D_2 : F_2 concentric parallel folds with steeply dipping NNE-trending axial surfaces D_3 : Minor disharmonic conjugate folds; NE-trending strike-slip faults in metasediments D_4 : NW-trending mylonite zones			D_2 : F_2 folds with steeply dipping NNE/NE-trending axial surfaces D_3 : Mylonitic shearing along granite-supracrustal contacts; faults in supracrustals D_4: Open to tight F_4 folds with steeply dipping NW-trending axial surfaces D_5: WNW and NW mylonitic shearing
~3.2	Intrusion of quartz monzonite pods							
~3.0	Multiphase intrusion of Lochiel granite Sheet	D_6: Folding and mylonitic re-foliation of D_2 fabrics near lochiel granite	~3.0 ~2.96	Multiple intrusion of Lochiel granite sheet Pegmatitic granite (?)				

IS THE CAMERON RIVER GREENSTONE BELT ALLOCHTHONOUS?;

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Many tectonic models for the Slave Province, N.W.T., Canada, and for Archean granite - greenstone terranes in general, are implicitly dependent on the assumption that greenstone belt lithologies rest unconformably upon older gneissic basement. Other models require originally large separations between gneissic terranes and greenstone belts. A key question relating to the tectonics of greenstone belts is therefore the original spatial relationship between the volcanic assemblages and presumed-basement gneisses, and how this relationship has been modified by subsequent deformation. Unconformities have been reported from the Cameron River Greenstone Belt northeast of Yellowknife, and from the Point Lake area to the north (1,2). What remains unclear in these examples is the significance of the so-called "later faulting" of the greenstone - gneiss contacts. Do the angular discordances really represent unconformities, or could they be better - interpreted as a consequence of the juxtaposition of originally widely separated terranes? Where unconformities between gneisses and overlying sediments are indisputable, such as at Point Lake, the significance of faults which occur below the base of the volcanic succession also needs to be evaluated. As part of an on-going investigation aimed at answering these and other questions, I mapped the extremely well-exposed Cameron River Greenstone Belt and the Sleepy Dragon Metamorphic Complex in the vicinity of Webb Lake and Sleepy Dragon Lake during the summer of 1985, extending the efforts of earlier workers (3,4,5,6).

The greenstone belt was found to consist predominantly of mafic pillowed to massive flows and numerous dike complexes. At the preserved base of the greenstone belt these dikes locally retain a sheeted aspect and display one-way chilling. Subordinate amounts of pyroclastic rocks and volcanic breccias are also present. Rocks of the Sleepy Dragon Metamorphic Complex are highly variable, and include both ortho- and paragneisses, along with numerous mylonite zones (3,7,8). Older gneisses and mylonites are intruded by several younger phases of mafic to silicic plutonic rocks which show different intensities of deformation.

The contact between the Cameron River Greenstone Belt and the Sleepy Dragon Metamorphic Complex was found to be a half-kilometer wide zone of very complex structure. All rocks within this high-strain zone have a strong steeply plunging stretching lineation, although rocks from throughout the area also have a less-intense generally vertical lineation. Transposed layering and intensely folded quartz segregations are common in this zone; sheath folds with verticaly plunging hinges are present in some localities, indicating very high shear strains. Macroscopic sense-of-shear indicators are not abundant but generally suggest that the Cameron River Belt was thrust over the Sleepy Dragon Complex. Supporting microscopic work is

currently underway. In one area north of Webb Lake some slivers of gneissic basement are intercalated with phyllonites of the major high-strain zone; further mapping will reveal the lateral extent of this tectonic juxtaposition, but it is apparently the first-documented example of Archean basement-involved thrusting in the Slave Province.

Pillow lavas of the Cameron River Belt immediately adjacent to the basal high-strain zone have aspect ratios locally exceeding 1:3:10. In an area to the northwest of Sleepy Dragon Lake these lavas are overturned in an anticline as shown by locally-consistent facing directions. The axial trace of this fold is parallel to the contact zone, and the fold's geometry is consistent with formation during thrusting of the Cameron River Belt over the Sleepy Dragon Complex. Preliminary mapping in the greenstone belt in the Webb Lake area has revealed the presence of a few other subparallel shear zones containing structures similar to those just described; a common origin is tentatively inferred pending more detailed mapping.

Interpreting the structures within the Sleepy Dragon Metamorphic Complex is difficult because of the complex deformation history of this terrane. The only structure which, at this point, can unambiguously be related to movement along the contact with the Cameron River Belt is a foliation which trends parallel to and increases in intensity towards the contact zone. The foliation cuts earlier structures including folded gneissic and mylonitic foliations; earlier foliations are folded about this later one (3). The fact that this late foliation is cut by some plutonic bodies suggests that a minimum age may be placed on the thrusting and emplacement of the Cameron River Greenstone Belt over the Sleepy Dragon Metamorphic Complex.

Numerous mafic dikes are present both at the base of the Cameron River Belt and within the Sleepy Dragon Complex near it's contact with the greenstone belt (6). The textures and xenolith content of the dikes in the Sleepy Dragon Complex appear to be generally different from the dikes in the greenstone belt. Deformational and metamorphic fabrics in the dikes of the Sleepy Dragon Metamorphic Complex suggest that they are of at least two, and probably three generations, while only two distinct generations of dikes are recognized from the Cameron River Greenstone Belt. Pending further field and laboratory work it is tentatively suggested that (a) the first two generations of dikes in the Sleepy Dragon Complex are not directly related to any dikes in the greenstone belt, (b) the earliest generation of (locally sheeted) dikes in the greenstone belt are not present in the basement complex, and (c) only the latest, relatively undeformed dikes are correlatable between the two terranes.

Although it is a bit premature to propose tectonic models for the Cameron River Greenstone Belt it is useful to keep a working hypothesis in hand. It is tentatively proposed that the Sleepy Dragon Complex is a preserved remnant of a rifted Archean continent; many of the metasedimentary gneisses may represent a

highly deformed Atlantic-type margin sequence originally deposited on top of older basement. The Cameron River Greenstone belt has many affinities with Phanerozoic ophiolites and/or island arc complexes, including the presence of sheeted dikes. Although an unconformable relationship has been reported between the Cameron River Belt and the Sleepy Dragon Complex, I have not yet been able to support this contention based on observed field relationships. In fact, all data collected to-date indicates that the greenstone belt is allochthonous. Structures at and near the base of the greenstone belt suggest that it has been imbricated and thrust over the Sleepy Dragon Metamorphic Complex, incorporating slices of gneiss in the process. It must be emphasized that these are only preliminary conclusions that need to be verified by several more seasons of detailed mapping but, so far, many similarities are seen between Slave Province greenstone belts and Phanerozoic collisional tectonic zones.

ACKNOWLGEMENTS

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BIBLIOGRAPHY

(1) Baragar, W.R., and J.C. McGlynn, 1976, G.S.C., Paper 76-14
(2) Henderson, J.B., 1981, in: A. Kroner (ed.), Precambrian Plate Tectonics, pp. 213 - 235
(3) Davidson, A., 1972, in: Report of Activities, Part A, G.S.C., Pap. 72-1a, pp. 109 - 115
(4) Lambert, M.B., 1974, in: Report of Activities, April to October, 1973, G.S.C., Pap. 74-1, Pt. A, pp. 177-179
(5) Lambert, M.B., 1977, G.A.C., Special Paper no. 16, pp. 331 - 340
(6) Lambert, M.B., 1982, G.S.C., Paper 82-1A, pp. 165 - 167
(7) Henderson, J.B., 1976, G.S.C. Open File 353
(8) Kusky, T.M., 1986, Cont. Geol. of the N.W.T., vol. 3, in press

SEDIMENTOLOGICAL AND STRATIGRAPHIC EVOLUTION OF THE SOUTHERN PART OF THE BARBERTON GREENSTONE BELT: A CASE OF CHANGING PROVENANCE AND STABILITY; Donald R. Lowe and Gary R. Byerly, Department of Geology, Louisiana State University, Baton Rouge, Louisiana 70803 USA

The sedimentological and stratigraphic evolution of the 3.5 to 3.3 Ga Barberton Greenstone Belt can be divided into three principal stages: (1) the volcanic platform stage during which at least 8 km of mafic and ultramafic volcanic rocks, minor felsic volcanic units, and thin sedimentary layers (Onverwacht Group) accumulated under generally anorogenic conditions, (2) a transitional stage of developing instability during which widespread dacitic volcanism and associated pyroclastic and volcaniclastic sedimentation was punctuated by the deposition of terrigenous debris derived by uplift and shallow erosion of the belt itself (Fig Tree Group), (3) an orogenic stage involving cessation of active volcanism, extensive thrust faulting, and widespread deposition of clastic sediments representing deep erosion of the greenstone belt sequence as well as sources outside of the belt (Moodies Group).

I. The platform stage of Barberton Greenstone Belt development is represented by rocks of the predominantly volcanic Onverwacht Group. Sediments deposited during this stage included (a) dacitic breccias, conglomerate, and coarse sands deposited as part of and adjacent to felsic volcanic centers and, less abundantly, proximal mafic lapillistones and tuffs; (b) distal felsic volcaniclastic and pyroclastic layers consisting mainly of fine ash, dust, and accretionary lapilli, (c) biogenic deposits such as carbonaceous oozes, carbonaceous muds, bacterial mats, and locally, stromatolites, and (d) orthochemical sediments including evaporites, barite, carbonate, and possibly siliceous deposits. The bulk of these sedimentary units show clear evidence of having been deposited under shallow-water conditions. The regional stratigraphic continuity and sedimentological integrity of sedimentary layers within this sequence, the predominantly shallow-water depositional setting, and the paucity of debris derived from the uplift and erosion of older rock sequences indicate that the overall depositional and tectonic setting was a broad, low-relief, shallow-water anorogenic platform (1).

II. Rocks traditionally assigned to the Fig Tree Group were deposited during a transitional phase of greenstone belt evolution. These are exposed in a complex succession of thrust sheets that provide numerous exposures of each part of the stratigraphic sequence (2). The lowest part of the Fig Tree is characterized by distal volcaniclastic units and carbonaceous cherts resembling those in the Onverwacht but showing rapid lateral facies changes. In particular, 40 to 50 m of predominantly carbonaceous chert in some structural belts can be correlated with a sequence of interbedded ultramafic lavas, banded cherts, carbonaceous cherts, stromatolites, and volcaniclastic units at least 500 m thick in other areas (2).

The overlying 200 to 500 m of rocks includes two principal components. By far the greatest thicknesses of Fig Tree strata consist of heavily altered dacitic pyroclastic and volcaniclastic detritus (3). This succession includes three main lithofacies: (a) plagioclase-phyric intrusive rocks that may locally grade into extrusive flows, (b) proximal, plagioclase-phyric breccias and conglomerates, probably developed as lava domes and surrounding coarse epiclastic units, and (c) regionally extensive ash

deposits, tuffs, and their current-worked equivalents, volcaniclastic sandstone and siltstone. The bulk of the finely laminated cherty ferriginous sediments characterizing Fig Tree rocks throughout much of the Mountain Land represent altered fine-grained dacitic volcaniclastic deposits. In contrast to previous interpretations, we consider the Fig Tree to represent a predominantly volcanic interval, perhaps more closely related petrogenetically to the Onverwacht Group than to the suprajacent orogenic Moodies succession.

Interbedded with these volcanic and volcaniclastic strata are thin, lenticular units of chert-pebble conglomerate and chert-grit sandstone showing rapid lateral facies changes and apparently representing debris derived from local uplifts within the greenstone belt. Most of the debris can be identified with underlying silicified rocks of the Fig Tree Group; there is little evidence for major uplift or deep erosion of the greenstone belt at this time.

III. Rocks which have traditionally been included within the Moodies Group represent three main clastic lithofacies: (a) a sequence of quartz-poor, highly altered sands and fine gravels derived by erosion of the subjacent dacitic rocks; (b) thick, coarse, chert-clast conglomerate and chert-grit sandstone derived by weathering and erosion of uplifted parts of the greenstone belt, and (c) quartzose and locally K-spar-rich sandstone representing the erosion of sources outside of the greenstone belt, possibly but not necessarily including the intrusive granitoid rocks and/or the Ancient Gneiss Complex or its equivalents.

Although the stratigraphic sections in most structural belts can be correlated with one another, there is as yet no satisfactory reconstruction of their original relative depositional positions. So-called northern facies rocks in the Mountain Land also belong to allochthonous terranes and their present location relative to units to the south is clearly of tectonic rather than depositional origin.

The overall sequence includes numerous minor unconformities and at least one major break. Within the Onverwacht Group, pauses in effusive activity are marked locally by weathering and erosion of flow surfaces, but no significant formation or accumulation of clastic debris. The inception of felsic volcanism both in the upper Hooggenoeg formation and the Fig Tree Group was accompanied by minor instability and local erosion of underlying rocks. Also, the formation of large, high-relief subaerial felsic volcanic edifices in Hooggenoeg and Fig Tree times was followed by extensive erosion and truncation of these complexes. The major structural unconformity within the Barberton sequence occurs locally at the base of the Moodies Group. Although a number of apparently conformable Fig Tree-Moodies transitions occur, over wide areas, the Moodies was deposited with angular unconformity on rocks as old as the Hooggenoeg Formation. This contact has additionally been complicated by structural movement.

The sedimentological development of the Barberton Greenstone Belt reflects three principal tectonic stages involving three contrasting sources of clastic sediment. The volcanic platform stage, represented by rocks of the Onverwacht and Fig Tree Groups, was primarily an interval of rapid effusion of lavas, subsidence, but little differential tectonic movement. The main sources of clastic detritus were first cycle, active, high-relief, felsic and, to a lesser extent, mafic volcanic centers. The second stage, represented by rocks of the Fig Tree Group, was one characterized by continuing, regionally extensive volcanism and developing

tectonic instability reflected by the presence of extensive lateral facies changes and small intra-platform uplifts that supplied shallow-level intraformational debris to local sedimentary systems. Latest Fig Tree and Moodies deposition was influenced by concurrent thrusting and orogenesis. Sediments were derived initially from both shallow and deep levels within the greenstone belt and, later, from distant quartz and K-spar rich sources outside of the belt.

REFERENCES: 1. Lowe, D. R. (1982) Precam. Res., 17, 1-29. 2. Lowe, D. R., et al. (1985) Precam. Res., 27, 165-186. 3. Byerly, G. R., and Lowe, D. R. (1984) Proc. Lunar Planet. Sci. Conf. 16th, p. 101-102.

EVIDENCE FOR STRUCTURAL STACKING AND REPETITION IN THE GREENSTONES OF THE KALGOORLIE DISTRICT, WESTERN AUSTRALIA

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INTRODUCTION

Most previous stratigraphic interpretations of the southern part of the Norseman-Wiluna Greenstone Belt have proposed polycyclic sequences (e.g. Horwitz and Sofoulis[1]; McCall[2]; Williams[3]; Glikson[4]; Gemuts and Theron[5]). These invoked two and sometimes three successive suites of mafic and/or ultramafic volcanics and intrusives separated by felsic volcanics and immature clastic sediments, however no distinctive lithological differences were reported between successive mafic-ultramafic sequences. When interpretations of Williams et al.[6] and Hallberg[7], further to the north, are integrated, a total of four separate major mafic-ultramafic suites emerges for a large part of the Norseman-Wiluna Belt. Although the author does not intend to imply that all polycyclic stratigraphies are wrong in principle such a situation seems suspiciously over-complex and stimulates the need to look critically at the individual areas where stratigraphies have been erected. For the Kalgoorlie area in the south, some of the schemes have already provoked scepticism (Burke et al.[8]; Archibald et al.[9]) and a simpler model consisting of one cycle subject to structural repetition has been evolved by workers in the Geological Survey of Western Australia (Griffin et al.[10]) for part of this area. The latter authors drew attention to the 'carbon copy' similarity between the elements of some polycyclic stratigraphies. Much more regionally extensive integrated structural and stratigraphic data is still required to evaluate the relationship between structure and stratigraphy more fully, an objective substantially limited by poor outcrop and deep weathering, but with due effort, far from unattainable.

OUTLINE OF STUDY

Regional mapping by the author in an area of approximately 20,000 km^2 centred on Kalgoorlie revealed many problems and anomalies in several of the published stratigraphic schemes. However since insufficient critical stratigraphic and structural evidence had been given in support of the schemes it has not been easy to check the bases on which they were erected. The following lines of investigation have been pursued.

* Regional distribution and interelationships of lithologically similar sequences previously regarded as distinct, based on mapping, mineral exploration data, and geophysical interpretation. Emphasis has been on the mafic-ultramafic suites because they are the most easy to define and map.

* Critical evaluation of contacts and their associated structural features.

RESULTS

There are several instances where mafic-ultramafic suites previously proposed as younger (e.g. Coolgardie-Kurrawang area in Glikson[4]) join or merge with their 'older' counterparts when mapped over various distances. They range in size from splinter-like splays a few kilometres long diverging from a major mafic belt by up to a kilometre, to extensive sheets which are traceable for tens of kilometres as separate entities before joining with and becoming indistinguishable from their 'older' counterparts. Some successions are isolated in metasedimentary terrain, and never connect with their sequences of origin; however this situation is unusual. In areas where like elements of two proposed cycles are juxtaposed or interconnected (e.g. Widgiemooltha and Spargoville areas in map of Gemuts and Theron[5]) there seems to be no clear reason why they should have been regarded as separate.

The apparent stratigraphic thicknesses of many of the previously proposed younger mafic-ultramafic sequences is very variable. While they may be measurered in kilometres in some areas, in many localities the sequences are attenuated and deformed. They may be traced for tens of kilometres as apparently conformable packages of all or most of the major mafic and ultramafic lithologies, though individually these lithologies may occur as lenses or sheets hundreds or even only tens of metres in thickness. While such sequences have been interpreted in the past as volcanic intercalations in a eugeosynclinal sedimentary pile (McCall[2]; Glikson[4]), or the beginnings of new volcanic cycles, their degree of deformation and tendency to be smaller scale carbon copies of their 'older' counterparts is more consistent with structural repetition. In a number of instances they are overlain by felsic volcanic rocks suggesting cyclic development in a uniformly facing sequence. This is here regarded as evidence that repetition has been mainly by faulting and not by recumbent or isoclinal folding.

OBSERVATIONS ON CONTACTS

Many previous stratigraphies (e.g. Williams[3]; Gemuts and Theron[5]; Glikson[4]) have been erected in areas where fragmentary facing evidence suggests thick uniformly facing sequences. The potential for strike dislocations has generally been overlooked despite heterogeneous shear deformation. There has been an absence of critical treatment of major formational contacts to establish whether they are normal or tectonised. This is understandable in some instances since such contacts are rarely well exposed, however diligent search by the author has revealed many key outcrops. The vast majority of these provide compelling evidence that all is not well with the published polycyclic stratigraphies. Examination of contacts, especially basal ones, and the contact areas of the previously proposed younger mafic-ultramafic suites, commonly reveals strong peneconcordant shearing, recrystallised mylonitic or other cataclastic rocks, or in one instance (the Kalpini formation which is the highest mafic-ultramafic suite of Williams[3]), an overturned but undeformed contact with clear facing evidence the reverse of that previously proposed. Mapping the relationship between the contact zones and primary layering often reveals subtle discordances not readily explained by unconformity.

MODEL

Many proposed structural repetitions or fault slices are linear, others are arcuate and folded around major upright structures. Linear belts are often controlled by throughgoing transcurrent deformation zones with pronounced sub-horizontal lineations. Arcuate systems however were conceivably generated by earlier processes such as thrusting or gravitational gliding predating upright folding. Although transcurrent shearing is a feasible mechanism for repetition for at least some of the more linear belts, it is possible that even many of these began life as early thrust sheets and became stretched and aligned by later transcurrent deformation. Early thrusts, recumbent folds and layer-parallel shear fabrics have been documented in several localities in the Norseman-Wiluna Belt where prevailing strikes deviate from the NNW regional grain, or where tight upright folding is subdued or absent (e.g. Chapman[11]; Gresham and Loftus-Hills[12]; Archibald[13]; Platt et al.[14]; Martyn and Johnson[15]; Spray[16]). In one instance (Chapman[11]) a narrow mafic-ultramafic belt in sediments has clearly been generated by an overthrust. Almost certainly the recognition of this structural style in east-west trending or gently domed areas is a consequence of preservation. It is undoubtedly present also in NNW trending linear domains but is overprinted and hard to recognise. It is emphasised that thrust repetition does not explain all of the previously proposed younger cycles in the district. Some are a consequence of misinterpretation, by placing too great a significance on isolated stratigraphic facing observations, or from attempts to correlate across major upright faults. Broad regional observations by the author suggest that thrust repetition may be much more strongly developed in the Kalgoorlie district than elsewhere in the Norseman-Wiluna Belt though this conclusion is tentative.

Thrusting does not appear to have occurred on a scale comparable with many Phanerozoic convergent plate boundaries. There is no evidence of juxtaposition of strongly contrasting domains, or of high pressure metamorphism. There is also a lack of pronounced east-west assymetry across the Norseman-Wiluna Belt as a whole. The tectonics can be viewed more in terms of a rearrangement of familiar elements of the local stratigraphy, a situation more consistent with a closed or intracratonic setting, rather than an open plate margin. This accords with models such as those of Groves[17]. As such, intrabasinal gravity gliding resulting from early uplift heralding later vertical tectonic events is the most favoured model by the author. This is consistent with the sedimentation style which is dominated by turbidites and includes debri flow deposits. Olistostromes have also been reported (Taylor, in Gee and Groves [18]). In some respects the scheme resembles that proposed by De Wit[19] for the Barberton Greenstone Belt. Felsic volcanism was intimately associated with sedimentation, and it is possible that concomitant granitic intrusion into a dense sheet of mafic-ultramafic volcanics may have triggered the instability that first led to the sedimentation and later to gravity gliding tectonics. Subsequent folding and faulting of the tectonically stacked sequence would have created the illusion of a polycyclic sequence which has suffered only upright folding and shearing. The upright tectonic events have generated their own set of interpretive problems. Peneplenation, and Tertiary lateritic weathering ultimately obscured much of the important evidence.

REFERENCES

1. Horwitz, R.C., and Sofoulis, J. 1965. Proc. Aust. Inst. Mining Metall. 214, p.45-59.
2. McCall, G.J.H., 1969. Proc. Roy. Soc. West. Aust., 52, p.119-128.
3. Williams, I.R., 1970. Geol. Surv. West. Aust. Rec. No. 1970/1.
4. Glikson, A.Y., 1971. Spec. Publ. No.3, Geol. Soc. Aust. p.443-460.
5. Gemuts, I. and Theron, A.C., 1975. Aust. Inst. Mining Metall. Mon. 5, p.66-74
6. Williams, I.R., Gower, C.F., and Thom, R., 1971. Geol. Surv. West. Aust. Rec.No.1971/26.
7. Hallberg, J.A., 1983. Geol. Surv. West.Aust. Rec. No. 1983/8.
8. Burke, K., Davey, J.F., and Kidd, W.S.F., 1976. The early history of the Earth. John Wiley and Sons, p. 113-129.
9. Archibald, N.J., Bettany, L.F., Bickle, M.J. and Groves, D.I., 1981 Spec. Publ. No. 7 Geol. Soc. Aust. p.491-504.
10. Griffin, T.J., Hunter, W.M., and Keats, W. 1983. Eastern Goldfields Geological Field Conference, Excursion and Guide. Geol. Soc. Aust. p.7-8.
11. Chapman, D.M., 1982. B.Sc. Hons. Thesis, Univ. of West. Aust.
12. Gresham, J.J. and Loftus-Hills, G.D., 1981. Econ. Geol. 76, p.1373-1416.
13. Archibald, N.J., 1979. Ph.D. Thesis, Univ. of West. Aust.
14. Platt, J.P., Allchurch, P.D. and Rutland, R.W.R., 1978 Precamb. Res. 7, p.3-30
15. Martyn, J.E. and Johnson, G.I., in press. Aust. J. Earth Sciences.
16. Spray, J.G., 1985. J. Struct. Geol. 7(2) p.187-203.
17. Groves, D.I., 1982. Revista Brasiliera de Geosciencias, 12, p.135-148.
18. Gee, R.D., and Groves, D.I. 1982. Univ. of West. Aust. Geol. Dept. and Extension Service Publ. No. 7, p.C3-C10.
19. De Wit, M.J., 1982. J. Struct. Geol. 4, 117-136.

PRELIMINARY STRUCTURAL MODEL FOR THE SOUTHWESTERN PART OF THE MICHIPICOTEN GREENSTONE BELT, ONTARIO; George E. McGill and Catherine H. Shrady, Dept. of Geology and Geography, University of Massachusetts, Amherst, MA 01003

The southwestern part of the Michipicoten Greenstone Belt includes a 100 km^2 fume kill extending northeastwards from the town of Wawa, Ontario. Except for a strip along the Magpie River that is covered by Pleistocene gravels, outcrop in the fume kill averages about 30-50%. Within this area are all the major lithologic belts characteristic of the southwestern fourth of the Michipicoten Greenstone Belt. All of the area mapped to date lies within Chabanel Township, recently mapped at 4" = 1 mile by Sage et al. (1). Following a brief reconnaissance in 1983, mapping at a scale of 1" = 400' was begun within and adjacent to the fume kill in 1984. We have concentrated on two objectives: 1) determination of the geometry and sequence of folding, faulting, cleavage development, and intrusion; and 2) defining and tracing lithologic "packages", and evaluating the nature of the contacts between these packages. Results for objective 1) are discussed in a companion abstract (2); this abstract will present tentative results for objective 2).

The entire Michipicoten Greenstone Belt has experienced relatively late movement on steep faults, most of which trend approximately NNW or NE (1,2,3). Some of this movement preceded the emplacement of diabase dikes, some followed. These displacements may be easily removed in order to reassemble older structures, which are of much greater tectonic interest.

For mapping and descriptive purposes, it long has been customary to divide the stratified rocks of the Michipicoten Greenstone Belt into 4 major lithologic groups (1,3): mafic-intermediate volcanics, intermediate-felsic volcanics, clastic sediments, and chemical sediments (including iron formation). This is certainly valid, because outcrop belts of these groups maintain integrity for long distances. However, there are along-strike intergradations among them, and there is no easy way to correlate between physically separated belts of similar lithology. This last problem means that there is no really dependable belt-wide stratigraphy, and relative ages of the various belts of similar lithology are known only in the few places where modern radiometric ages have been measured (4,5).

Our detailed mapping (Fig. 1) indicates that the situation is more complex than one would infer from published maps and descriptions (1,3,6). There are several lithologic packages within the single belt of clastic sediments in Chabanel Township, all of which appear to be bounded by fault contacts. In some cases, stratigraphic way up reverses across these faults, in other cases it does not. At map scale, the package boundaries follow bedding or volcanic layering on one or both sides, but locally this is not so, and at outcrop scale it commonly

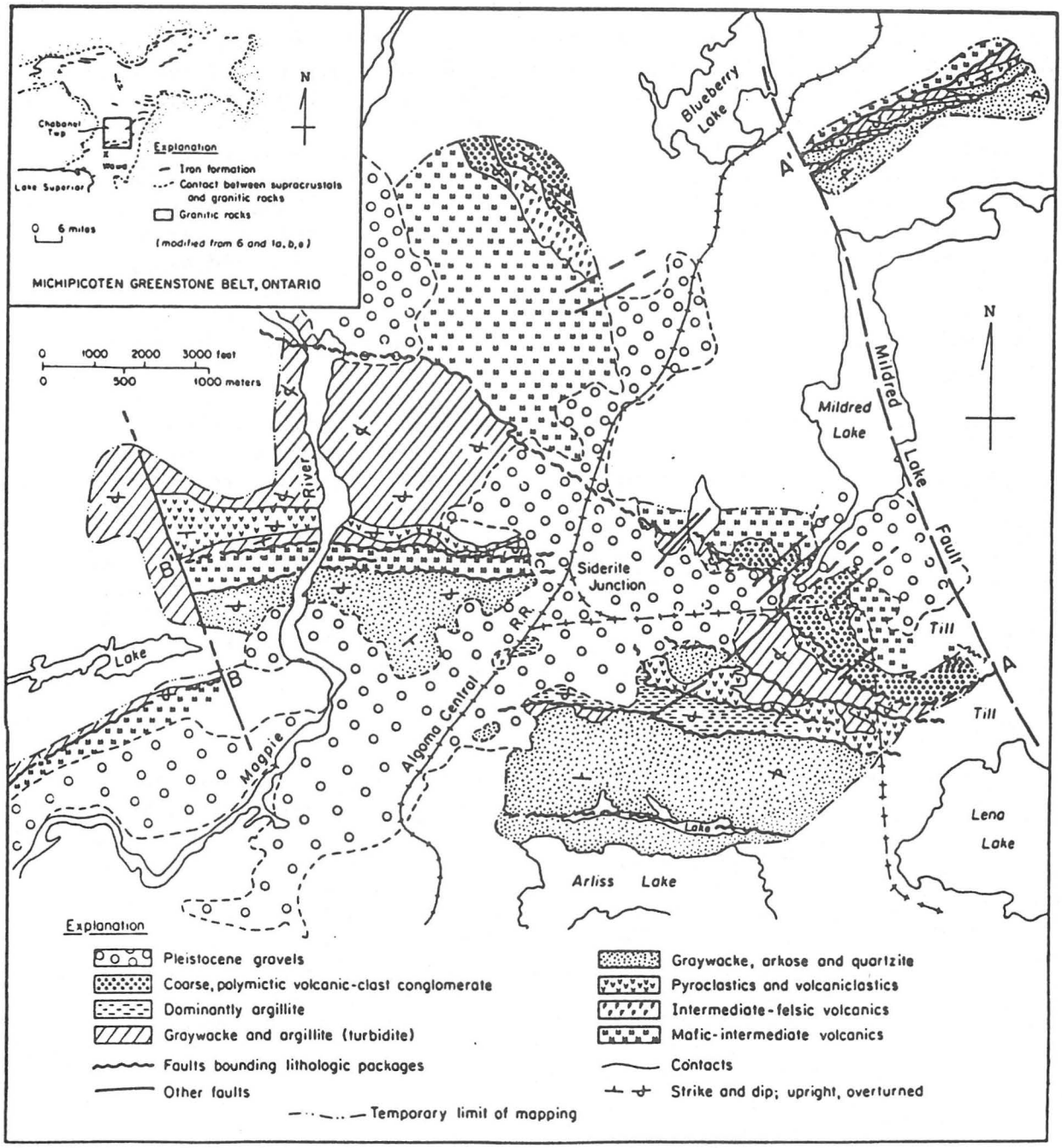

Fig. 1. Geologic sketch map of the central part of Chabanel Township, Ontario. All intrusive igneous rocks omitted for simplicity. B-B' and A-A' indicate corresponding points across late faults.

is not so. In places, these faulted boundaries are characterized by locally developed cleavages, excessive flattening or elongation of pebbles, or minor folds.

The area we have mapped seems to be a zone of faults and folds separating a large region to the south underlain by overturned rocks with tops north from an even larger region to the north underlain by overturned rocks with tops south (1,6). This relationship would seem to indicate an antiformal fold in the inverted limb of a very large nappe, but we have not been able to define such a structure, and rocks that should correlate across the structure are not the same age (R. Sage, pers. com.). Major faulting thus is necessary, but earlier or synchronous folding at township or larger scale would seem necessary to account for the opposed overturning. Almost all of the rocks north and south of our area are volcanic, so it may never be possible to determine if these terranes consist of continuous sections or if they, too, are divided into fault-bounded packages.

Because we have yet to sort out the sequence of minor and major structures with sufficient confidence, and because completed detailed mapping covers such a small fraction of the total belt, we prefer to be rather conservative about interpreting our data. Key observations include a "stratigraphy" that consists mostly of fault-bounded "packages", the apparent early age of these faults, and the large areal extent of the inverted sequences facing each other. The most attractive and probably the simplest explanation for these relationships involves early imbricate thrusting--before the imposition of the almost universal steep dips. However, this interpretation remains to be proved.

References

(1) Sage, R.P., et al., (1982) Ontario Geol. Surv. Prelim. Maps of Precambrian Geology: a. Chabanel Twp., Map P. 2439; b. Esquega Twp., Map P. 2440; c. McMurray Twp., Map P. 2441; d. Lastheels Twp., Map P. 2442; e. Musquash Twp., Map. 2568.
(2) Shrady, C.H. and McGill, G.E. (1986) This volume.
(3) Sage, R.P., et al., (1984) Prelim. Map of Precambrian Geology of Corbiere Twp., Ontario Geol. Surv. Map P. 2607.
(4) Turek, A., et al., (1982) Can. Jour. Earth Sci., 19, 1608-1626.
(5) Turek, A., et al., Can. Jour. Earth Sci., 21, 457-464.
(6) Goodwin, A.M. (1962) Geol. Soc. America Bull., 73, 561-586.

THERMAL IMPLICATIONS OF METAMORPHISM IN GREENSTONE BELTS AND THE HOT ASTHENOSPHERE-THICK CONTINENTAL LITHOSPHERE PARADOX; Paul Morgan, Department Geosciences, Purdue University, West Lafayette, IN 47907.

From considerations of secular cooling of the Earth and the slow decay of radiogenic heat sources in the Earth with time, the conclusion that global heat loss must have been higher in the Archean than at present seems inescapable. The mechanism by which this additional heat was lost and the implications of higher heat loss for crustal temperatures are fundamental unknowns in our current understanding of Archean tectonics and geological processes. Higher heat loss implies that the average global geothermal gradient was higher in the Archean than at present, and the restriction of ultramafic komatiites to the Archean and other considerations suggests that the average temperature of the mantle was several hundred degrees hotter during the Archean than today (1). In contrast, there is little petrologic evidence that the conditions of metamorphism or crustal thickness (including maximum crustal thickness under mountains) were different in Archean continental crust from the Phanerozoic record (see 1). Additionally, Archean ages have recently been determined for inclusions in diamonds from Cretaceous kimberlites in South Africa (2), indicating temperatures of 900 to 1300 degC at depths of 150 to 215 km (45 to 65 kbar) in the Archean mantle (3), again implying relatively low geothermal gradients at least locally in the Archean. In this contribution the thermal implications of metamorphism are examined, with special reference to greenstone belts, and a new thermal model of the continental lithosphere is suggested which is consistent with thick continental lithosphere and high asthenosphere temperatures in the Archean.

High-grade metamorphism is common in Archean terrains (4, 5), and includes some greenstone belts, such as in the Yilgarn block of SW Australia (6). High metamorphic temperatures (700 degC or more) and often high metamorphic pressures (5 to 10 kbar or greater) are indicated by the mineral assemblages in these terranes, and they are underlain in most cases by continental crust of normal thickness (7, 8). Conductive thermal relaxation models have been proposed to predict the thermal conditions of metamorphism in the crust following tectonic activity such as underthrusting (e.g., 9-11). As demonstrated by Ashwal and Morgan (7), however, simple thermal relaxation of thickened crust cannot reasonably produce the high temperatures required by granulite metamorphism with a thick section of crust (30 km or more) below the shallowest depth of granulite metamorphism without requiring the lower part of the crust to be supersolidus. Basically the temperature range for granulite metamorphism is so close to estimates of the crustal solidus for reasonable crustal compositions (e.g., 12), that a positive geothermal gradient below the shallowest depth of granulite metamorphism causes the geotherm to intersect the solidus above the Moho. Ashwal and Morgan (7) conclude that unless granulite metamorphism occurs only near the base of the crust and the thick section of crust now below the exposed granulites was added after metamorphism, major crustal magmatic activity is associated with granulite metamorphism. Such extreme thermal conditions are not required by lower grades of metamorphism, but any metamorphic gradients which indicate a high geotherm suggest the upward transport of heat by magma unless the crust is thin.

If it is accepted that magmatic heat transport is an essential component of the crustal thermal regime during the peak thermal conditions recorded by the metamorphic mineral assemblages in the crust (at least where high geothermal gradients are indicated), then maximum temperatures recorded in

these systems were buffered by the solidus. The occurrence of young granulites at the top of sections of normal thickness crustal sections similarly indicates that modern maximum geothermal gradients are buffered by the solidus. A similar conclusion is indicated by heat flow data from areas of recent tectonism in which high heat flow must result from magmatic heating of the crust (e.g., 13). Maximum temperatures at shallow depth are buffered by the boiling point curve at hydrostatic or lithostatic pressures, below which maximum temperatures are buffered by the crustal solidus. As these maximum crustal temperatures are commonly encountered in areas of active tectonism and magmatism today, it is impossible for maximum temperatures recorded by Archean metamorphic assemblages to have been higher than modern maximum temperature conditions unless the solidus was different. Thus, in this buffered system, higher heat loss in the Archean is not expected to be recorded by metamorphic assemblages indicating higher geothermal gradients than peak modern conditions, although these peak crustal thermal conditions may have been more widespread in the Archean than at present.

The occurrence of high-grade (granulite) metamorphism in Archean greenstone belts suggests that either the high-grade areas were produced near the base of the crust and subsequently the crust has been thickened below the high-grade terranes, and/or magmatism was an important process during the high-grade metamorphism. The intimate association of plutons with the greenstone belts in "granite-greenstone" terranes suggests the importance of magmatism during this high grade metamorphism, and is consistent with models which suggest basal melting of stacked simatic thrust sheets during the evolution of at least some greenstone belts (14-16).

Perhaps the most paradoxical indicator of Archean thermal conditions with respect to higher global heat loss is the relatively low Archean geothermal gradients indicated by the formation of diamonds of Archean age. The diamond stability field is consistent with geotherms predicted for modern shield areas with thick (150 km or greater) lithosphere (e.g., 13). Meyer (3) has suggested that diamonds were formed in the asthenosphere which in turn suggests that perhaps the higher temperatures deduced for the Archean mantle from the occurrence of komatiitic lavas were not universal. A more common interpretation of the diamond data is that they indicate the existence of thick "keels" of subcontinental lithosphere below at least some areas during the Archean (1, 16). However, as the lithosphere is intimately related to the thermal boundary of upper mantle convection, it would be expected that this boundary layer and the lithosphere would have been thinner during the Archean with higher global heat loss and mantle temperatures. A possible solution to this paradox may be found in the intrinsic heat production of continental lithosphere.

There are two basic variable parameters that control the stable thickness of the continental thermal boundary layer (lithosphere), the heat production within the layer and the heat input to its base (13, 17). The layer thins if heat input to its base increases, and thickens if the heat input decreases. This heat input depends upon the temperature difference between the lower portion of the stable boundary layer and the underlying convection cell, or more specifically the temperature gradient in the lowest portion of the layer. As this gradient decreases to zero, the heat input to the base of the lithosphere decreases to zero (negative gradients are not permissible in a stable thermal boundary layer). The thickness of stable continental lithosphere with zero heat input at its base is independent of the global heat loss, assuming that the heat can be lost elsewhere (oceanic and other continental lithosphere), and this may possibly be a mechanism for maintaining

thick continental lithosphere at a time of high global heat loss and high average mantle temperatures.

The condition for zero heat flux into the base of the stable continental lithosphere is that the temperature increase within the lithosphere due to its intrinsic radiogenic heat production creates a geotherm that is asymptotic to the asthenosphere isotherm (or adiabat with an adiabatic basal heat flux). For thick lithosphere this condition requires a small but significant component of heat production in the mantle lithosphere, and an example of such a heat production distribution and geotherm are given in Figure 1. This condition has the interesting property that thicker lithosphere is indicated for higher asthenosphere temperatures for similar heat production distributions. If heat production distributions of this type are realistic it is unlikely that they are accidental (see also 18), and the concentration of radiogenic heat production into the lithosphere by metasomatism and crustal building processes may be related to the stabilization of continental lithosphere.

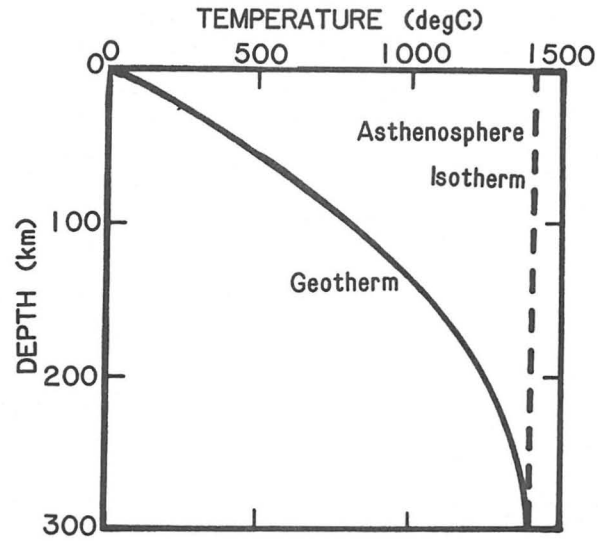

Figure 1. Example of continental lithosphere geotherm asymptotic with asthenosphere isotherm as a result of its intrinsic radiogenic heat generation. A two component crustal heat generation model is assumed for this geotherm: An upper crustal component decreasing exponentially with depth from 2.7 $\mu W/m^{3}$ at the surface with a depth scale length of 7 km, and an additional uniform component of 0.09 $\mu W/m^{3}$ (geotherm model modified from 19).

References: (1) Richter, F.M. (1985) Earth Planet Sci. Lett. 73, 350-360. (2) Richardson, S.H., Gurney, J.J., Erlank, A.J. and Harris, J.W. (1985) Nature 310, 198. (3) Meyer, H.O.A. (1985) Am. Mineral. 70, 344-355. (4) Condie, K.C. (1981) Archean Greenstone Belts, Elsevier, Amsterdam. (5) Windley, B.F. (1982) The Evolving Continents, Wiley, Chichester. (6) Gale and Klein (1981) in (5). (7) Ashwal, L.D. and Morgan, P. (1984) Heat sources for granulite metamorphism, unpublished manuscript. (8) England, P. and Bickle, M. (1984) J. Geol. 92, 144-146. (9) Bickle, M.J., Hawkesworth, C.J., England, P.C. and Athey, D.R. (1975) Earth Planet. Sci. Lett. 26, 13-28. (10) England, P.C. and Richardson, S.W. (1977) J. Geol. Soc. Lond. 134, 201-213. (11) Thompson, A.B. (1981) Terra Cognita 1, 11-20. (12) Wyllie, P.J. (1977) Tectonphysics 43, 41-71. (13) Morgan, P. (1984) Phys. Chem. Earth 15, 107-193. (14) McGregor, V.R. (1979) in Trondjhemites, Dacites and Related Rocks (F. Barker, ed.), Elsevier, Amsterdam, 133-147. (15) Barker, F., Arth, J.G. and Hudson, T. (1981) Phil. Trans. Roy. Soc. Lond. A301, 293-303. (16) de Wit, M. (1985) What the oldest rocks say (abs.), Workshop on the Earth as a Planet, LPI and GSA, Orlando, Florida, Oct. 27, 1985. (17) Morgan, P. and Sass, J.H. (1984) J. Geodynamics 1, 143-166. (18) Morgan, P. (1985) J. Geophys. Res. 90, C561-C570. (19) Davies, G.F. and Strebeck, J.W. (1982) Geophys. J. R. Astr. Soc. 69, 623-634.

GEOCHEMICAL CHARACTERS AND TECTONIC EVOLUTION OF THE CHITRADURGA SCHIST BELT: AN ARCHAEAN SUTURE (?) OF THE DHARWAR CRATON, INDIA. S.M.NAQVI, NATIONAL GEOPHYSICAL RESEARCH INSTITUTE, HYDERABAD, INDIA.

The Chitradurga schist belt extending for about 450 km in a NS direction and 2-50 km across, is one of the most prominent Archaean (2.6 b.y.) tectonic features of the Indian Precambrian terrain, comprising about 2 to 10 km thick sequence of volcanosedimentary rocks. The basal unit of this belt is composed of an orthoquartzite-carbonate facies, unlike many other contemporary greenstone belts of the Gondwana land which begin with a basal mafic-ultramafic sequence. Eighty percent of the belt is made up of detrital and chemogenic sediments, their sucession commencing with a poorly preserved quartz pebble basal conglomerate and current bedded quartzites which, in turn, rest on tonalitic gneisses, the latter having been further remobilized alongwith the schist belt. Deposition of current bedded mature arenites indicate the existence of platformal conditions near the shore line. Polymictic graywacke conglomerates, greywackes, shales, phyllites, carbonates, BIFs (oxide, carbonate and sulfide) BMF's (Banded Maganese Formations) and cherts thus constitute the main sedimentary rocks of the belt. The polymictic conglomerates contain debris of rocks of older greenstone sequences, as well as an abundant measure of folded quartzites, BIF's and gneissic fragments which represent earlier orogenies.

Four different types of greywackes are recognised in the belt from N to S. Most of these have been derived from the surrounding tonalitic gneisses which contained older greenstone sequences as enclaves of various dimensions. However, the younger sequences in the north contain debris from the intrabasin volcanism also. The K-granites and gneisses are found to be progressively abundant in the source area of these graywackes as indicated by the granitic component of the debris of the younger graywackes sequences. Their REE patterns are characterized by both positive and negative Eu anomalies, the latter especially in the interbedded shales with grewackes. Geochemistry of the graywackes and chemogenic sediments thus indicate their deeper oceanic environment of formation. Although stratigraphic relation between the shallow water and deeper water sediments is uncertain, the basal orthoquartzites-carbonate sequences indicating platformal environment perhaps represent a facies change due to shallow water conditions along the shore line, and the greywacke suite those of deeper water away from it. Similar facies change is observed in the BIF's from shallower oxide to deeper sulfide facies.

The ultramafic rocks, mostly found in the lower sections of the belt, show pillow structures and spinifex texture and are komatiitic in composition. The mafic, intermediate and acid volcanics are found as detached outcrops in presumably higher

stratigraphic sections and show tholeiitic and calc alkaline affinities, probably produced by 5-15% melting. The ultramafic lavas were produced by deeper mantle melting source, the geochemical characteristics belonging to the oceanic class.

Most of the rock suites in the belt have been metamorphosed to greenschist facies. However, its eastern margin is found to be in thrust contact with the higher amphibolite facies rocks (700°C at 6-7 Kbr), and the southern part near Mysore consist of predominantly ultramafic rocks metamorphosed to amphibolite and granulite facies. The northern part of the belt near Gadag is least metamorphosed. Irrespective of the grade of metamorphism or of inferred ages of the various stratigraphic groups, the belt shows a remarkable structural homogeneity of 3 phases of deformation from N to S and E to W and a convexity towards East. Both major and minor F1 folds are tight isoclinal with shallow to steep plunges and subvertical to subhorizontal axial planes. The variation in the attitude and orientation of the F1 axes has been controlled by the F2 episode which has coaxially folded both the subparallel bedding and the first generation axial plane schistocity cleavage. Only at F1 hinges the intersection between S1 and S2 is discernible. F3 is found as general warps on F2 limbs. The F1 axial plane schistocity cleavage and F2 crenulation charge are generally dipping (horizontal to subvertical) towards the east. High grade rocks on the eastern margin have been thrust westwards over the low grade central part. Structural data indicate considerable crustal shortening along the belt. Inversion of stratigraphic sequence is reflected, at many places by the youning directions obtained from current bedding, graded bedding and pillow couvexities. Horizontal compression and collision tectonics therefore, appear to have played a significant role in the development of the structural configuration of the belt.

As the 3000 m.y. old gray banded gneisses, found on the eastern and western sides of the Chitradurga schist belt are similar, the existing observations suggest the following two possible models: i) The belt developed in a rift on the juvenile Archaean continental crust which collapsed upon loading by sediments, resulting in a shallow subduction and horizontal compression. (ii) The belt evolved on an "Oceanic" crust between two juvenile continental blocks to the East and West. Shallow subduction and horizontal movement of the Eastern block would then result in the present structural geometry and consequent welding of the two along this probable suture.

GEOCHEMICAL AND ISOTOPIC CONSTRAINTS ON THE TECTONIC SETTING OF THE SERRA DOS CARAJAS BELT, EASTERN PARA, BRAZIL; Wm. J. Olszewski, Jr., Dept. of Earth Sciences, Univ. of New Hampshire, Durham, NH 03824; Allan K. Gibbs and Karl R. Wirth Institute for the Study of the Continents, and Dept. Geol. Sci., Cornell University, Ithaca, NY 14853

The lower part of the Serra dos Carajas belt (Fig 1) is the metavolcanic and metasedimentary Grao Para Group (GPG) (1-6). The GPG is thought to unconformably overlie the older (but undated) Xingu Complex, composed of medium and high-grade gneisses and amphibolite and greenstone belts. The Lower Metavolcanic Sequence of the Grao Para Group (LMS) is estimated to be about 4-6 km thick, consisting of massive, vesicular, and porphyritic mafic volcanic flows and agglomeratic breccias and about 10-15% massive, flow-banded, brecciated, and tuffaceous prophyritic rhyolite (6). The LMS is overlain by the extensive, 100-400 m thick, and high-grade banded iron formations of the Carajas Formation, followed by an Upper Sequence (US) of 1-3 km of mixed volcanic and clastic and chemical sedimentary rocks. The stratigraphy of the US is poorly known, but it is thought to contain some quartz-rich arenites, suggesting mature continental provenance (6). Much thicker quartz-rich sandstones and conglomerates overlie the Upper Sequence, with unknown degree of conformity.

Petrographic, geochemical, and isotopic analyses of the bimodal metavolcanics of the LMS show these to be basalts, basaltic andesites, trachyandesites (shoshonitic), and rhyolites (6,8). Spilitic alteration is locally apparent, but the coherence of alkali element ratios and readily-altered trace element compositions suggests that most samples did not undergo strong alteration. Good correlation between HREES, Ti, and magnesium number in the mafic rocks demonstrate the effects of fractional crystallization in the mafic rocks. LREES, Si, K, Rb, Cs, and Ba do not correlate with magnesium number, suggesting that variable enrichments of these elements (fig. 2) reflect variable contamination of the basaltic melts with crustal material. Several contamination components must have been involved, since these elements are only weakly correlated among themselves, and with U, Th, Nb, and Ta. Rhyolite patterns show significant negative Eu anomalies.

Zircons from two quartz porphyritic rhyolites give an age of 2758 + 39 Ma (7), the best estimate of the age of eruption of the LMS. Rb-Sr whole-rock analyses of mafic rocks yield an isochron of 2687 + 54 Ma, similar within the range of calculated errors of the zircon age. Thus the GPG's Late Archean age is well established. The high initial Sr isotopic ratio 0.7057 for the mafic rock isochron is significantly higher than values of CHUR (0.7012, 9) or depleted mantle (0.7008, 10) for 2758 Ma. This indicates contamination by older continental crust. Sm-Nd results are too restricted in distribution to yield a usable isochron.

e Sr vs. e Nd values (Fig. 3) show a cluster around e Sr +50 and e Nd +3. These indicate that the magma was more likely derived from a depleted source than from a CHUR-like source. The high e Sr values are probably either the results of seawater interaction, leaving Nd isotope ratios intact; or contamination with older, presumably mafic crust that had elevated Rb/Sr ratios, but mantle-like Sm/Nd ratios. One rhyolite has similar e Nd and e Sr values, suggesting derivation from similar sources by similar processes. Three of the mafic samples have negative e Nd and positive e Sr values, possibly indicating contamination by older granulitic and granitoid crust. Note that the ranges of diversity in the e Sr and e Nd data can be seen in the basalts alone: the isotopic variation does not correlate directly with silica content. Diverse sources of contamination are indicated, and might be found

in the diverse lithologies of the underlying Xingu Complex.

The geochemical data indicate that the GPG has many features in common with ancient and modern volcanic suites erupted through continental crust. The mafic rocks clearly differ from those of most Archean greenstone belts, and modern MORB, IAB, and hot-spot basalts. The geological, geochemical, and isotopic data are all consistent with depostion on continental crust, presumably in a marine basin formed by crustal extension. The isotopic data also suggest the existence of depleted mantle as a source for the parent magmas of the GPG. The overall results suggest a tectonic environment, igneous sources, and petrogenesis similar to many modern continental extensional basins, in contrast to most Archean greenstone belts. The Hammersley basin in Australia and the circum-Superior belts in Canada may be suitable Archean and Proterozoic analogues, respectively.

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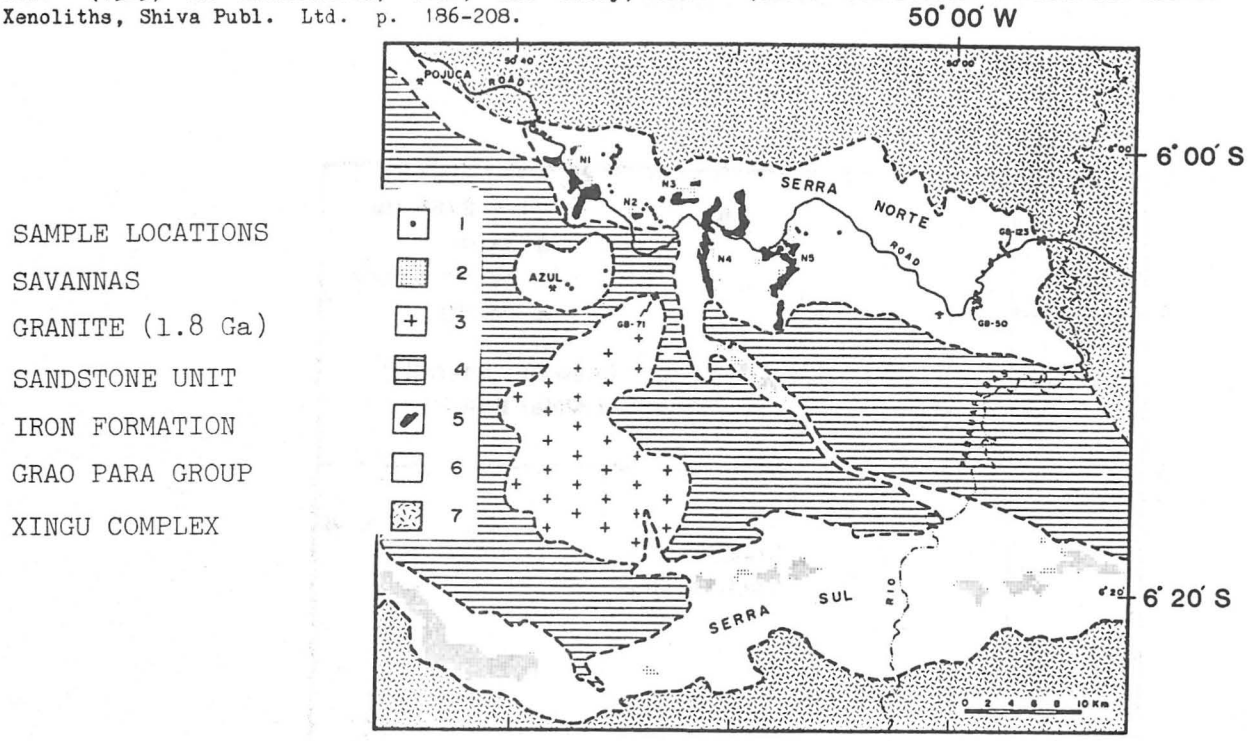
REFERENCES - [1] Hirata, W.K., Rigon, J.C., Kadekaru, K., Cordeiro, A.A.C., and Meireles, E.A. (1982) Anais Simp. Geol. Amazonia I, 100-110. [2] Santos, J.O.S., and S.O.C. Loguercio (1984) in Schobbenhaus et al. (eds.) Geologia do Brasil, DNPM, Brasilia, p. 93-127. [3] Amaral, G. (1984) in F.F.M. de Almeida and Y. Hasui, O Precambriano do Brasil, Edgard Blucher, Sao Paulo, p. 6-35. [4] Gibbs, A.K., Wirth, K.R., and Hirata, W.K., (1985) Field report on a visit to the Grao Para Group, Serra dos Carajas, Brazil (unpublished), 35 p. [5] Cordani, U.G., Tassinari, C.C.G., and Kawashita, K. (1984) Ciencias da Terra, 9:6-11. [6] Gibbs, A.K., Wirth, K.R., Hirata, W.K., and Olszewski, Wm. J. Jr. (1985 submitted) II Simposio Geol. Amazonia, Soc. Bras. Geol. Nucl. Norte, Dec. 1-9, 1985, Belem. 17 p. [7] Wirth, K.R., Gibbs, A.K., and Olszewski, Wm. J. Jr. (1985 submitted) II Simp. Geol. Amazonia, Soc. Bras. Geol. Nucl. Norte, Dec. 1-9, 1985, Belem. 12 p. [8] Lemos, V.P., and Villas, R.N. (1983) Revista Bras. Geocien. 13:165-177. [9] DePaolo, D.J., and Wasserburg, G.J. (1979) Geochim. Cosmochim. Acta 43:6156-617. [10] McCulloch, M.T., and Chappell, B.W. (1982) Earth Planet. Sci. Lett. 58:51-64. [11] Cox, K.G. (1983) in Hawkesworth, C.L., and Norry, M.J. (eds.) Continental Basalts and Mantle Xenoliths, Shiva Publ. Ltd. p. 139-157. [12] Thirlwall, M.F., and Jones, N.W. (1983) in Hawkesworth, C.L., and Norry, M.J. (eds.) Continental Basalts and Mantle Xenoliths, Shiva Publ. Ltd. p. 186-208.



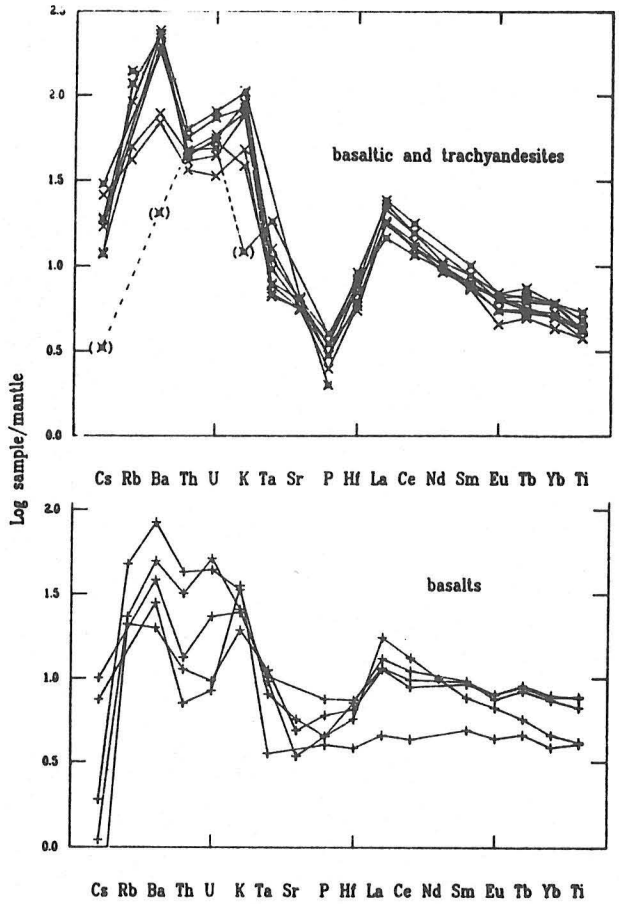


Fig. 2: Incompatable element diagrams for mafic rocks, normalized for model primitive mantle of Wood et al. 1979, and Hf, Eu, and Yb interpolated from chondrite data. Karoo basalts have similar enrichment patterns.

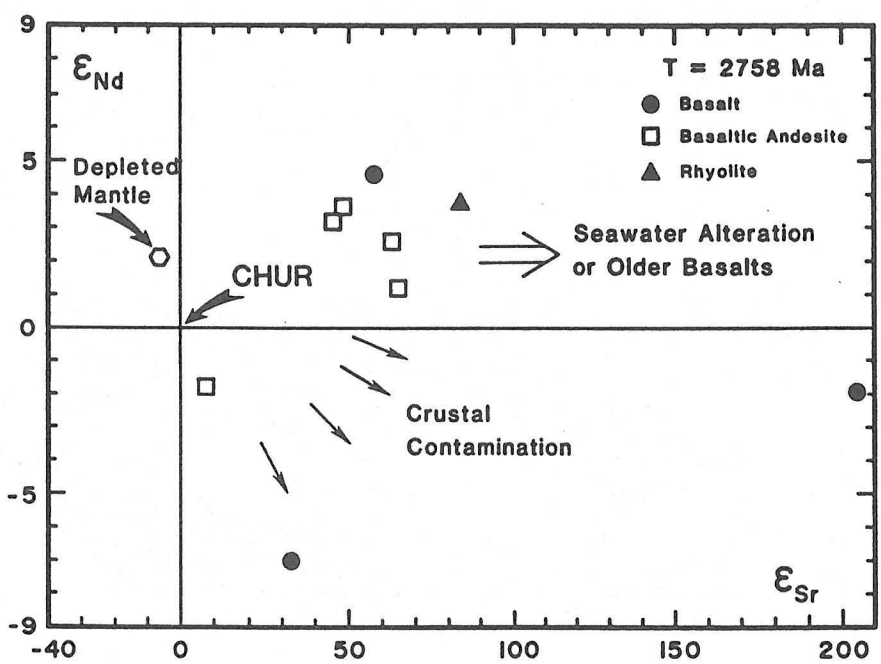


Fig. 3: Sr - Nd diagram for Grao Para Group metavolcanic rocks.

POLYPHASE THRUST TECTONICS IN THE BARBERTON GREENSTONE BELT.

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In the circa 3.5 by old Barberton greenstone belt, the supracrustal rocks form a thick and strongly deformed thrust complex. Structural studies in the southern part of the belt have shown that 2 separate phases of over-thrusting (D_1 and D_2) successively dismembered the original stratigraphy. Thrust nappes were subsequently refolded during later deformations (D_3 and D_4). This poster deals with the second thrusting event which, in the study region appears to be dominant, and (unlike the earlier thrusting), affects the entire supracrustal pile.

The supracrustal rocks form a predominantly NE/SW oriented, SE dipping tectonic fan (the D_2 fan) in which tectonic slices of ophiolitic-like rocks are interleaved with younger sedimentary sequences of the Diepgezet and Malalotcha Groups[1] (Fig. 1). Two distinct levels of decollement can be distinguished within this fan: (1) Within the ophiolitic sequence, usually below the pillow lavas. These zones are delineated by strongly sheared serpentinite lenses and talcose schists. Asbestos fiber is commonly developed in such sheared lenses, as for example in the Havelock and the Msauli asbestos deposits. (2) At the base of the Diepgezet Group, within ferruginous shales and banded cherts. This upper decollement zone is not always obviously sheared, but it is ubiquitously folded in a disharmonic manner and is thought to have been gravity induced, on a dynamic slope, during sedimentation, because: (1) The finely laminated rocks at this stratigraphic level are conformably to unconformably overlain by a 2 to 3 km thick medium to coarse grained clastic sequence (the rest of the Diepgezet Group and the Malalotcha Group; the Malalotcha Group is derived from a quartz-rich source and from the reworking of folded Diepgezet Material). (2) Within the D_2 fan, individual tectonic units may be folded independantly of one another (Fig. 2). The D_2 folds are mostly isoclinal, with fold axes broadly parallel to the thrust contacts (Fig. 2), and are contemporaneous with the emplacement of the nappes. Another set of D_2 folds is contemporaneous with the deposition of the Malalotcha Group sediments and probably formed in tectonically ponded basins, during periods of thrust propagation along the lower decollement level.

Structural and sedimentological data indicate that the D_2 tectonic fan was formed during a prolonged, multi-stage regional horizontal shortening event during which several types of internal deformation mechanisms were successively and/or simultaneously active. Movement appears to have been predominantly to the NW and to the N. During D_2, periods of quiescence and sedimentation followed periods of thrust propagation. Although the exact kimematics which led to the formation of this fan is not yet known, paleoenvironmental interpretations together with structural data suggest that D_2 was probably related to (an) Archean collision(s).

References

(1) Lamb, S. (1984) Structures and sedimentology on the eastern margin of the Archean Barberton greenstone belt, northwest Swaziland, Ph.D Thesis, University of Cambridge, UK. (2) Paris, I.A. (1985) The geology of the Simbubule area, Barberton. Ph.D Thesis, University of the Witwatersrand, Johannesburg, South Africa.

THRUST TECTONICS IN THE BARBERTON BELT

Paris, I.

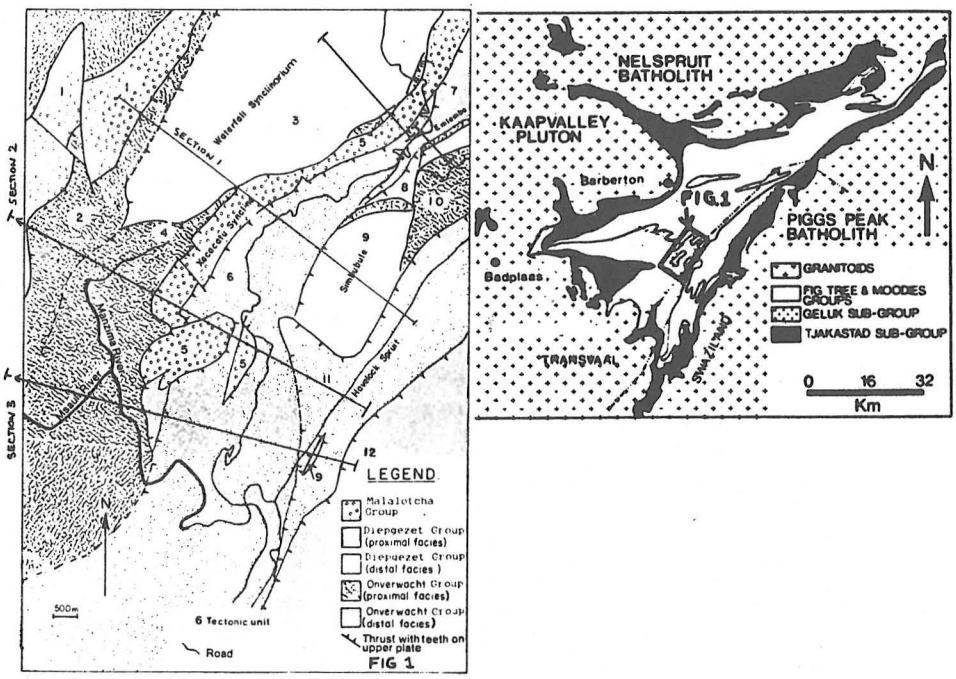

FIG 1

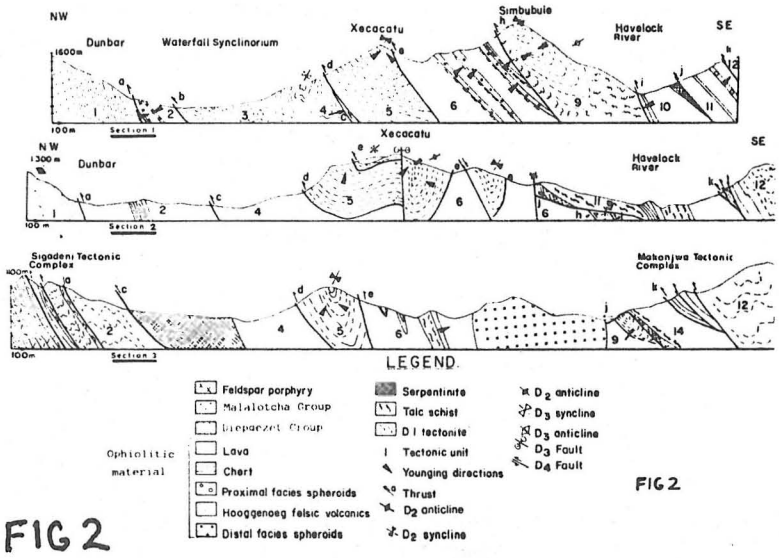

FIG 2

Figure Caption: (1) Simplified geological map of part of the investigated area (for location see inset). (2) Three sections, as located on Fig. 1 showing part of the thrust complex. Note how some of the thrust-slices (individually numbered) composed of sediments (Diepgezet and Malalotcha Groups) are tight to isoclinally folded. Folding and thrusting are related to the same regional deformation (D_2).

AGE CONSTRAINTS ON THE EVOLUTION OF THE QUETICO BELT, SUPERIOR PROVINCE, ONTARIO; J.A. Percival and R.W. Sullivan, Geological Survey of Canada, 601 Booth Street, Ottawa, Ontario K1A 0E8

Much attention has been focussed on the nature of Archean tectonic processes and the extent to which they were different from modern rigid-plate tectonics. The Archean Superior Province (1) has linear metavolcanic and metasediment-dominated subprovinces of similar scale to Cenozoic island arc-trench systems of the western Pacific (2), suggesting an origin by accreting arcs (3,4). Models for the evolution of metavolcanic belts in parts of the Superior Province suggest an arc setting (4,5) but the tectonic environment and evolution of the intervening metasedimentary belts are poorly understood. In addition to explaining the setting giving rise to a linear sedimentary basin, models must account for subsequent shortening and high-temperature, low-pressure metamorphism (6-8). Correlation of rock units and events in adjacent metavolcanic and metasedimentary belts is a first step toward understanding large-scale crustal interaction. To this end, zircon geochronology has been applied to metavolcanic belts of the western Superior Province (9-13); this study reports new age data for the Quetico metasedimentary belt, permitting correlation with the adjacent Wabigoon and Wawa metavolcanic subprovinces.

The 10-100 km-wide Quetico belt extends at least 1200 km from beneath cover in the west to the Kapuskasing structure and probably continues 800 km further east, as the Opatica belt. It is mainly fault-bounded against adjacent metavolcanic rocks but stratigraphic contacts are present locally. The belt consists of marginal zones of metasedimentary schist and an interior zone of migmatite and granite. Marginal metasediments have preserved sedimentary structures suggesting a homogeneous sequence of turbiditic greywacke, possibly derived from adjacent volcanic highlands (14). Conglomerate and cross-bedded sandstone of the Seine Group (15) occur sporadically along the northern margin of the belt and have been interpreted as proximal fan deposits of the Quetico turbidites (16) or as a younger sequence (15,17).

The most prominent structural features of the belt are the regular east-trending bedding which dips steeply near the margins and moderately in the interior, and a pervasive, gently east-plunging lineation. Several early sets of folds have been recognized in detailed studies (18-20). Symmetrical low-pressure metamorphic zonation characterizes marginal schists, where grade increases from chlorite-muscovite at the margins, through biotite, staurolite, and garnet-andalusite zones, to garnet-cordierite-sillimanite grade adjacent to the interior zone of migmatite and intrusive granite. Common assemblages of garnet-andalusite throughout marginal schists and locally in the interior indicate low metamorphic pressure (bathozone 2; 3.3 kbar (21)). Granulite facies occurs in the east near Flanders Lake (22) and adjacent to the Kapuskasing zone (23), where metamorphic pressure is 4-6 kbar (24). The regional metamorphic culmination is coincident with interior plutons, suggesting that the granites transmitted heat to high levels in the crust.

Plutonic rocks, classified into three compositional groups, have restricted spatial distribution: 1) a suite of small diorite-monzonite plugs cuts marginal schists and extends locally into adjacent metavolcanic belts; 2) biotite-magnetite leucogranite with local tonalite and amphibolite inclusions, occurs near the schist-migmatite contact; and 3) peraluminous granite, with garnet, cordierite, muscovite, sillimanite, apatite and tourmaline, are prevalent in the interior zone, particularly the Sturgeon Lake batholith (8). Late pegmatites are ubiquitous in the interior zone and common in the higher-grade parts of the marginal schist unit.

QUETICO BELT, SUPERIOR PROVINCE
Percival, J.A. and Sullivan, R.W.

U-Pb zircon geochronology in the Wawa subprovince indicates major volcanic activity between 2749 and 2696 Ma (25) followed by D_1 deformation at about 2696, deposition of alkaline ("Timiskaming") volcanics at 2689, D_2 deformation, and intrusion of post-tectonic plutons at 2684 Ma (9) to 2668 Ma (26) (Fig. 1). In the Wabigoon subprovince, volcanics were erupted in the interval 2755-2702 Ma, with post-tectonic plutons younger than 2695 Ma (12) (Fig. 1).

A chilled porphyritic dacite sill cutting biotite-grade Quetico metasediments yielded an imprecise U-Pb zircon date of 2743 ± 16 Ma, providing a minimum age for sediment deposition. A single tonalite clast from metaconglomerate at Max Creek, interpreted to be Seine equivalent, has zircons dated at 2684 ± 10 Ma, interpreted as the age of the source pluton. Together these dates show that the Quetico metasediments and Seine Group are not facies equivalent. Monazites from the geologically oldest plutonic rock type, a foliated biotite granite with zircons with relict cores, are discordant, with an upper intercept of 2684 Ma. Monazite from massive peraluminous granite with probable inherited zircon is concordant at 2670 Ma. Zircon and monazite from a pegmatite dyke form a discordia line with an upper intercept of 2671 Ma (Fig. 1). The data do not permit definition of the length of time of sediment deposition nor is the thickness of the sequence known; thus inferences on lithospheric thickness (28) cannot be made.

Preliminary synthesis suggests that sediment deposition on extending crust forming the Quetico basin probably occurred during volcanism in adjacent terranes, possibly continuing until volcanism ceased. Closure of the basin during D_1 and/or D_2 events, dated in adjacent belts, led to folding of the sedimentary pile and thickening of the weak crust. Conglomerates were deposited adjacent to marginal transcurrent faults. During subsequent thermal relaxation, partial melts were extracted from lower crustal metasedimentary and tonalitic rocks in a crustal root zone as well as from the mantle. The derived granites and diorites ascended passively to within 10 km of the surface, producing a regional low-pressure aureole in the host schists. A back-arc or inter-arc setting is favoured over an accretionary prism environment for the Quetico sediments because of its symmetry and high-temperature metamorphism which probably occurred in a region of high heat flow.

REFERENCES

1) Goodwin, A.M. et al. (1972) Geol. Assoc. Can. Sp. Pap. 11, 527-623.
2) Hamilton, W. (1979) U.S. Geol. Surv. Prof. Pap. 1078, 345 p.
3) Langford, F.F. and Morin, J.A. (1976) Am. J. Sci. 276, 1023-1034.
4) Blackburn, C.E. (1980) Geoscience Canada 7, 64-72.
5) Blackburn, C.E. et al. (1985) Geol. Assoc. Can. Sp. Pap. 28, 89-116.
6) Pirie, J. and Mackasey, W.D. (1978) Geol. Surv. Can. Pap. 78-10, 37-48.
7) Thurston, P.C. and Breaks, F.W. (1978) ibid, 49-62.
8) Percival, J.A. et al. (1985) Geol. Surv. Can. Pap. 85-1A, 385-397.
9) Corfu, F. and Stott, G.M. (1985) Inst. Lake Superior Geol. 32 (abstr.), 15-16.
10) Turek, A. et al. (1984) Can. J. Earth Sci. 21, 457-464.
11) Davis, D.W. and Edwards, G. (1982) ibid 19, 1235-1245.
12) Davis, D.W. et al. (1982) ibid 19, 254-266.
13) Nunes, P.D. and Thurston, P.C. (1980) ibid 17, 710-721.
14) Ojakangas, R.W. (1985) Geol. Assoc. Can. Sp. Pap. 28, 23-47.
15) Lawson, A.C. (1913) Geol. Surv. Can. Mem. 40, 111 p.
16) Wood, J. (1980) Precambrian Res. 12, 227-255.
17) Poulsen, K.H. et al. (1980) Can. J. Earth Sci. 17, 1358-1369.

18) Kehlenbeck, M.M. (1984) Geoscience Can. 11, 23-32.
19) Borradaile, G.J. (1982) Precamb. Res. 19, 179-189.
20) Sawyer, E.W. (1983) ibid 22, 271-294.
21) Carmichael, D.M. (1978) Am. J. Sci. 278, 769-797.
22) Coates, M.E. (1968) Ont. Dep. Mines Geol. Rep. 68, 22 p.
23) Percival, J.A. (1985) Geol. Surv. Can. Pap. 85-1A, 1-5.
24) Percival, J.A. (1985) Geol. Assoc. Can. Abstr. Prog. 10, A48.
25) Turek, A. et al. (1982) Can. J. Earth Sci. 19, 1608-1626.
26) Krogh, T.E. and Turek, A. (1982) ibid 19, 859-867.
27) Davis, D.W. (1985) Personal communication.
28) McKenzie, D. et al. (1980) Earth Planet. Sci. Lett. 48, 35-41.

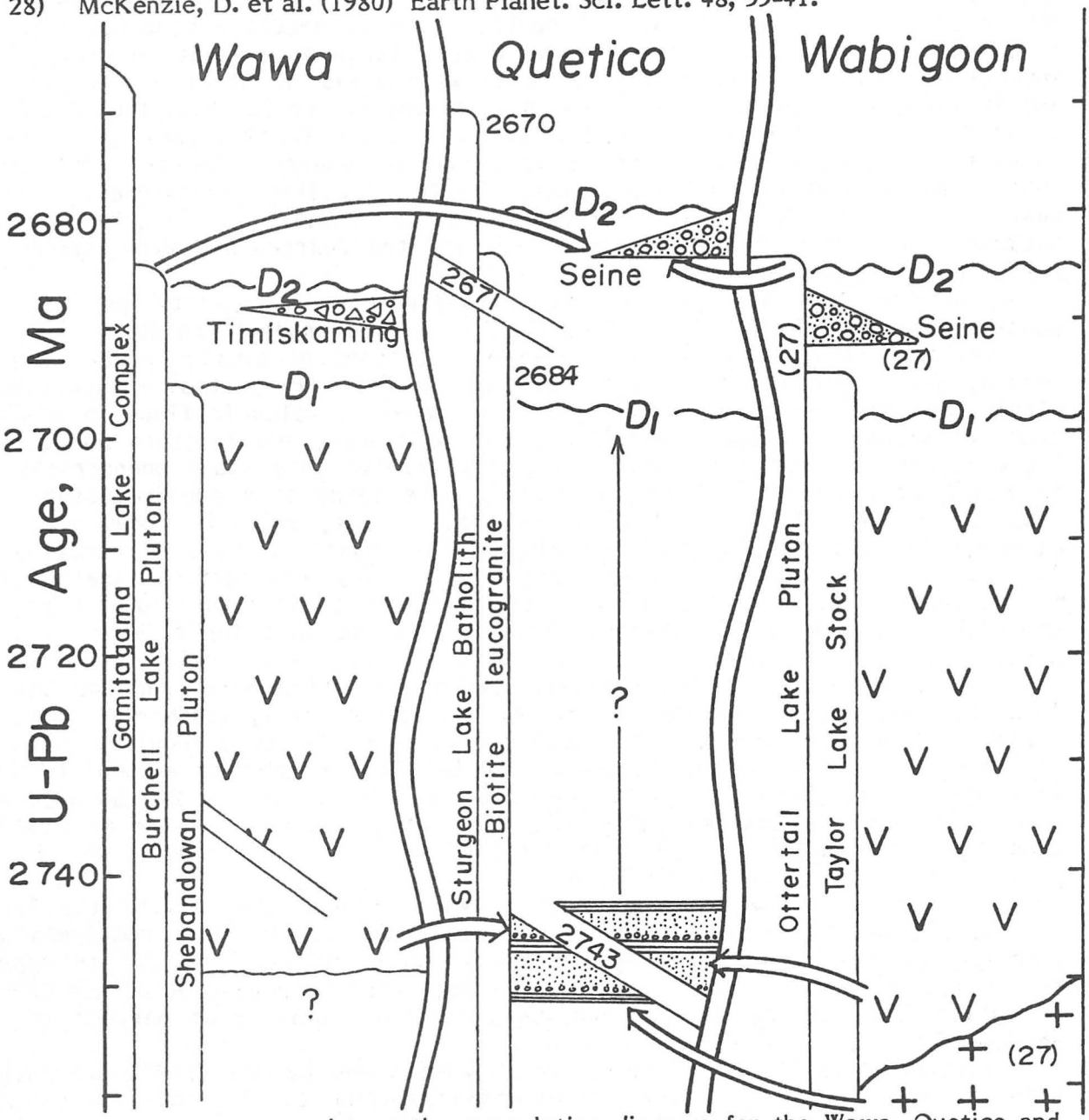

Fig. 1: Age summary and tentative correlation diagram for the Wawa, Quetico and Wabigoon subprovinces. Arrows crossing subprovince boundaries indicate sedimentary provenance.

ARCHEAN MEGACRYSTIC PLAGIOCLASE UNITS AND THE TECTONIC SETTING OF GREENSTONES

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Large (up to 20 cm), equidimensional, commonly euhedral, plagioclase megacrysts of highly calcic composition (An_{80-90}) occur commonly in all Archean cratons in one or more of three distinct associations:
1) as cumulate crystal segregations of anorthosite or as megacrysts in basaltic dikes, sills, and flows in greenstone belts that vary in metamorphic grade from greenschist to granulite. Throughout 100's of thousands of square kilometers of northwestern Ontario and Manitoba the plagioclase megacrysts occur in pillowed and massive flows, sills, dikes, large inclusions in dikes, and intrusive anorthositic complexes (Fig. 1) with areas of up to a few 100 km^2 and spanning a period of at least 100 m.y. in the 2.7 to 2.8 b.y. time frame,
2) as basaltic dike swarms in stable cratonic areas forming parallel to subparallel patterns over hundreds of thousands of square kilometers intruding both granitic gneisses and supracrustal belts including greenstones. These swarms include the Ameralik-Saglek system at 3.1 to 3.4 b.y. (Fig. 2) [1], the Matachewan system at 2.5 to 2.6 b.y. [2], and the Beartooth-Bighorn system at 2.2 to 2.3 b.y. [3], and
3) as anorthositic complexes associated with marbles and quartzites (Sittampundi, India and Messina, South Africa) in granulite grade terrains.

Initial attempts to correlate tectonic settings of similar modern cryst-bearing units with their Archean counterparts were only partially successful. Plagioclase phenocrysts of An_{80-90} occur in basaltic volcanic flows in oceanic crust at spreading ridges, hotspots, aseismic ridges, and fracture zones [4]. These recent occurrences, however, normally involve only small phenocrysts up to a few millimeters in size and usually more lathy than equidimensional in shape [5]. In contrast to these normal occurrences, volcanic flows over the Galapagos hotspot display more equidimensional crysts up to 3 cm across [4]. Although these oceanic environments might be satisfactory tectonic analogs for many greenstone occurrences, they certainly are not satisfactory for the extensive dike swarms in stable cratonic masses. Thus we turn for clues to a more detailed understanding of the petrogenesis of the crysts and related melts.

The crysts are quite homogeneous, varying by little more than one to two An units over several centimeters thereby suggesting nearly isothermal crystallization at nearly constant melt composition over the time required to grow crystals commonly 6 to 8 cm across and up to 20 cm across and accumulate them in large masses. Thin, more sodic rims on the order of 100 to 200 μm wide are common on large crysts when the groundmass plagioclase laths are more sodic than the large crysts. The rims normally approach the composition of the plagioclase in the groundmass (Table 1).

The nature of the parent melts, or melts in equilibrium with the large crysts, has been an open question because: 1) the anorthositic complexes are clearly cumulates with bulk compositions too rich in Al_2O_3 and CaO to represent melts [6], and 2) the disparity in composition between plagioclase crysts and plagioclase of the matrix suggests a lack of equilibrium between crysts and the melt represented by their matrix.

Initial attempts to determine melt compositions by use of REE concentrations in megacrysts in conjunction with distribution coefficients for plagioclase and basaltic melts were fraught with problems resulting from modification of plagioclase REE concentrations by alteration, recrystallization, and tiny inclusions. By utilizing several splits from each cryst in several samples from the BVL anorthosite, mixing lines were determined and the least

modified REE concentrations were calculated for pristine plagioclases [7]. These values in conjunction with the most recent distribution coefficients indicate melts with nearly flat REE patterns at 10X to 20X chondrites with perhaps a slight depletion in the light REE's. The calculated patterns compare well with several cryst-bearing basalts in greenstone belts (Fig. 3) as well as with the non-cryst-bearing basalts. These patterns are those of the least enriched tholeiitic basalts which are very common in greenstone belts. Comparison of these basalts with those in the cratonic dike swarms shows many similarities (Fig. 3, Table 2) but the initial data suggests that the cratonic dikes are slightly enriched in SiO_2, K_2O, and light REE. It is tempting to attribute these differences to contamination of the melts as they rise through continental crust but the melts of the Galapagos when compared with MORB show some of the same enrichments (Table 2) which in this case cannot be attributed to continental contaminants. Further work on the pristine REE contents of plagioclase megacrysts is underway and should help determine whether megacrysts in enriched melts formed from the more enriched or less enriched tholeiitic melts, or both.

At present the petrogenetic data require, at a minimum, isothermal crystallization of plagioclase megacrysts from tholeiitic melts (the least enriched ones in greenstone belts) followed by segregation of the plagioclase crystals which then become entrained in rising melts to form intrusions or volcanic flows. Furthermore, the occurrences seem to require large volumes of melt at similar temperatures for long periods of time over huge areas having both oceanic and cratonic associations. Continual generation of similar melt and continuous addition of the melt to extensive networks of crystallizing chambers is also strongly implied. The major remaining questions with significant implications for the setting and evolution of greenstone belts are: 1) Does the cryst-producing melt have the same composition and crystallize under the same conditions beneath greenstone belts, stable cratons, and current oceanic crust? 2) Where do the plagioclase crysts form and accumulate; in low or high pressure environments? 3) Is there a systematic change in the time of megacryst emplacement across large areas such as might be produced by plates overriding zones of melt production or other such time-dependent mechanisms?

Refs: [1] McGregor V. R. et al. (1985) Workshop: The Worlds Oldest Rocks, Lunar and Planetary Inst., Houston. [2] Ernst R. E. (1982) Ont. Geol. Surv. Misc. Pap. 106, p. 53-56. [3] Miller J. D. (1980-81) Wyo. Geol. Assoc. Earth Sci. Bull., v. 13-14, p. 187-215. [4] Cullen A. et al. (1985) preprint from Center for Volc. Univ. Oreg. [5] Blanchard D. P. et al. (1976) Jour. Geophys. Res., 81, p. 4231-4246 and Donaldson C. (1977) EPSL, 37, p. 81-89. [6] Phinney W. C. (1982) LPI Tech. Rpt. 82-01, Lunar and Planetary Inst., Houston, p. 121-124. [7] Morrison D. A. et al. (1985) Lunar and Planet. Sci. Conf. XVI, Lunar and Planetary Inst., Houston, p. 589-590.

Table 1

Matachewan Dikes: Plag. Compositions

LOCATION	CRYSTS			MATRIX LATHS	
	main cryst	inner rim	outer rim	core	rim
Sta 8	An79-81	An56	An42-49	An61-64	An47
9-11	80-82	66-69	54	65-69	54
10	75-78	61-62	54	58-61	45-50
16	81-84		62-67	68-70	52-53
16	82-84		64-65	64-66	46-53
16	83-85		72-76	71-72	66-69
18	Too Altered			50-54	33
34	81-83			No laths	
35	81-83		78-80	61-63	51-54

Table 2. Matrices of Cryst-Bearing Tholeiites

	RECENT		ARCHEAN	
	MORB Leg 37	Pinta Galapagos	BRNIB Bird River Greenstone	KD4-6 Cratonic Dike
SiO_2	48.36 %	48.60	48.85	50.09
TiO_2	0.89	2.10	0.96	0.94
Al_2O_3	16.85	16.68	15.00	14.67
Fe_2O_3	3.39	6.52	2.14	2.37
FeO	5.41	3.95	10.14	9.74
MnO	0.17	0.17	0.22	0.19
MgO	7.21	7.23	6.50	6.43
CaO	12.53	12.15	12.07	10.54
Na_2O	2.15	2.80	1.85	2.32
K_2O	0.19	.43	0.29	0.61
P_2O_5	0.08	.31	0.09	0.10
H_2O	2.35	.47	1.76	1.93
La	4.41 ppm	13.00	4.04	9.64
Ce	10.5	36.20	10.12	19.6
Sm	2.11	5.49	2.42	2.96
Eu	0.80	2.02	.895	0.95
Yb	2.19	3.40	2.41	2.414
Lu	.38	.40	.377	0.408

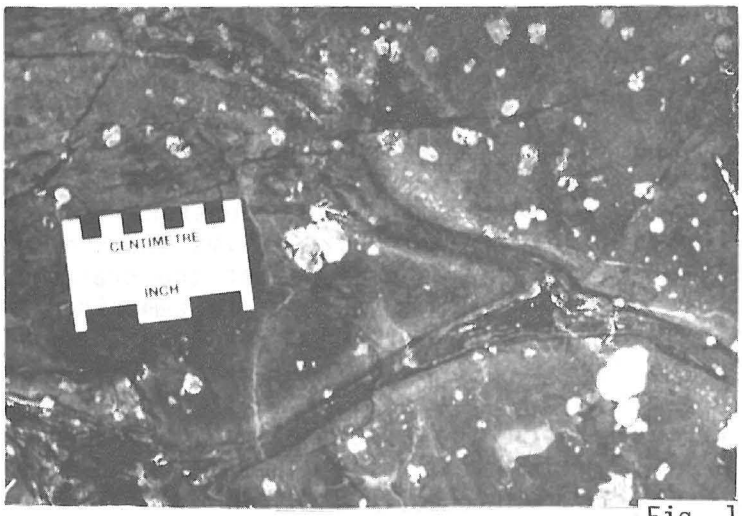

Fig. 1

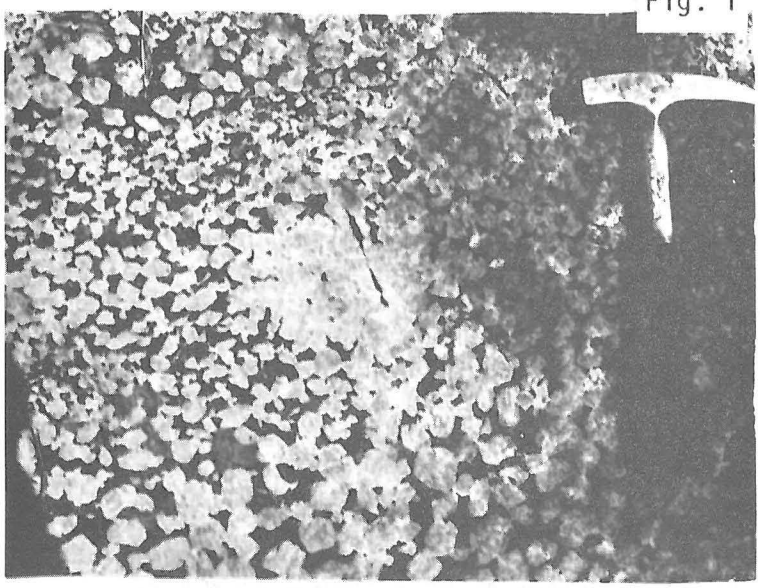

Fig. 2

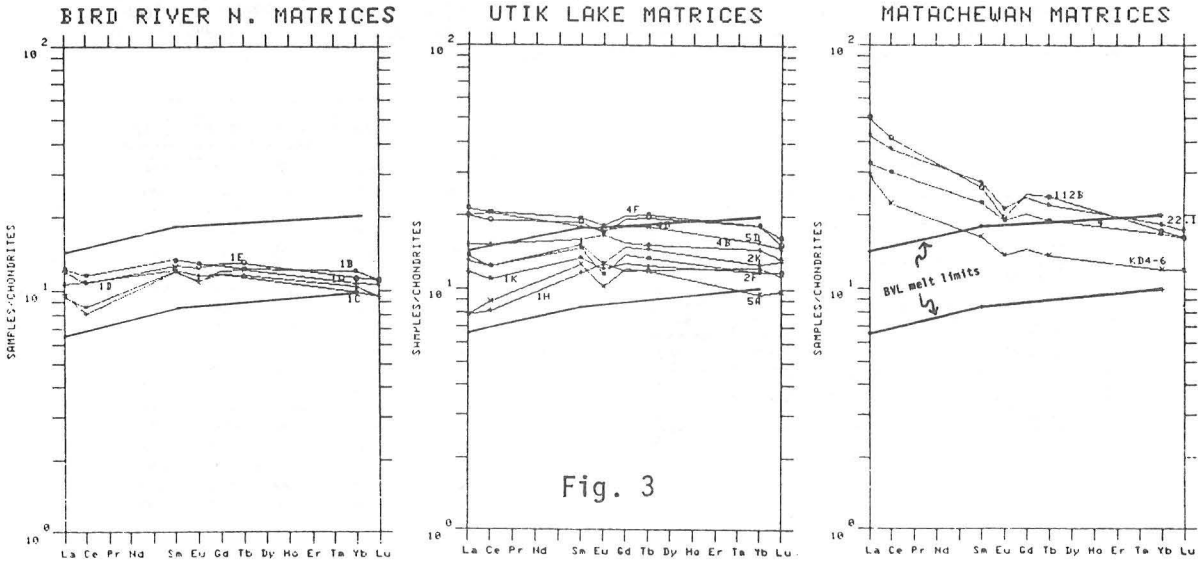

Fig. 3

RAINY LAKE WRENCH ZONE: AN EXAMPLE OF AN ARCHEAN SUBPROVINCE BOUNDARY IN NORTHWESTERN ONTARIO; K. Howard Poulsen, Economic Geology and Mineralogy Division, Geological Survey of Canada

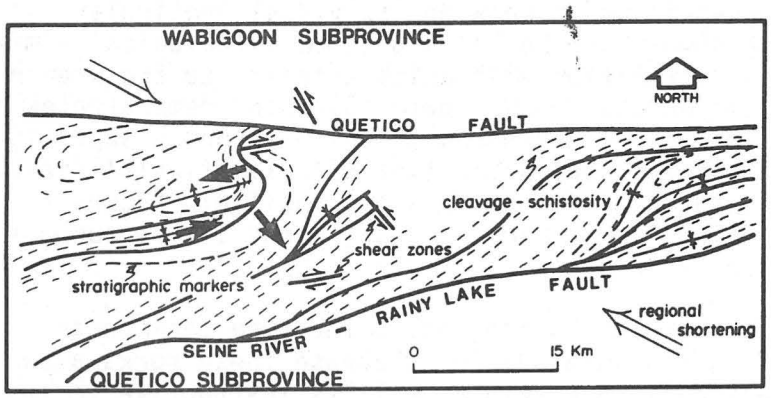

Fig. 1- Schematic diagram illustrating structural features of Rainy Lake Wrench Zone. Short solid arrows identify downward facing units.

The Superior Province of the Canadian Shield comprises an alternation of subprovinces with contrasting lithological, structural and metamorphic styles (1). Rocks of the Rainy Lake area form a fault bounded wedge between two of these subprovinces, the Wabigoon granite-greenstone terrain to the north and the Quetico metasedimentary terrain to the south (Fig. 1). The Quetico and Seine River-Rainy Lake Faults bound this wedge within which interpretation of the stratigraphy has been historically contentious. In the eastern part of the wedge, volcanic rocks and coeval tonalitic sills are unconformably overlain by fluviatile conglomerate and arenite of the Seine Group; in the western part of the wedge, metamorphosed wacke and mudstone of the Coutchiching Group are cut by granodioritic plutons. The Coutchiching Group has previously been correlated with the Seine Group and with the turbiditic Quetico metasediments of the Quetico Subprovince and these correlations are the cornerstone of earlier tectonic models which relate the subprovinces (2,3).

The structural geology of the Rainy Lake area is characterized by the following attributes:

(i) lenticular lithostratigraphic domains with discordant boundaries,
(ii) steep boundary faults,
(iii) regular orientation and sense of displacement of small ductile shear zones,
(iv) regionally developed sub-vertical foliation which transects large lithological folds,
(v) shallow bimodal orientations of minor folds and lineations and a preponderance of folds of dextral asymmetry,
(vi) downward facing folds in the Rice Bay, Nickel Lake and Bear Pass areas (arrowed, Fig. 1).

These observations compare favourably with the known characteristics of dextral wrench or "transpressive" zones based both on experimental data and natural examples (4,5,6,7,8). Much of this deformation involved the Seine Group, the youngest stratigraphic unit in the area (9), and predates the emplacement of late-to-post-tectonic granodioritic plutons for which radiometric data indicate a Late Archean age.

The interpretation of a wrench zone separating the Wabigoon and Quetico Subprovinces has important implications regarding the tectonic models which can

be used to relate them. Of great importance is the high probability that this zone contains rocks which are actually allochthonous relative to those adjacent in the Quetico and Wabigoon. Given this type of structural environment, not only is correlation of stratigraphic units between individual lenticular domains difficult to establish simply on the basis of some lithological similarity but more important, the correlation with units exterior to the wrench zone is even more suspect. New geochronological data (9) which demonstrates a 40 Ma difference in age between the Seine and Coutchiching strongly supports this argument. Therefore the concept that Seine-type alluvial-fluvial rocks, which are restricted spatially to the wrench zone are transitial "facies" between Wabigoon volcanics and Quetico turbidites (2,3) finds little support in a wrench zone interpretation.

Pettijohn (10) was the first to emphasize that Seine-type sedimentary sequences occur all along the subprovince margin. Because these rocks also correlate spatially with a well defined wrench zone it is instructive to inquire whether an alternate hypothesis might account for these observed relationships without relying on the concept of facies equivalence. The link between alluvial-fluvial sedimentation and wrench zones is well-known in Cenozoic environments where thick alluvial, fluvial and lacustrine sequences are restricted to narrow "pull-apart" basins associated with large transcurrent faults (11,12,13). Such basins are localized by bends in marginal faults and by intersections with fault splays. Lateral and vertical facies variations are present within such basins (14) but these rocks are not contiguous with rocks external to the basin. The size and geometry of the wrench fault system at the southern margin of Wabigoon subprovince and the areal extent of the Seine-type rocks are comparable with younger examples in which there is also a juxtaposition of rocks of differing lithology. In many of these examples, and possibly in the present one as well, the juxtaposed terranes have depositional histories which are quite independent so that present geographic geometry has no simple paleographic significance.

The proposal that wrench faulting is significant at the subprovince boundary is not a new one. Hawley (15) first suggested a model of this type for rocks in the Atikokan area to the east of Rainy Lake but the emphasis in the past has been placed only on the late-stage displacements on the Quetico Fault (2,16) rather than the possibility presented here that such late faulting is merely a reflection of a broader zone of wrenching which also became a locus for sedimentation.

References

(1) Stockwell, C. H. (1964) Geol. Surv. Can. Paper 65-17, 29p.
(2) MacKasey, W. D., Blackburn, C. E. and Trowell, N.F. (1974), Ont. Div. Mines, MP 58, 30p.
(3) Goodwin, A. M. (1977) Can. J. Earth Sci., 14, p. 2737-2759.
(4) Wilcox, R. E., Harding, T. P. and Seely, D. R. (1973) AAPG Bull., 57, p. 74-96.
(5) Tchalenko, J. S. (1970) Geol. Soc. Am. Bull., 81, p. 1625-1640.
(6) Lowell, J. D. (1972) Geol. Soc. Am. Bull., 83, p. 3091-3102.
(7) Freund, R. (1970) Geol. Soc. Am. Bull., 78, p. 188-200.
(8) Ramsay, J. G. (1980) J. Struct. Geol., 2, p. 83-99.
(9) Davis, D. W. and Corfu, F. (this volume)
(10) Pettijohn, F. J. (1943) Geol. Soc. Am. Bull., 54, p. 925-972.

(11) Schubert, C. (1980) J. Struct. Geol., 2, p. 463-468.
(12) Norris, R. J. and Carter, R. M. (1982) Tectonophysics, 87, p. 11-24.
(13) Miall, A. D. (ed.) (1981) Geol. Assoc. Can., Sp. Paper 23, 272p.
(14) Mann, P. and Burke, K. (1982) Geol. Soc. Am. Abst., New Orleans, p. 555.
(15) Hawley, J.E. (1930) J. Geol., 38, p. 521-547.
(16) Schwerdtner, W. M., Stone, D., Osadetz, K., Morgan, J. and Stott, G. M., (1979) Can. J. Earth Sci., 16, p. 1965-1977.

A PALAEOMAGNETIC PERSPECTIVE OF PRECAMBRIAN TECTONIC STYLES

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The considerable success derived from palaeomagnetic studies of Phanerozoic rocks with respect to the tectonic styles of continental drift (1) and plate tectonics (2), etc. have not been repeated by the many palaeomagnetic studies of Precambrian rocks. This is undoubtedly related to the vast amount of Precambrian time compared with Phanerozoic time, and the concomitant uncertainties of magnetisation ages and rock ages, yet it is still surprising that there is little evidence of consolidation or even convergence of opinions regarding tectonic styles prevalent during the Precambrian. After all, there are 30 years of research with results covering the major continents for Precambrian times that overlap considerably yet there is no concensus even in the grossest terms. There is good evidence that the usual assumptions employed by palaeomagnetism are valid for the Precambrian which only serves to exacerbate the problem. The existence of magnetic reversals during the Precambrian, for instance, is difficult to explain except in terms of a geomagnetic field that was predominantly dipolar in nature. It is a small concession to extend this notion of the Precambrian geomagnetic field to include its alignment with the Earth's spin axis and the other virtues of an axial geocentric dipole that characterise the recent geomagnetic field. In addition it is not a forceful argument to claim that early studies of Precambrian rocks need to be re-done, since re-studies have often only served to confirm the early works. Therefore we submit that the palaeomagnetic results derived from Precambrian rock units are not easily dismissed. It is simply untenable that the majority of the data are spurious and claims that synopses of Precambrian data are invalid, cannot be sustained in such terms. Such arguements posed against the evidence for continental drift have long been debunked. There are, nevertheless, differing interpretations of Precambrian palaeomagnetic data and it is the purpose of this brief article to address this problem.

Methods that have been used to interpret Precambrian palaeomagnetic data fall into two classes. The first class assumes the existence of a "Pangaea" or some supercontinent and proceeds to use the palaeomagnetic data, *a posteriori*, to support the model. The second class, which we prefer, accepts the palaeomagnetic data at face value (as synthesised by workers closely in touch with the results) and proceeds to view the overall relationships of the data, isolated from preconceived notions. This latter approach has led us to suggest that the present day geographical relationships of continents (from which a reasonable amount of data for the Precambrian are available) yields the more satisfactory comparison. Of course small adjustments of the continents refine this comparison, but overall an excellent agreement in Precambrian pole paths can be realised by leaving the continents in their present locations.

Limitations of the available data in our earlier comparisons (3) restricted the time span of comparisons between different continents to 2300 Ma - 1900 Ma for North America and Africa and 1800 Ma - 1600 Ma for North America, Greenland and Australia. Recently two results have been derived from igeneous rock about 2900 Ma in age, in Australia and Africa. The palaeomagnetic pole positions from these rock units are in close proximity,

suggesting that the present geographic relationship of Australia and Africa is valid for 2900 Ma ago. The pole position from the Millindinna Complex, Australia, dated at 2860±20 Ma is at 11.9°S, 161.3°E, dp=6.8°, dm=8.4°(4), while the pole position from the Usushwana Complex, Africa, dated at 2880 Ma is at 11.6°S, 165.8°, dp=5.1°, dm=7.5°(5). Thus there is evidence that during the Precambrian North America and Australia were in their present relative geographic locations for 1800 Ma-1600 Ma, as were North America and Africa for 2300 Ma-1900 Ma, and now Africa and Australia, at least for 2900 Ma ago.

These observations are not easily reconciled with Phanerozoic palaeomagnetic results as we have already discussed(3), but they are a matter of record and must be explicable. In terms of greenstone terranes it is obvious that tectonic models postulated to explain these observations are paramount in understanding Precambrian geology. What relevance the current geographical relationships of continents have with their Precambrian relationships remains a paradox, but it would seem that the ensialic model for the development of greenstone terranes is favoured by the Precambrian palaeomagnetic data.

(1) Irving, E., 1964. Paleomagnetism and its application to geological and geophysical problems. Wiley, N.Y., pp. 399.
(2) McElhinny, M.W., 1973. Palaeomagnetism and plate tectonics. Cambridge University Press, Cambridge, pp. 358.
(3) Embleton, B.J.J. and Schmidt, P.W., 1979. Nature 282, 705-707.
(4) Schmidt, P.W. and Embleton, B.J.J., 1985. J. Geophy. Res., 90, 2967-2984.
(5) Layer, P.W., Kroner, A. and McWilliams, M.O., 1985, pers. comm.

THE WISCONSIN MAGMATIC TERRANE: AN EARLY PROTEROZOIC GREENSTONE-GRANITE TERRANE FORMED BY PLATE TECTONIC PROCESSES; Klaus J. Schulz, U.S. Geological Survey, Reston, VA 22092 and Gene L. LaBerge, Department of Geology, University of Wisconsin-Oshkosh, Oshkosh, WS 54901 and U.S. Geological Survey

The Wisconsin magmatic terrane (WMT) is an east trending belt of dominantly volcanic-plutonic complexes of Early Proterozoic age (~1850 m.y.) that lies to the south of the Archean rocks and Early Proterozoic epicratonic sequence (Marquette Range Supergroup) in Michigan. It is separated from the epicratonic Marquette Range Supergroup by the high-angle Niagara fault, is bounded on the south, in central Wisconsin, by Archean gneisses, is truncated on the west by rocks of the Midcontinent rift system, and is intruded on the east by the post-orogenic Wolf River batholith.

Although the history of the WMT is complex in detail, integration of recent studies (Sims and others, in press) provides an overview of its nature and evolution. The WMT shows many similarities to Archean greenstone-granite (AGG) terranes (Condie, 1981). In fact, until recent U/Pb zircon dating, considerable controversy existed as to the age of the rocks of the WMT. Insofar as the comparisons between the WMT and AGG terranes are valid, understanding of the tectonics of the WMT may provide important insights into the tectonic processes involved in the evolution of at least some AGG terranes.

As in many AGG terranes, a major portion of the WMT is comprised of volcanic rocks and lesser volcanogenic sediments variably metamorphosed to lower greenschist to amphibolite facies. The supracrustal rocks show a complex stratigraphy with at least three successions distinguished on the basis of differences in composition, metamorphism, and structural fabric (LaBerge and Myers, 1984; Sims and others, in press). The older units are dominantly subaqueous basaltic lavas and consanguineous intrusive rocks which are overlain locally by intermediate to felsic volcanic and volcaniclastic units, some in part subaerial (LaBerge and Myers, 1984). Both bimodal (basalt-rhyolite) and calc-alkaline (basaltic andesite through rhyolite) suites are present with the former hosting volcanogenic massive sulfide deposits (May and Schmidt, 1982). Komatiites have not been recognized within the WMT. The older basaltic units are dominantly tholeiitic in character, show strong to moderate depletion of light REE elements ($[La/Yb]_N$=0.09-0.89) and high-field-strength elements (Hf, Zr, Ta, etc.), and are lithologically and compositionally similar to recent back-arc basin basalts (e.g. Mariana Trough, Wood and others, 1981), island-arc tholeiites (e.g. Scotia arc, Hawkesworth and others, 1977), and some ophiolitic basalts (e.g. Troodos, Kay and Senechal, 1976). The younger calc-alkaline units are enriched in LIL elements ($[La/Yb]_N$=2.5-9.5), are also depleted in high-field-strength elements, and are similar to volcanic sequences found in recent island-arcs (e.g. Sunda Arc, Whitford and others, 1979).

Sedimentary rocks are locally found to overlie and(or) interfinger with the volcanic succession. They include graywacke, argillite, thin iron-formation, chert, and minor conglomerate, some containing granitoid boulders (LaBerge and Myers, 1984).

Intrusive rocks within the WMT appear to have been largely diapirically emplaced and show a temporal progression from gabbro and diorite through tonalite and granite. They range from calcic to calc-alkaline in character, although locally slightly alkaline varieties are also present (Sims and others, 1985). The granitoids show an overall increase from north to south across the terrane in their average K_2O/Na_2O ratios and SiO_2 contents. Gneissic rocks, found in domes and block uplifts, are mostly tonalite to granodiroite and are also calc-alkaline (Sims and others, 1985). Both lithologically and chemically, the WMT granitoids appear similar to those formed at compressional plate-margins (Brown, 1982).

Ultramafic rocks are present in the WMT, particularly along the northern and southern margins. They are mostly serpentinized, but perioditic and pyroxenitic lithologies are recognized. These ultramafic rocks are often spatially associated with gabbroic rocks and were in some cases structurally implaced. The ultramafic-gabbroic bodies are lithologically and chemically similar to recent ophiolitic fragments.

Structure within the WMT is complex and consists regionally of large structural blocks having diversely oriented internal structures that are bounded by ductile deformation zones ("shear zones"; LaBerge and Myers, 1984; Sims and others in press). Within the blocks, the supracrustal rocks show generally steep dips and open to isoclinal folds. The deformation zones bounding the blocks record pronounced flattening in the foliation planes and a strong component of verticle movement (Palmer, 1980). This intense deformation along zones is regional in scope, and generally younger than the pervailing internal structural fabric within the blocks. Domes along the northern margin of the terrane, representing large-scale, antiformal fold-interference structures, modified by diapirism and by intrusion of granitoids, have further deformed and metamorphosed the mantling supracrustal rocks (Sims and others, 1985).

U-Th-Pb zircon ages on the volcanic and associated gneissic and granitoid rocks that comprise the WMT (VanSchmus, 1980; Sims and others, in press) indicate that they formed from 1,890 to 1,830 Ma. Detailed isotopic dating in the northeastern portion of the WMT (Sims and others, 1985) indicates that volcanism, granitoid intrusion, metamorphism, and deformation within this region occurred from 1,865 to 1,835 Ma ago, a time span of 30 m.y.

The overall lithologic, geochemical, metallogenic, metamorphic, and deformational characteristics of the WMT are similar to those observed in recent volcanic arc terranes formed at sites of plate convergence. It is concluded that the WMT represents an evolved oceanic island-arc terrane accreated to the Superior craton in the Early Proterozoic. This conclusion is strengthened by the apparent absence of Archean basement from most of the WMT, and the recent recognition of the passive margin character of the epicratonic Marquette Range Supergroup (Larue and Sloss, 1980). On the basis of the new data for the WMT and the epicratonic sequence in Michigan, Schulz and others (in press) have proposed the following tectonic model: 1) early crustal rifting and spreading along the southern margin of the Superior craton, 2) subsequent subduction and formation of a complex volcanic arc, and, 3) with oblique convergence, collision of the arc with the continental margin (epicratonic) sequence and Archean crust of upper Michigan culminating

in the Penokean orogeny. This tectonic model is similar to plate tectonic histories recently presented for other Early Proterozoic terranes of North America (Hoffman, 1980; Lewry, 1981; Karlstrom and others, 1983). This indicates that the events and processes occurring in the Lake Superior region were not unique, and that the tectonic processes operating were generally similar to those recognized for the Phanerozoic. Given the general similarity of some AGG terranes to the Early Proterozoic magmatic terranes, it seems likely that subduction and plate collisions were also operative in the Archean.

References:

Brown, G. C., 1982, In: Thorpe, R. S., ed., Andesites: New York, John Wiley and Sons, p. 437-461.

Condie, K. C., 1981, Archean Greenstone Belts: Elsevier Scientific Publ. Co., New York, 434 p.

Hawkesworth, C. J., O'Nions, R. K., Pankhurst, R. J., Hamilton, P. J., and Evensen, N. M., 1977, Earth and Planet, Sci. Letters, v. 36, p. 253-262.

Hoffman, P. F., 1980, In: Strangway, D. W., ed., Geol. Assoc. Canada Special Paper 20, p. 523-549.

Karlstrom, K. E., Flurkey, A. J., and Houston, R. S., 1983, Geol. Soc. America Bull., v. 94, p. 1257-1274.

Kay, R. W., and Senechal, R. G., 1976, Journal Geophy. Research, v. 81, p. 964-970.

LaBerge, G. L., and Myers, P. E., 1984, Geol. Soc. America Bull., v. 95, p. 246-253.

Larue, D. K., and Sloss, L. L., 1980, Geol. Soc. America Bull., Part I, v. 91, p. 450-452.

Lewry, J. F., 1981, Nature, v. 294, p. 69-72.

May, E. R., and Schmidt, P. G., 1982, Geol. Assoc. Canada Spec. Paper 25, p. 447-480.

Palmer, E. A., 1980, Unpublished M.S. thesis, Univ. of Minnesota-Duluth, Duluth, Mn, 127 p.

Schulz, K. J., Sims, P. K., Morey, G. B., in press, Precambrian: Conterminous U.S., DNAG v. C-2, Geol. Soc. America.

Sims, P. K., Peterman, Z. E., and Schulz, K. J., 1985, Geol. Soc. America Bull., v. 96, p. 1101-1112.

Sims, P. K., Schulz, K. J., Peterman, Z. E., and VanSchmus, W. R., in press, Precambrian: Conterminous U.S., DNAG v. C-2, Geol. Soc. America.

VanSchmus, W. R., 1980, In: Morey, G. B., and Hanson, G. N., eds., Geol. Soc. America Special Paper 182, p. 159-168.

Whitford, D. J., Nicholls, I. A., and Taylor, S. R., 1979, Contrib. Mineral Petrol., v. 70, p. 341-356.

Wood, D. A., Marsh, N. G., Tarney, J., Joron, J.-L., Fryer, P., and Treuil, M., 1981, In: Hussong, D. M., and Uyeda, S., and others (eds.) Init. Rep. Deep Sea Drilling Proj. 60, p. 611-645, U.S. Government Printing Office, Washington, D. C.

NEW INSIGHTS INTO TYPICAL ARCHEAN STRUCTURES IN GREENSTONE TERRANES OF WESTERN ONTARIO: W.M. Schwerdtner, Department of Geology, University of Toronto, Toronto, Canada M5S 1A1

Ongoing detailed field work in selected granitoid complexes of the western Wabigoon and Wawa Subprovinces, southern Canadian Shield, has led to several new conclusions: (1) Prominent gneiss domes are composed of prestrained tonalite-granodiorite and represent dense hoods of magmatic granitoid diapirs. The diapiric material commonly was a syenite-diorite crystal mush. (2) The deformation history of the prestrained gneiss remains to be unraveled. (3) The gneiss lacked a thick cover of mafic metavolcanics or other dense rocks at the time of magmatic diaprisim. (4) The synclinoral structure of large greenstone belts is older than the late gneiss domes and may have been initiated by volcano-tectonic processes. Multi-phase granitoid plutonism greatly tightened the synclinoria. (5) Small greenstone masses within the gneiss are complexly deformed, together with the gneiss. (6) No compelling evidence has been found of ductile early thrusting in the gneiss terranes. Zones of greenstone enclaves occure in hornblende-rich contaminated tonalite and are apt to be deformed magmatic septa. Elsewhere, the tonalite gneiss is biotite-rich and hornblende-poor.

These conclusions rest on several new pieces of structural evidence. (1) Oval plutons of syenite-diorite have magmatic strain fabrics and sharp contacts that are parallel to an axial-plane foliation in the surrounding refolded gneiss. (2) Gneiss domes are lithologically composite and contain large sheath-like structures which are deformed early plutons, distorted earlier gneiss domes, or early ductile nappes produced by folding of planar plutonic septa. (3) The predomal attitudes of gneissosity varied from point to point. It is difficult to prove by conventional structural methods what caused the state of early deformation in the large gneiss domes. New approaches are being developed based on the patterns of total and incremental finite strain in the granitoid terranes under study.

DEFORMATIONAL SEQUENCE OF A PORTION OF THE MICHIPICOTEN GREENSTONE BELT, CHABANEL TOWNSHIP, ONTARIO; Catherine H. Shrady and George E. McGill, Dept. of Geology and Geography, University of Massachusetts, Amherst, MA 01003

Detailed mapping at a scale of one inch = 400 feet is being carried out within a fume kill, having excellent exposure, located in the southwestern portion of the Michipicoten Greenstone Belt near Wawa, Ontario. A simplified geological map of the area described here is presented in a companion abstract (Fig. 1 in 1).

The rocks are metasediments and metavolcanics of lower greenschist facies. U-Pb geochronology indicates that they are at least 2698 ± 11 Ma old (2). The "lithologic packages" (1) strike northeast to northwest, but the dominant strike is approximately east-west. Sedimentary structures and graded bedding are well preserved, aiding in the structural interpretation of this multiply deformed area.

Deformation in this area is tentatively divided into six phases (0-5). Phase 0 is soft sediment deformation. Folds of this type are generally small (amplitudes ranging from several millimeters to tens of centimeters); however, some early larger scale (up to 10 meters in amplitude) tight to isoclinal folds with no or a very poorly developed axial plane cleavage may be slump folds.

Included within Phase 1 of deformation is the regional overturning resulting in rocks that dip north and young to the south in the northern part of our area and extending well to the north (1,3,4), and rocks that dip south and young north in the southern part of our area and farther south (1,3a). To what extent the regional steep dips are attributable to this phase of deformation or to later refolding is, at present, not known. Also included within Phase 1 are an approximately bedding parallel cleavage, and pebbles within conglomeratic units flattened parallel to this cleavage. It is thought that these latter two features are associated and likely relate to the regional overturning.

Cut by and therefore pre-dating Phase 2 cleavage, but of uncertain temporal relationship to the structures included within Phase 1, are areally significant faults that separate lithologic packages. These faults regionally follow but locally truncate bedding. In places, they are associated with an apparently old fracture cleavage.

Phase 2 is characterized by a penetrative northwest to north striking cleavage of moderate dip. Phase 2 cleavage crenulates Phase 1 cleavage where both are clearly present; however, in much of the area, these two cleavages cannot be separated. Related examples of mesoscopic folds are rare, and associated structures of regional significance have not been recognized.

Phase 3 cleavage is penetrative where well developed and crenulates both Phase 1 and Phase 2 cleavages. Within the area mapped, Phase 3 cleavage strikes northeast with generally steep northwest or southeast dips; dip direction and angle commonly change within individual outcrops. Dips as low as 30° are locally present in the northwest part of the area. It is not clear whether variation in dip indicates the existence of two distinct northeast striking cleavages or whether it is due to later minor folding about sub-horizontal axial surfaces. Phase 3 cleavage is axial planar to folds that are open to tight, range in scale from several millimeters to tens of meters in amplitude, and refold earlier folds. At one locality, Phase 3 cleavage and associated folds appear related to late movement on a fault that approximately parallels bedding. It is not yet clear if this fault is entirely young, or whether it is a reactivated older structure.

Steeply dipping northeast and north-northwest trending faults constitute Phase 4. However, some movement on these faults post-dates diabase dikes (Phase 5) that trend north-northwest and northeast. Locally developed fracture cleavages appear to be associated with diabase dike emplacement, but because the dikes commonly follow trends of older faults, some or all of these fracture cleavages may be related to the faults rather than to the dikes.

In summary: we have tentatively identified at least six phases of deformation within a relatively small area of the Michipicoten Greenstone Belt. These include the following structural features in approximate order of occurrence: 0) soft-sediment structures; 1) regionally overturned rocks, flattened pebbles, bedding parallel cleavage, and early, approximately bedding parallel faults; 2) northwest to north striking cleavage; 3) northeast striking cleavage and associated folds, and at least some late movement on approximately bedding parallel faults; 4) north-northwest and northeast trending faults; and 5) diabase dikes and associated fracture cleavages. Minor displacement of the diabase dikes occurs on faults that appear to be reactivated older structures.

References

(1) McGill, G.E. and Shrady, C.H. (1986) This volume.
(2) Turek, A., et al. (1984) Can. Jour. Earth Sci., 21, 457-464.
(3) Sage, R.P., et al. (1982) Ontario Geol. Surv. Prelim. Maps of Precambrian Geology: a. Chabanel Twp., Map P. 2439; b. Esquega Twp., Map P. 2440; C. McMurray Twp., Map P. 2441; d. Lastheels Twp., Map P. 2442; e. Musquash Twp., Map P. 2568.
(4) Sage, R.P., et al. (1984) Prelim. map of Precambrian geology of Corbiere Twp., Ontario Geol. Surv. Map P. 2607.

A CONTINENTAL RIFT MODEL FOR THE LA GRANDE GREENSTONE BELT;

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Stratigraphic relationships and the geochemistry of volcanic rocks constrain the nature and timing of the tectonic and magmatic processes in the pre-deformational history of the La Grande greenstone belt in the Superior Province of north-central Quebec (Fig. 1). With the exception of a locality in the western part of the belt the lowermost supracrustals in this belt are obscured by syntectonic granitoid intrusives. The supracrustal succession in the western part of the belt consists of a lower sequence of immature clastic sediments and mafic volcanoclastics, overlain by pillowed and massive basalts (Fig. 1, A-A'). Further east, along tectonic strike, a lower sequence of mafic volcanoclastics and immature clastic sediments is overlain by a thick sequence of pillowed and massive basalts, and resedimented coarse clastic sediments and banded iron formation. These are overlain by massive basaltic andesites, andesites and intermediate volcanoclastics intercalated with immature clastic sediments (Fig. 1, B-B'). In contrast, in the eastern part of the belt lenses of felsic volcanics and volcanoclastics occur at the base of the succession and pillowed and massive basalts are overlain by komatiites at the top (Fig. 1, C-C').

The lower sequences of clastic sediments in the central part of the belt reflect a mixed intrabasinal and extrabasinal provenance, but the upper clastic sediments have a uniquely extrabasinal tonalitic provenance. In addition metasedimentary and granitoid xenoliths have been found in the volcanic pile in the central and eastern parts of the belt and a local unconformable contact is believed to exist between the supracrustal succession and an underlying tonalitic basement in the west (1). Therefore a model in which the La Grande belt formed on a sialic crust is favoured.

The largest volumes of eruptive rocks in the La Grande belt are tholeiitic basalts (Fig. 2). These basalts are not primary mantle-derived liquids, but have undergone a polybaric fractionation history (1, 2 and 3). Their parental magmas are believed to have been basaltic komatiites (Fig. 2). The basaltic komatiites and most magnesian basalts lie along a steep slope in Al-Si space (Fig. 2) which is best explained by the fractional crystallization of orthopyroxene and olivine (4, 1). Co-existence of these two silicate phases and a liquid of basaltic composition is restricted

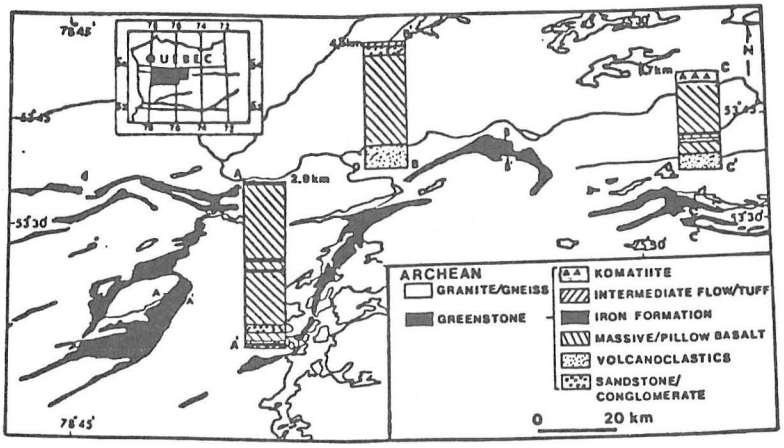

Figure 1 Geology of the La Grande greenstone belt.

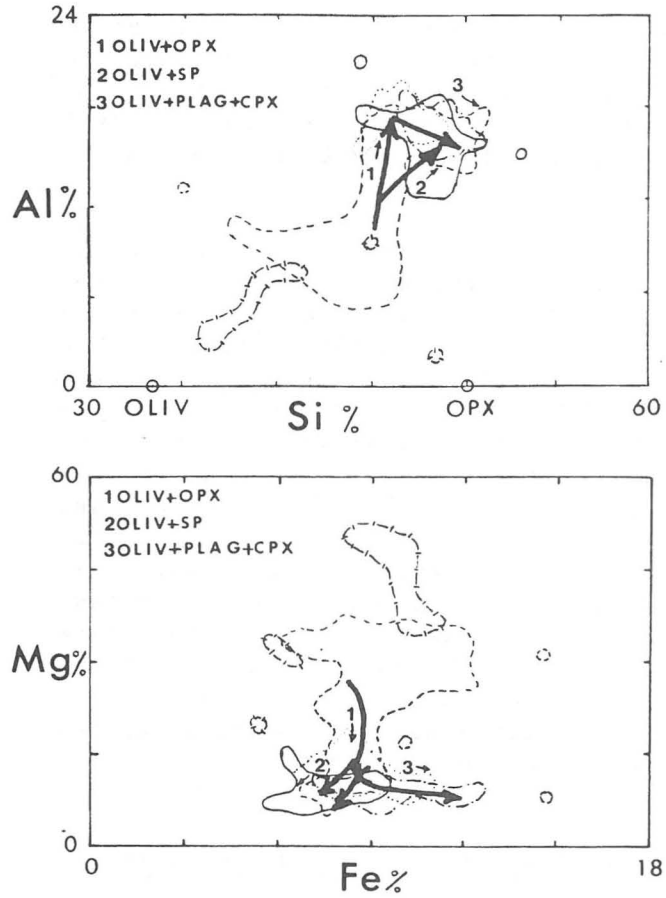

Figure 2 Al-Si and Mg-Fe in cation%. The solid line encloses basalts from section A-A', dotted line is basalts from section B-B', dash-bar and dash-dot are komatiites and basalts respectively from section C-C' and the dashed line includes komatiites and basalts from Lac Guyer (north of C-C').

to pressures on the order of 10 kb (5). Thus the basalts represent komatiitic liquids which have been modified by differing extents of fractionation at depths on the order of 30 km before migrating to higher levels in the crust (3, 1 and 6). A spectrum of basaltic compositions are found in the La Grande belt of which the endmembers are an Fe-enriched suite and those which have negligible Fe variation (Fig. 2). The Fe-enriched basalts have undergone extensive low pressure fractionation of a gabbroic assemblage, which is probably the result of a more protracted residence time in upper crustal conduit system than the relatively constant Fe group. The degree of fractionation of the komatiitic liquids and their location in space and time may reflect the variable efficiency of a crustal density filter (cf. 7). Thus, the occurrence of komatiitic lavas in the upper levels of the supracrustal succession may be due to late failure of the crustal barrier. Their restricion to the eastern parts of the belt may reflect development of a major rift only there. Ponding of mafic magmas within the sialic crust may have resulted in the melting of the crust and the early eruption of rhyolitic magmas in the east (4). Toward the central parts of the belt, komatiitic magmas ingested sialic crust, were modified by fractional crystallization and were ultimately erupted as basaltic andesites and andesites. These contaminated magmas are characterized by high compatible element (eg. Ni and Cr) and fractionated, enriched light rare earth element abundances (up to 100X chondrite) (8).

The La Grande greenstone belt can be explained as the product of continental rifting (6). The restricted occurrence of komatiites, and eastwardly directed paleocurrents in clastic sediments in the central part of the belt are consistent with rifting commencing in the east and propagating westward with time (Fig. 3). The increase in depth of emplacement and deposition with time of the lower three units (Fig 1, section B-B') in the central part of the belt reflects deposition in a subsiding basin (6). These supracrustal rocks are believed to represent the initial rift succession (c.f. 9). Model calculations (Fig.3) reveal that the extension factor for lithosphere neccessary to account for the observed initial subsidence in the

central part of the belt (6) is comparable in magnitude with that measured in Modern sedimentary basins where the continental lithosphere is believed to have been rapidly thinned (10). The occurrence of clastic sediments of granitic provenance high in the succession in the central parts of the belt may reflect the uplift and erosion of marginal forebulges that formed as a result of lithospheric flexure.

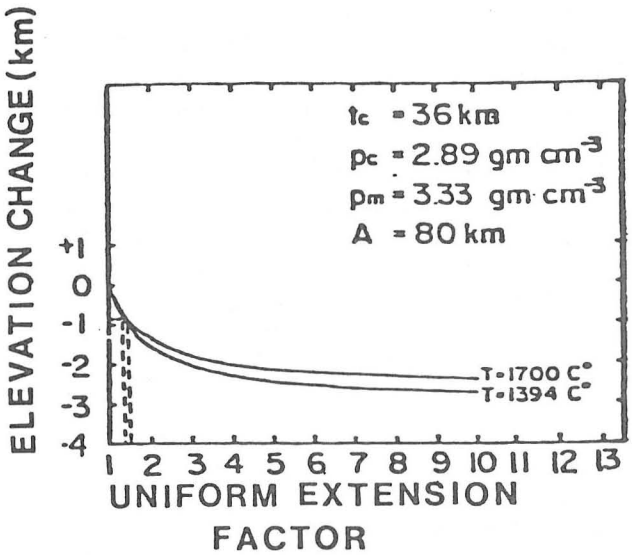

Figure 3 Initial elevation change versus uniform extension factor. For an initial elevation change of .9 km corresponding to the subsidence that is observed in the lower three units of section B-B'corrected for the basin fill and 1 km of water requires a uniform extension factor of approximately 1.5. The symbols used are: crustal thickness (tc), crustal and mantle densities (pc) and (pm) respectively, temperature at the base of the slab (T) and lithosphere thickness (A).The thermal expansion coefficient used is 3.2 x 10^{-5} C. The calculations were performed using the method of Royden and Keen (11).

Rivard B. and Francis D. (1986) Contrib. Mineral. Petrol. (in preparation).

Skulski T., Hynes A. and Francis D. (1984) in Chibougamau-Stratigraphy and Mineralization, CIM special V. 34, Guha J. and Chown E. editors, p. 57-72.

Stamatelopoulou-Seymour K., Francis D. and Ludden J. (1983) Contrib. Mineral. Petrol., 84, p. 6-14.

Stamatelopoulou-Seymour K. and Francis D. (1986) Precamb. Res. (in preparation).

O'Hara M. (1968) Earth-Sci Rev., 4, p. 69-133.

Skulski T., Hynes A. and Liu M. (1986) Precamb. Res. (in preparation).

Francis D., Ludden J. and Hynes A. (1983) J. Petrol., 24, p. 556-582.

Skulski T. (1986) unpublished MSc thesis, Department of Geological Sciences, McGill University.

Bickle M. and Eriksson K. (1982) Phil. Trans. Roy. Soc. Lond., A 305, p. 225-247.

McKenzie D. (1978) Earth. Planet. Sci. Lett., 40, p. 25-32.

Royden L. and Keen C.E. (1980) Earth Planet. Sci. Lett., 51, p. 343-361.

TWO CONTRASTING METAMORPHOSED ULTRAMAFIC-MAFIC COMPLEXES FROM GREENSTONE BELTS, THE NORTHERN KAAPVAAL CRATON AND THEIR SIGNIFICANCE IN ARCHAEAN TECTONICS

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The character of Archaean ultramafic-mafic complexes can, given their prominance in greenstone belts, provide critical clues to help deduce the tectonic setting of these belts. Here we describe two contrasting, metamorphosed, ultramafic-mafic complexes, the first a partially serpentinised dunitic body with associated chromite from Lemoenfontein, one of several peridotitic bodies occurring as discrete lenses and pods in granulite facies gneisses of the northern Kaapvaal craton. The second, the Rooiwater complex is a major layered igneous body, now metamorphosed in the amphibolite facies, but without pervasive deformation, which crops out in the northern Murchison greenstone belt.

The Lemoenfontein body is circular, about 100m in diameter, having the form of a steeply plunging boudin which complements the regional structural pattern. The surrounding granulite facies gneisses were isotopically reset about 2650 Ma and may be considerably older. The Lemoenfontein rocks are partially serpentinised dunite, displaying a prominent tectonic fabric defined by the preferred orientation of olivine grains, chromite pods and disseminated chromite stringers, all of which are believed to have been through the granulite facies metamorphism. Chromite is present as massive high-grade ore, 'leopard' (nodular) ore, tectonically layered ore and disseminated ore. Zones of chromite enrichment range in thickness from 1 to 30cm. The Lemoenfontein chromites are similar to those mined in the Ultramafic Formation of the Selukwe greenstone belt, Zimbabwe.

Olivines from Lemoenfontein are Fo94 to Fo96 with NiO contents from 0.35 to 0.59wt%. The mineral chemistry of the chro-

mites of all different types (pods, trains, and inclusions in silicate grains) is very similar indicating either complete metamorphic equilibration or they represent consistent primary compositions. The Lemoenfontein chromites have refractory characteristics (low TiO_2, Al_2O_3 and alkali metals) and plot on geochemical fence diagrams in or close to the fields of other podiform chromites. Rocks which in Phanerozoic series are closely associated with alpine-type peridotites or ophiolite suites.

The Rooiwater complex is a thick on end differentiated igneous body, of age greater than 2650 Ma, probably intruded at 2960 Ma. The complex is heterogenously deformed with much of the 7.5km exposed thickness showing no pervasive deformation. Metamorphosed pyroxenite, anorthosite, gabbro, sulphide-bearing gabbros, thick magnetitite layers and differentiated granites are compatible with the hypothesis that the body is a layered intrusion although it is now allochthonous and intruded by younger unrelated granites. Southward increasing TiO_2 and decreasing V_2O_5 contents in magnetitite layers combined with a general southerly disposition of differentiated hornblende granite suggest that the Rooiwater complex is southward facing. A paucity of ultramafic cumulates and up to 1.5km of highly differentiated hornblende granite suggests that the original magma was more felsic than that of similar layered intrusions.

The Lemoenfontein chromites and associated ultramafic rocks are lithologically and chemically similar to their Phanerozoic equivalents of ophiolitic origin, interpreted as obducted oceanic crust. Similarly we interpret the Lemoenfontein complex as being a remnant of Archaean oceanic material. In contract, the Rooiwater complex is, despite the lack of exposed intrusive contacts, similar to layered igneous complexes such as Ushushwana or Bushveld. These complexes are intrusive in continental environments. We conclude that contrasting ultramafic-mafic complexes represent a heterogeneity in greenstone belts with either

oceanic or continental environments involved. Whether this heterogeneity relates to a temporal or spatial (or both) control remains uncertain.

ZIRCON Lu-Hf SYSTEMATICS: EVIDENCE FOR THE EPISODIC DEVELOPMENT OF ARCHEAN GREENSTONE BELTS P. E. Smith, M. Tatsumoto* and R. M. Farquhar, Dept. of Physics, Geophysics Division, University of Toronto, Canada M5S 1A7; U.S. Geological Survey*, Federal Center, MS 963, P.O. Box 25046, Lakewood, CO 80225

A combined U-Th-Pb and Lu-Hf isotopic study of zircons was undertaken in order to determine the provenance and age of an Archean granite-greenstone terrain and to test the detailed application of the Lu-Hf system in various Archean zircons.

The eastern Wawa subprovince of the Superior province consists of the low grade Michipicoten and Gamitagama greenstone belts and the granitic terrain. Earlier studies have established the structural and stratigraphic relationships of the area (1-4). The adjacent high grade Kapuskasing zone is believed to represent the lower crustal levels to the greenstone belts (5).

The rock units of this area have been the subject of extensive geochronological studies using zircon U-Pb (6, 7) and whole rock U-Th-Pb methods (Smith, et al in prep.). The three volcanic cycles recognized in the area have mean ages of 2748 My (cycle I), 2732 My (cycle II), and 2714 My (cycle III). Syntectonic granitoids which surround the supracrustal rocks date from the cessation of cycle I volcanic rocks, to the time of post-tectonic plutonism dated at 2666 $\pm$ 2 My. The oldest rocks yet dated come from a granite dated at 2888 $\pm$ My which is possibly the basement to the volcanic rocks. Zircon ages from the Kapuskasing zone appear to reflect updating during the regional metamorphism (8).

The Lu, Hf, U and Th contents of zircons from these rocks reveal patterns that may be indicative of their source regions (Fig. 1). Zircons from rocks of granitic composition appear to have distinct enrichments in U and Th relative to zircons from rocks of more intermediate composition. More striking however, is the severe depletion of Lu and Hf from the zircons from the Kapuskasing area. The lowest Hf content measured so far, 1790 ppm, is from zircons from a mafic gneiss. The elemental patterns in the lower crustal zircons suggest that Lu and Hf loss accompanies Pb loss during high grade metamorphism.

The U-Pb age corrected Hf isotopic ratios from the zircons indicate significant long-lived heterogeneity of source regions for the greenstone belts (Fig. 2). Overall the heterogeneity in the ratios may be attributed to three isotopically distinct sources: (1) a high Lu/Hf source, (2) a moderately enriched Lu/Hf source; and (3) a sub-chondritic Lu/Hf source.

The high Lu/Hf source is represented by a sub-volcanic intrusive from cycle II and two tholeiites (whole rock determinations) from the lower stratigraphic levels of cycles I and II. The epsilon Hf values range from +8.7 to +11.6 and the source is believed to represent the depleted mantle.

The second source has epsilon Hf values ranging from +1.4 to +5.9. There is an apparent alignment of dacitic volcanic rocks and their sub-volcanic equivalents from cycles I and II with the tonalitic syntectonic granitoids. It is believed that the source of these rocks was the lower crust and it can be inferred that previous intracrustal differentiation led to a high Lu/Hf lower crustal reservoir. The process which led to the enhanced Lu/Hf ratio was most likely Hf loss as attested to by the Kapuskasing zircons. A greater than chondritic Lu/Hf ratio for the lower crust may explain the apparent non-coherence of initial Nd and initial Hf ratios for an Archean tonalite reported in the literature (9, 10).

The low Lu/Hf source is represented by rhyolites capping the sequences of cycles I and III and by post-tectonic potassic granitoids. Their epsilon Hf values ranging from -1.3 to +1.4, significantly lower than the coeval dacites, are indicative of an upper crustal source.

The Hf isotopic data from the three volcanic cycles indicate that the typical lithological features of a greenstone belt cycle could be accommodated in a crustal growth model that involved decreasing depth of melting in three isotopically distinct reservoirs: mantle, lower crust and upper crust. The model age of the sources given by the intersection of the lower crustal curve with the bulk earth evolution curve (11) is about 2900 My, in good agreement with the zircon U-Pb basement age. This linear array also has a similar intersection age to that of Proterozoic carbonatite complexes studied by Bell et al (12). The general convergence of the other reservoir vectors around this age suggests that mantle depletion, crustal extraction and intracrustal differentiation were all part of the same episodic event. It is also apparent that recycling of older basement was important in the formation of many of the later greenstone belt rocks.

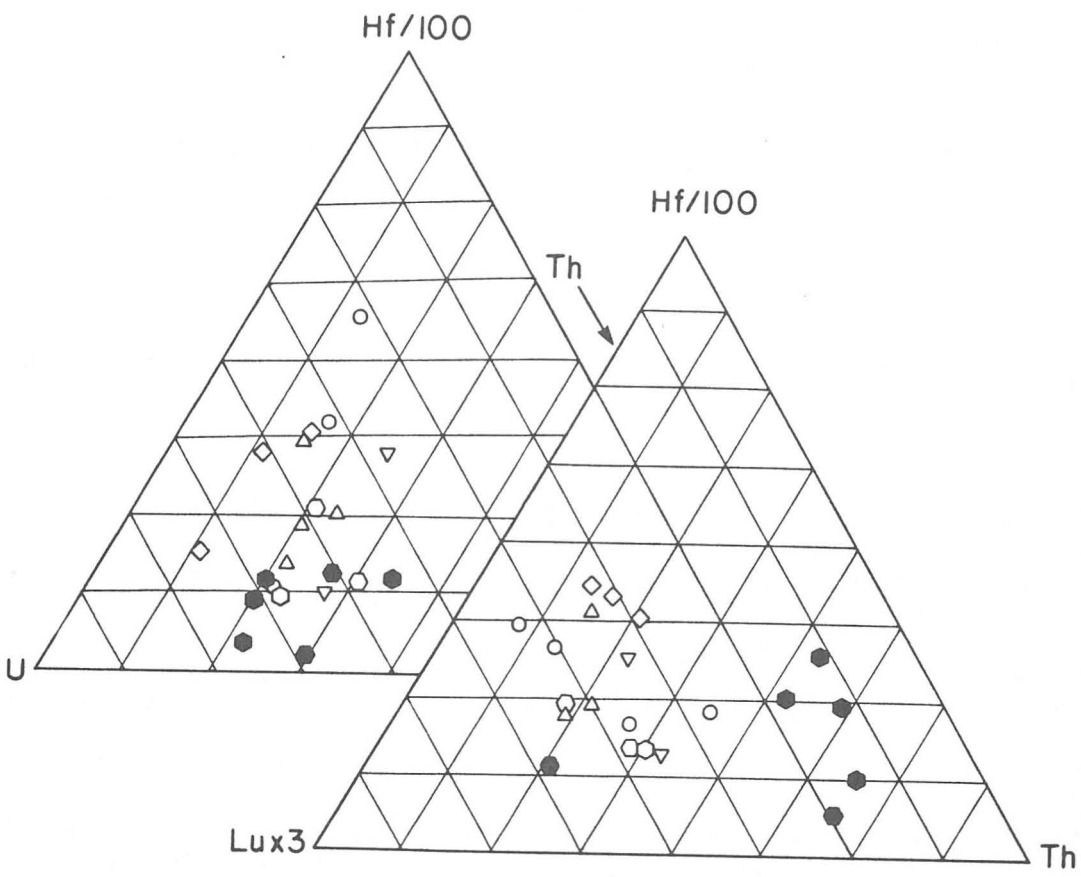

Figure 1

Relative abundances of Lu, Hf, U and Th for eastern Wawa subprovince zircons. Symbols are: △ Dacitic volcanic rocks; ▽ rhyolites; ○ sub-volcanic granitoids; ◇ syntectonic granitoids; ⬡ post-tectonic granitoids; ⬡ basement granite zircons; ⬢ Kapuskasing zircons; ⬢ conglomerate boulder zircons.

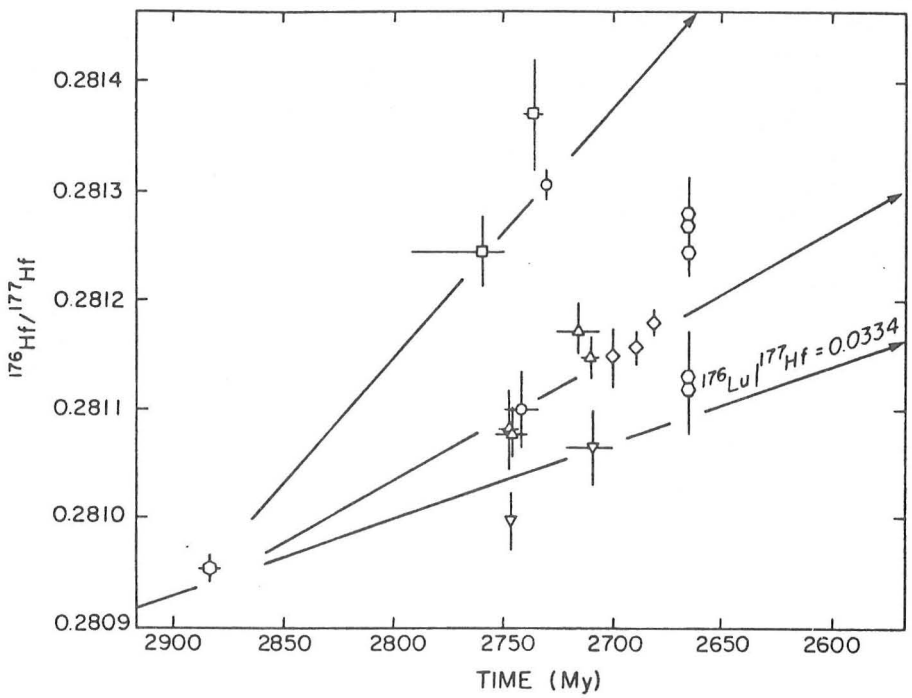

Figure 2

Initial $^{176}Hf/^{177}Hf$ vs T diagram for zircons and whole rocks (□). Symbols as in Figure 1.

References

(1) Goodwin, A. M. (1962) Geol. Soc. Am. Bull. 73, 561-586.
(2) Ayres, L. D. (1969) Ont. Dept. Mines Geol. Rep. 69.
(3) Attoh, K. (1980) Current Res. part A. Geol. Surv. Can. Pap. 80-1 A, 101-106.
(4) Sage, R. P. (1980) Ont. Geol. Surv. Miscel. Pap. 96, 47-50.
(5) Percival, J.A. and Card, K. D. (1983) Geology 11, 323-326.
(6) Turek, A., Smith, P. E. and Van Schmus, W. R. (1982) Can. J. Earth Sci. 17, 866-876.
(7) Turek, A., Smith, P. E. and Van Schmus, W. R. (1984) Can. J. Earth Sci. 21, 457-464.
(8) Percival, J. A. and Krogh, T. E. (1983) Can. J. Earth Sci. 20, 830-843.
(9) Patchett, J. P., Kouvo, O., Hedge, C. E. and Tatsumoto, M. (1981) Contr. Miner. Petrol. 78, 279- 297.
(10) Jahn, B., Vidal, P. and Kroner, A. (1984) Contr. Miner. Petrol. 86, 398-408.
(11) Patchett, J. P. and Tatsumoto, M. (1980) Contr. Miner. Petrol. 75, 263-267.
(12) Bell, K. Blenkinsop, J., Cole, T. J. S. and Menagh, D. P. (1982) Nature 298, 251-253.

EARLY PRECAMBRIAN CRUSTAL EVOLUTION OF SOUTH INDIA

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The Early Precambrian sequence in Karnataka, South India provides evidences for a distinct trend of evolution which differs from trends exhibited in many other Early Precambrian regions of the world. The supracrustal rock associations preserved in greenstone belts and as inclusions in gneisses and granulites suggest the evolution of the terrain from a stable to a mobile regime. The stable regime is represented by 1. layered ultramafic-mafic complexes, 2. orthoquartzite-basalt-rhyodacite-iron formation, and 3. orthoquartzite-carbonate-Mn-Fe formation. The mobile regime which can be shown on sedimentological grounds to have succeeded the stable regime witnessed accumulation of a greywacke-pillow basalt-dacite-rhyolite-iron formation association. Detrital sediments of the stable zone accumulated dominantly in fluvial environment and the associated volcanics are subaerial. The volcanics of the stable regime are tholeiites derived from a zirconium and LREE-enriched source. The greywackes of the mobile regime are turbidites, and the volcanic rocks possess continental margin (island-arc or back-arc) affinity; they show a LREE-depleted to slightly LREE-enriched pattern. The evolution from a stable to a mobile regime is in contrast to the trend seen in most other regions of the world, where an early dominantly volcanic association of a mobile regime gives way upward in the sequence to sediments characteristic of a stable regime.

Structures in greenstone belts, in the gneisses surrounding them, and also in the inclusions in the gneisses are similar in style, sequence, and orientation. This structural unity which is present in spite of the three thermal peaks recorded by radiometric ages around 3300, 3000 and 2600 m.y. ago, indicates long range stability of tectonic stress regimes in the Archaean lithosphere. The continuation of structures and rock formations across the greenstone-granulite boundary suggests that the two provinces did not evolve in separate tectonic blocks but represent only different crustal levels.

The preservation of detrital pyrite-uraninite bearing conglomerates, iron formations, and carbonate rocks provide an unique opportunity for the elucidation of evolutionary changes from oxygen-deficient to oxygenic atmosphere-hydrosphere conditions. Large scale development of iron formations and limestones in the greenstone belts of South India at least 3000 m.y. ago suggests that these may be the earliest large-scale sinks for the photosynthetically produced oxygen. Detailed palaeobiological and biogeochemical studies of these rock formations are necessary.

BASEMENT-COVER RELATIONS AND INTERNAL STRUCTURE OF THE CAPE SMITH KLIPPE: A 1.9 Ga GREENSTONE BELT IN NORTHERN QUEBEC, CANADA; M.R. St-Onge and P.F. Hoffman, Geological Survey of Canada, 588 Booth St., Ottawa, Ont. K1A 0E4, S.B. Lucas, Geological Sciences Dept, Brown University, Providence, R.I. 02912, D.J. Scott, Geology Dept, McMaster University, Hamilton, Ont. L8S 4L8, and N.J. Bégin, Geological Sciences Dept, Queen's University, Kingston, Ont. K7L 3N6.

The Cape Smith Belt is a 380x60 km tectonic klippe (1 and references therein) composed of greenschist- to amphibolite-grade mafic and komatiitic lava flows and fine-grained quartzose sediment, intruded by minor syn- to post-tectonic granitoids. Previously studied transects in areas of relatively high structural level show that the belt is constructed of seven or more north-dipping thrust sheets which verge toward the Superior Province (Archean) foreland in the south and away from an Archean basement massif (Kovik Antiform) external to the Trans-Hudson Orogen (Early Proterozoic) in the north. A field project (mapping and structural-stratigraphic-metamorphic studies) directed by MRS was begun in 1985 aimed at the structurally deeper levels of the belt and underlying basement, which are superbly exposed in oblique cross-section (12 km minimum structural relief) at the west-plunging eastern end of the belt. Mapping now complete of the eastern end of the belt confirms that all of the metavolcanic and most of the metasedimentary rocks are allochthonous with respect to the Archean basement, and that the thrusts must have been rooted north of Kovik Antiform. The main findings (2) are:

1. A thin autochthonous to parautochthonous low-strain sedimentary sequence on the south margin of the belt rests directly on Archean basement showing no evidence of Proterozoic transposition.

2. The bulk of the belt is separated from the autochthon by a sole thrust which, except at the south margin of the belt, is located at the basement-cover contact. The hangingwall and footwall rocks of the sole thrust record high ductile strains over a zone of increasing width, from south to north, toward the hinterland. Late syn-metamorphic thrusts faults with relatively small displacements cut the sole thrust and its associated shear zone, and place basement gneisses over cover rocks.

3. Lensoid meta-ultramafic tectonic blocks occur locally within the basal shear zone. Their metamorphic anthophyllite-actinolite assemblage differs from the serpentine-tremolite assemblage of cumulate meta-ultramafics occurring in sills at higher structural levels. The blocks may have been tectonically transported from mantle depths during thrusting, although this idea remains to be tested.

4. The allochthonous rocks above the sole thrust occur in a series of thrust sheets bounded by south-verging (D1) thrust faults, which are defined by structural repetitions of stratigraphy and splay from the sole thrust. Favorable lithologies at all structural levels (excepting the southern autochthonous margin) have a pevetrative syn-metamorphic schistosity (S1) which is planar to south-facing tight to isoclinal folds of bedding (F1).

5. A transverse stretching lineation (L1) common in the lower stgructural levels and pervasive in the basal shear zone, when considered with the F1 fold asymmetry and overall thrust-ramp geometry, indicates relative southward translation of the cover during D1.

6. A pelitic interval above the sole thrust on the north margin of the belt contains the metamorphic assemblage kyanite-staurolite-garnet-biotite-muscovite-plagioclase-quartz. The asemblage is indicative of metamorphic T of 550°C and minimum P of 5.5 Kbars.

7. Mesoscopic late- to post-metamorphic chevron to rounded parallel folds (F2) of the S1 fabric have a marked limb asymmetry suggestive of a gravitational origin as folds cascading off basement-cored macroscopic D2 antiforms into pinched cover-rock synforms. The distribution of north- versus south-vergent mesoscopic folds however is not always consistent with the mapped limbs of the macroscopic folds, possibly reflecting diachronous development of the macroscopic folds.

8. Macroscopic high-angle D3 crossfolds affect both the basement and cover in the eastern half of the belt and provide a cumulative structural relief of 12-15 km. D3 fold hinges are readily documented by reversals in plunge azimuth of the D2 folds. Plunge projections permit the construction of a composite structural cross-section linking the highest and lowest structural levels of the belt.

The main implication of these observations is that the presence of Archean basement beneath the belt has no direct bearing on the question of the tectonic setting of the mafic-ultramafic magmatism.

REFERENCES

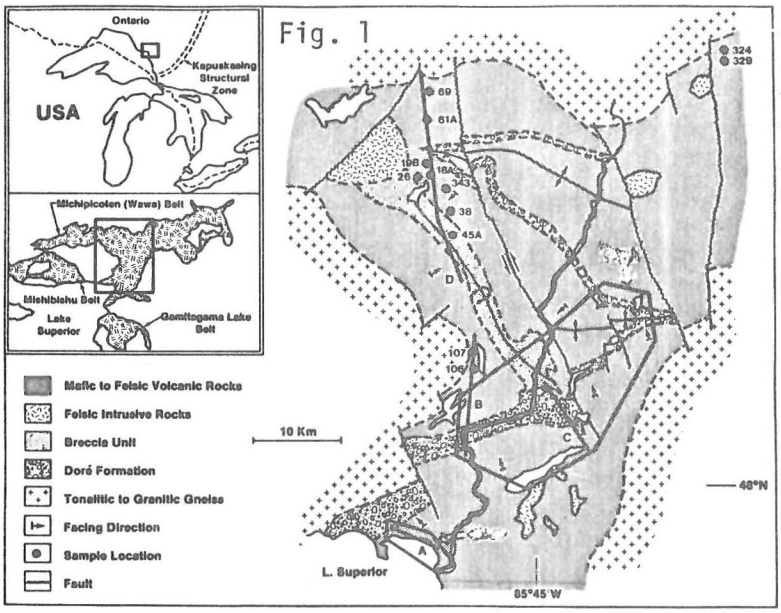
(1) Hoffman, P.F. (1985) Can. J. Earth Sci., 22, 1361-1369.
(2) St-Onge, M.R. et al. (1986), Geol. Surv. Can. Pap. 86-1A.

RHYOLITIC COMPONENTS OF THE MICHIPICOTEN GREENSTONE BELT, ONTARIO: EVIDENCE FOR LATE ARCHEAN INTRACONTINENTAL RIFTS OR CONVERGENT PLATE MARGINS IN THE CANADIAN SHIELD? Paul J. Sylvester, SN4, NASA/Johnson Space Center, Houston, TX 77058; Kodjo Attoh, Dept. of Geology, Hope College, Holland, MI 49423; Klaus J. Schulz, U.S. Geological Survey, Reston, VA 22092

Rhyolitic rocks often are the dominant felsic end member of the bimodal volcanic suites that characterize many late Archean greenstone belts of the Canadian Shield [1]. The rhyolites primarily are pyroclastic flows (ash flow tuffs) emplaced following plinian eruptions [2], although deposits formed by lava flows and phreatomagmatic eruptions also are present. Based both on measured tectono-stratigraphic sections and provenance studies of greenstone belt sedimentary sequences [3], the rhyolites are believed to have been equal in abundance to associated basaltic rocks.

In many recent discussions of the tectonic setting of late Archean Canadian greenstone belts, rhyolites have been interpreted as products of intracontinental rifting [2,4]. A study of the tectono-stratigraphic relationships, rock associations and chemical characteristics of the particularly well-exposed late Archean rhyolites of the Michipicoten greenstone belt, Ontario (figure 1) suggests that convergent plate margin models are more appropriate.

Three time-equivalent stratigraphic sequences of volcanism (figure 2), each including both mafic and felsic rocks, have been recognized in the Michipicoten greenstone belt [5,6,7,8]. The lower volcanic sequence is most well-preserved and therefore has been studied in most detail. It consists of a largely mafic unit (MV1) conformably overlain by a thick (up to about 700m), mainly felsic volcanic succession (FV1), which was emplaced approximately 2743 Ma ago [9]. In the Michipicoten Harbour area, an undated basal felsic flow unit is structurally discontinuous with the mafic sequence. Along the northern margin of the belt, epiclastic sediments are deposited on apparently older granitoid basement, and are overlain by felsic volcanics (and iron formation) that may be time-correlative with the Michipicoten Harbour felsic flows.



A range of depositional environments apparently existed for the felsic volcanic rocks of the lower volcanic sequence. Subaerial non-welded massive ash flows, shallow water accretionary lapilli-bearing hyalotuffs and deeper water bedded pyroclastic deposits all have been recognized [6,7,10]. Similarly, sedimentary rocks that overlie the lower volcanic sequence were deposited in both subaerial (braided fluvial and alluvial fan) and subaqueous (turbidite) environments [11].

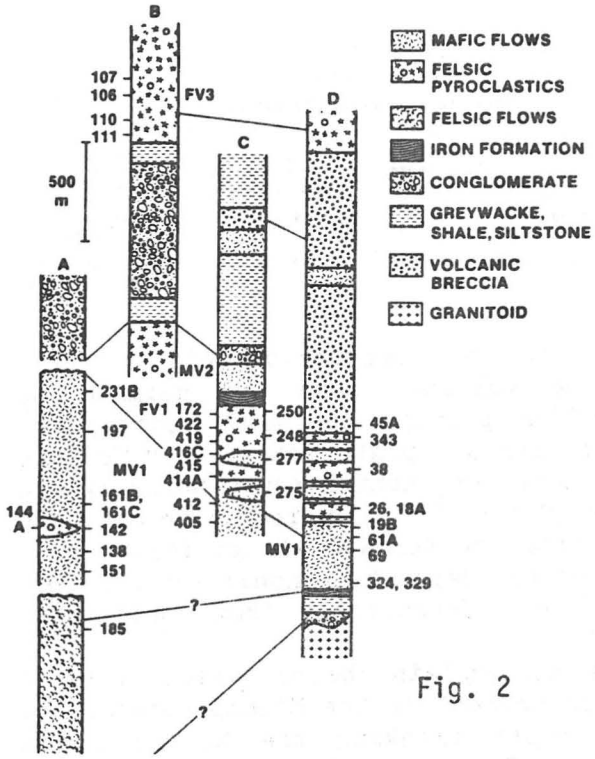


Fig. 2

Voluminous Cenozoic rhyolitic pyroclastic deposits are erupted on continental (rather than oceanic) crust and exhibit distinctive chemical characteristics and rock associations depending on whether that crust was the site of intracontinental rifting or subduction. Three examples of Cenozoic rhyolites associated with intracontinental, extension-related tectonism are presented in Table 1. The Trans-Pecos volcanic province of west Texas represents a rift dominated by alkaline to peralkaline rocks of bimodal basalt-rhyolite composition. The rhyolites are dominated by low-silica (<75wt%) compositions that tend to be depleted in alumina and lime relative to iron and the alkalis. The Rio Grande rift of New Mexico consists of a more continuous spectrum of mafic to felsic compositions that are commonly described as calc-alkaline [14]. Rhyolitic rocks, such as the Bandelier Tuff, are dominated by high-silica compositions. The Yellowstone Plateau volcanic field represents a third extension-related rhyolite group characterized by an association with continental flood basalts and "hot spot" activity. Yellowstone rhyolites are compositionally similar to the subalkaline rhyolites of the Rio Grande rift.

Cenozoic ash flow tuffs of rhyolitic composition also are erupted in voluminous proportions in continental inner arc regions of convergent plate margins. Relative to rhyolites formed in intracontinental rifts or hot spots, inner arc subduction-related rhyolites tend to have higher ratios of alumina and lime to iron and the alkalis (> about 1.4) and a more continuous spectrum of low- to high-silica compositions. Three examples of inner arc Cenozoic rhyolites are listed in Table 2. They differ mainly with respect to whether a field association with voluminous coeval intermediate volcanics is present (San Juan field), ambiguous (Sierra Madre Occidental) or not found

TABLE 1

VOLUMINOUS CENOZOIC RHYOLITIC ASH-FLOW TUFFS

Extension-related, intracontinental suites

	Dominant SiO_2 range	$(Al_2O_3 + CaO)/(FeO_t + alkalis)$	Associated volcanics
1. Trans Pecos Volcanic Province, Texas [12]	70-75 wt.%	0.76 - 1.25	hawaiite, mugearite, trachyte
2. Bandelier Tuff, Jemez Mountains, New Mexico [13,14]	74 - 77	1.21 - 1.43	basaltic andesite to rhyodacite
3. Yellowstone Plateau Volcanic Field [15]	75 - 77	1.24 - 1.37	olivine tholeiite

TABLE 2

VOLUMINOUS CENOZOIC RHYOLITIC ASH-FLOW TUFFS

Subduction-related, continental inner arc suites

	Dominant SiO_2 range	$(Al_2O_3 + CaO)/(FeO_t + alkalis)$	Associated volcanics
1. Taupo Volcanic Zone, New Zealand [16]	69-77 wt.%	1.57 - 1.71	minor high-Al basalt to dacite
2. Mid-Tertiary Upper Volcanic Sequence, Sierra Madre Occidental, Mexico [17]	70 - 77	1.62 - 1.85	minor basaltic andesite to dacite
3. Oligocene Ash-Flows, San Juan Volcanic Field, Colorado [18]	64 - 76	1.34 - 2.05	voluminous andesite to qtz latite

(Taupo volcanic zone).

If Cenozoic rhyolites may be used as a guide, the Michipicoten lower volcanic sequence (FV1) rhyolites, which are characterized by a continuous spectrum of silica compositions and relatively high ratios of alumina and lime to iron and the alkalis (Table 3), are more likely to be subduction-related than intracontinental rift-related. The Taupo volcanic zone and neighboring Kermadec-Tonga island arc system [19] offer perhaps the most appropriate plate tectonic analogue. At this convergent plate margin, rhyolitic pyroclastic rocks erupted from the New Zealand continental crust actually are deposited largely on the adjacent sea floor [20], which also is the depositional site for tholeiites derived from the Kermadec-Tonga island arc. The resulting ocean floor/continental slope deposits should consist of interfingering rhyolites and basalts derived independently from continental and oceanic platforms, respectively.

TABLE 3

LOWER CYCLE MICHIPICOTEN RHYOLITES (FV1)

	Dominant SiO_2 range	$(Al_2O_3 + CaO)/(FeO_t + alkalis)$	Associated volcanics
Low-silica type	70-73 wt.%	1.69 - 2.44	minor enriched basalt, dacite
High-silica type	74 - 78	1.67 - 2.28	minor enriched basalt, dacite

A similar tectonic-depositional model may explain the so-called cyclical mafic to felsic stratigraphic relationships present in the Michipicoten belt. The presence of pre-existing granitoid crust flanking the belt and the well-known compositional similarity between Cenozoic island arc tholeiites and Archean greenstone belt tholeiites [21], such as those present in the Michipicoten belt [22], support this interpretation. However, the existence of subaerial and shallow subaqueous depositional environments for some Michipicoten volcanic, volcaniclastic and sedimentary units requires either intermittent, local emergence of the volcanic pile or the existence of at least small continental blocks underlying parts of the belt.

References: (1) Thurston et al. (1985) Geol. Assoc. Canada Spec. Pap. 28, p. 7-21. (2) Ayres, L.D. and Thurston, P.C. (1985) Ibid., p. 343-380. (3) Ayres, L.D. (1983) Can. J. Earth Sci. 20, p. 1168-1194. (4) Gelinas, L. and Ludden, J.N. (1984) Phys. Earth Planet. Int. 35, p.77-88. (5) Goodwin, A.M. (1962) Geol. Soc. Am. Bull. 73, p. 561-586. (6) Attoh, K. (1980) Geol. Surv. Canada, Paper 80-1A, p. 101-106. (7) Attoh, K. (1981) Geol. Surv. Canada, Paper 81-1B, p. 49-54. (8) Sage, R.P. (1979) Ontario Geol. Surv., Misc. Pap. 90, p.48-53. (9) Turek, A., Smith, P.E. and Van Schmus, W.R. (1984) Can. J. Earth Sci. 21, p. 457-464. (10) Morton, R.L. and Nebel, M. (1983) Precamb. Res. 20, p. 39-62. (11) Ojakangas, R.W. (1983) LPI Tech. Rpt. No. 83-03, p. 66-70. (12) Noble, D.C. and Parker, D.F. (1974) Bull. Volcanol. 38, p. 803-827. (13) Sommer, M.A. and Schramm, L.S. (1983) Bull. Volcanol. 46, p. 306-316. (14) Guilbeau, K.P. and Kudo, A.M. (1985) Geol. Soc. Am. Bull. 96, p. 108-113. (15) Hildreth, W., Christiansen, R.L. and O'Neil, J.R. (1984) J. Geophys. Res. 89, p. 8339-8369. (16) Cole, J.W. (1979) N.Z. J. Geol. Geophys. 22, p. 631-657. (17) Cameron, K.L. and Hanson, G.N. (1982) Geochim. Cosmochim. Acta 46, p. 1489-1503. (18) Lipman, P.W. (1975) U.S. Geol. Surv. Prof. Pap. 852, 128pp. (19) Ewart, A., Brothers, R.N. and Mateen, A. (1977) J. Volcanol. Geotherm. Res. 2, p. 205-250. (20) Walker, G.P.L. (1980) Ibid., 8, p. 69-94. (21) Condie, K.C. (1985) J. Petrol. 26, p. 545-563. (22) Sylvester, P., Attoh, K. and Schulz, K.J. (1985) Geol. Assoc. Canada Program with Abstr. 10, p. A60.

GEOPHYSICAL CHARACTERISTICS AND CRUSTAL STRUCTURE OF GREENSTONE TERRANES, CANADIAN SHIELD;

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A knowledge of the deep structure and geometry of greenstone belts is fundamental to tectonic models of Archean evolution. In the Canadian Shield long linear granite-greenstone terranes of generally low metamorphic grade alternate with temporally-equivalent metasedimentary belts of higher grade. The focus of geophysical investigations of these terranes has been to examine geometries and contact relationships within individual terranes, and to look at the broader and deeper aspects of structure and inter-terrane relationships.

Major greenstone belts are characterized by positive gravity anomalies in the range 15-30 mGal that primarily reflect the relatively high density mafic and ultramafic metavolcanic components (1). These anomalies are sometimes interrupted by negative anomalies caused by felsic plutons and are poorly developed where high metamorphic grade basement is present and/or boundaries are gently-dipping. Modelling reveals that many greenstone belts are more or less basin-shaped, some having deep keels, and that their steep surface boundaries extend to depth. Model depths of polycyclic greenstones are 2-8 km and non-polycyclic are 3-12 km (1). The generally smaller depths of the former have been attributed to granitic intrusion decreasing vertical extent by stoping, or to listric normal faulting or thrusting (1). Models indicate abrupt changes in depth of up to ~10 km between supracrustals of the Wawa greenstone and Quetico metasedimentary terranes and point to a major faulted contact (2). Granitic intrusions at and within boundaries of greenstones are associated with prominent negative gravity anomalies. Modelling indicates that they have depths ranging from 2-16 km with depths in the middle of the range being characteristic (3,4). Generally, the contacts of the granites are modelled as steeply dipping. Some granites extend several kilometres deeper than adjacent greenstones but in other cases greenstones are interpreted to underlie the granite. For example, interpretation of a combined gravity-seismic study of the Aulneau batholith of the Wabigoon subprovince suggests that it is floored by up to 10 km of greenstones (3). Gravity studies in Wabigoon subprovince have contributed to classifying granites into epizonal sheets and deep diapiric batholiths intruded in two separate periods (4).

Regionally, greenstone belts generally correspond to magnetic lows and associated granites to magnetic highs (5,6). Magnetization studies (6) indicate values that are generally < 0.05 A/m for greenstones and > 0.05 A/m for granites. Linear positive anomalies within the English River gneiss belt have drawn attention to pyroxene amphibolite gneisses, probably derived from metavolcanics (7). Their occurrence is significant in that they are in an area where volcanism is thought not to have been important. Aeromagnetic shaded relief maps have been used to assist in mapping surface geology in the Abitibi greenstone belt (8). Various features correlate with diorite-gabbro and peridotite-serpentinite intrusions, diabase dykes, major faults, iron formations and zones of contact metamorphism around granitic intrusions. The magnetic signature

of the Abitibi belt, however, is not noticeably different from that of the bordering terranes. Modelling has been limited. Interpretation of a 300 km N-S profile across the Abitibi belt (8) indicates that the greenstones extend to a maximum depth of 13.6 km in the south, with an average depth of ~9 km compared to 6 km in the north. This agrees with seismic refraction results that suggest the bottom of the belt dips southward increasing in depth from 6 to 14 km (9). Surface magnetic units over granites of the Wabigoon belt have been modelled as extending to the intermediate discontinuity (16-19 km) with an increase in magnetization occurring at a few kilometres depth (6). Magnetization is low or absent below the discontinuity.

Seismic reflection studies within the Aulneau batholith and adjacent greenstones (10,11) have mapped a near-vertical contact between granite and greenstone to a depth of several kilometres (confirmed by later gravity studies) and a vertical fault zone. Although there are no detectable velocity differences between the greenstones and granites, the impedance contrast is sufficient to produce recognizable reflections from the near-vertical contact. The lower surface of the batholith, as interpreted from gravity, did not produce reflections, perhaps due to its undulatory nature (12). There is also a poor correlation between the average depth of the Yellowknife greenstone belt as determined from seismic (~10 km) and gravity (~3 km) studies (13,14). In contrast, the seismic refraction survey (9) across the Abitibi belt yielded a geometry for the bottom of the belt similar to that based on magnetic interpretation (8). The seismic investigations in the vicinity of the Aulneau batholith (10,11) also detected several deep horizontal or near-horizontal reflectors. The most prominent reflectors are at intermediate depths of about 19 and 22 km and the Moho at 38 km. The three reflectors appear to be continuous beneath the granite and greenstones suggesting that complex structure, which typifies the upper crust, is absent at depth. A similar picture of the Wabigoon crust has been found by long-range refraction - wide angle reflection experiments (15,16), but in the Quetico metasedimentary belt to the south no sharp boundaries are found within or at the base of the crust which is about 40-42 km thick (16). In the English River gneiss belt to the north seismic refraction studies indicate thinner crust with an average thickness of 34 km (17). The average depth of the intermediate discontinuity remains about the same (~18 km). In detail, the Moho is upwarped by roughly 8 km in the northern part of the belt, whereas the intermediate discontinuity exhibits a complementary downwarp with an amplitude of 10 km. Re-examination of the orginal data (12) indicates that the axis of this proposed warping lies close to the northern margin of the gneiss belt where it coincides with a sedimentary basin.

Magnetotelluric investigations have been carried out in the western Wabigoon belt (18). A 3.9 km thick near-surface resistive zone under the metavolcanics is considerably less resistive (21,300 Ω-m) than one 7.4 km thick under the granitic gneiss (3,280,000 Ω-m). It suggests that crust underlying metavolcanic rocks is partially fractured and contains saline fluids and/or that the metavolcanics extend throughout the resistive zone. Heat flow studies reported from several Precambrian shields indicate that the average heat flow in greenstones is roughly 10% lower than in crystalline terranes (19). Heat generation data from the Churchill and Superior provinces of the Canadian Shield indicate that

greenstones are approximately 7 km thick.

A general conclusion is that greenstone belts are not rooted in deep crustal structures. Geophysical techniques consistently indicate that greenstones are restricted to the uppermost 10 km or so of crust and are underlain by geophysically normal crust. Gravity models suggest that granitic elements are similarly restricted, although magnetic modelling suggests possible downward extension to the intermediate discontinuity around ~18 km. Seismic evidence demonstrates that steeply-dipping structure, which can be associated with the belts in the upper crust, is not present in the lower crust. Horizontal intermediate discontinuities mapped under adjacent greenstone and granitic components are not noticeably disrupted in the boundary zone. Geophysical evidence points to the presence of discontinuities between greenstone-granite and adjacent metasedimentary terranes. Measured stratigraphic thicknesses of greenstone belts are often twice or more the vertical thicknesses determined from gravity modelling. Explanations advanced for the discrepancy include stratigraphy repeated by thrust faulting and/or listric normal faulting (1), mechanisms which are consistent with certain aspects of conceptual models of greenstone development. Where repetition is not a factor the gravity evidence points to removal of the root zones of greenstone belts. For one region, this has been attributed to magmatic stoping during resurgent caldera activity (20).

Geophysical studies in the Canadian Shield have provided some insights into the tectonic setting of greenstone belts. Much work, however, remains to be done, particularly in the use of geophysics in evolutionary models of greenstone development. Future needs include detailed, integrated studies, the introduction of relatively new methods such as Vibroseis seismic reflection, greater use of magnetotellurics and the application of other electromagnetic methods such as very low frequency (VLF) surveys.

References

(1) Gupta V. K. and Thurston P.C. 1985. Geology (in preparation). (2) Cheadle S.P. 1982. MSc Thesis, Lakehead University. (3) Brisbin W.C. and Green A.G. 1980. C.J.E.S., 17, 968-977. (4) Szewczyk Z.J. and West G.F. 1976. C.J.E.S., 13, 1119-1130. (5) Magnetic Anomaly Map of Canada. 1984. G.S.C. Map 1255A. (6) Hall D.H. 1968. C.J.E.S., 5, 1277-1296. (7) Urquhart W.E.S. 1977. Geotraverse Conference Abstract, 5-7. (8) Losier L. 1984. MSc Thesis, Ecole Polytechnique de Montreal. (9) Parker C.L. 1984. MSc Thesis, McGill University. (10) Green A.G., et al., 1978. C.J.E.S., 15, 301-315. (11) Green A.G., et al., 1979. C.J.E.S., 16, 1599-1612. (12) Hall D.H. and Brisbin W.C. 1982. C.J.E.S., 19, 2049-2059. (13) Clee, T.E., et al., 1974. C.J.E.S., 11, 1534-1549. (14) Gibb R.A. and Thomas M.D. 1980. C.J.E.S., 17, 1506-1516. (15) Hall D.H. and Hajnal Z. 1973. Bull. Seis. Soc. Am., 63, 885-910. (16) Wright J. and West G.F. 1976. Geotraverse Conference, 47-55. (17) Hall D.H. and Hajnal Z. 1969. C.J.E.S., 6, 81-99. (18) Kryzan A. and Strangway D.W. 1977. Geotraverse Conference, 121-131. (19) Drury M. 1985. Abstract (this volume). (20) Gupta V.K. et al., 1982. Precam. Res., 16, 233-255.

VOLCANOLOGICAL CONSTRAINTS ON ARCHEAN TECTONICS P. C. Thurston and L.D. Ayres, Ontario Geological Survey, Toronto, Ont., Canada, M5S 1B3; Dept.of Earth Science, Univ.of Manitoba, Winnipeg, Man., Canada. R3T 2N2.

Terranes within Archean shields can be classified as granite-greenstone megabelts, contemporaneous sedimentary megabelts (1) and basement enclaves within either of the above (2,3). Stratigraphic and geochronological work in the Superior Province has shown the granite-greenstone megabelts represent proximal volcanism, sparse deep water clastic sedimentation, and late alluvial fan-submarine fan sedimentation(4). The sedimentary megabelts represent stratigraphically equivalent deep water sedimentation of wacke-pelite couplets (5), submarine fan conglomerates and minor distal facies volcanism(6). Basement enclaves include meta-igneous and metasedimentary gneiss and fragmented metavolcanic relics with poorly preserved primary textures.(2,7).

Greenstone belts of Australia have been subdivided into >3 Ga platform-phase greenstones and <3Ga rift-phase greenstones (8). Platform phase units are basal komatiite flows and tholeiitic flows with an upper unit of minor pyroclastics. Volcanism in the platform phase is typified by abundant pillowed amygdular flows, overlain by minor airfall tuff and relatively distal debris flow volcaniclastic units. Sedimentary units include chert, quartzite, and stromatolitic carbonates with minor wackes, indicative of shallow water platform sedimentation (9). Examples in the Superior Province, generally about 3 Ga old (10) include quartz-rich wackes in the lower sequence at North Spirit Lake in the Sachigo Subprovince (11), quartzites with fuchsite clasts in the lower sequence of the Wabigoon Subprovince at Armit Lake (12), and carbonate-rich sediments in the Lumby Lake greenstone belt (13) within the Wabigoon Diapiric Axis basement enclave (3). Volcanologically one can conclude from the thickness of the shallow water volcanic rocks and sediments that accumulation took place on a shallow platform (9) and as well, large scale subsidence kept pace with the rate of accumulation of volcanic rocks.

Rift-phase greenstones are relatively deep water amygdule-poor pillowed tholeiites succeeded upward by vesiculated pillowed flows and calc-alkaline pyroclastic and volcanoclastic units (8). Considering the maximum water depth for pyroclastic erutions (14) and the thickness of pyroclastic sections in many rift-phase greenstone belts, Ayres (15) has suggested many Plinian eruption columns became subaerial. Classically (15) most Archean pyroclastic units were considered to have been deposited subaqueously. Recent studies have shown however that many Archean pyroclastic units were deposited subaerially (16,15). Sedimentologic studies of rift-phase greenstones show some deep-water clastic deposits(17), but increasingly shallower water deposits (alluvial fan) at stratigraphically high levels. The structural pattern in rift-phase belts is alternating synclinoria and anticlinoria either breached by diapirism or sheared out (8). Most Superior Province greenstone belts younger than 2.9-2.8 Ga (18) are probably rift-phase based on the following. a)structural style with synclinoria dominating with only rare dome and basin structures. b) Quartz-rich and carbonate rich sedimentation is scarce in the Abitibi (4), kWabigoon(20), and younger (<2.9Ga) sequences of the Uchi (18) and Sachigo (11,18) Subprovinces. c) Volcanism is typified by bimodal tholeiite-rhyolite sequences (21) with shoaling upward attributes (18,4). Evidence for small scale operation of rift-related volcanism is seen in the Six Mile Lake cycle at Sturgeon Lake (22) where a tholeiitic basalt-calcalkaline rhyolite sequence 2755 Ma (23) is rich in incompatible elements relative to later sequences and is cut by abundant mafic dikes assumed to feed the uounger (2718Ma) (24) cycles related to wide-scale rifting.

A survey of volcanic cyclicity (20) reveals the following types of cycles within the Superior Province. (+=fractionation relation; -=no fractionation KOM =komatiitic; TH=tholeiitic; CA=calc-alkaline; ALK=alkaline magma clans

1)KOM Perid Kom +dacite

2)KOM Perid.kom - TH bas + rhy - -CA bas + rhy -ALK

3)TH bas + andes - Th andes - Ca dac +rhy

4)TH bas + andes - Ca bas + rhy - ALK

5)CA bas + rhy

6)TH bas - CA dac + rhy - THbas

Increasing stratigraphic height to the right in each entry.

Cycle types 3,4, and 5 above were formerly thought to represent fractionation sequences,but recent work has shown that many are bimodal(21). The fact that the above cycle types are bimodal has profound volcanologic and petrogenetic implications in that the bimodalism is not simply the paucity of intermediate composition magmatic liquids. Trace element geochemistry and field evidence suggests, when corrected for unerupted volume in zoned magma

chambers,and loss of vitric fines in high level winds during Plinian eruptions are made, preserved volumes of felsic volcanics in the Archean represent ± 15% of the original felsic magma (21). In effect, we concluded that Archean bimodal volcanism represents subequal volumes of mafic and felsic magma which are involved in greenstone belt volcanism.

Determination of paleoenvironment (above), eruption type, eruption rate, magma chamber size and type, developmental processes, and the life span of individual volcanoes places many genetic constraints on greenstone belt tectonics. In mafic sequences subequal volumes of pillowed and massive flows (18) suggest eruption by sheet flow processes (25) dominate over eruption from shield volcanoes (18). In felsic sequences the volumetric dominance of ignimbrites (21) and the notion that sedimentary basins contain large amounts of tephra suggest Plinian eruptions were dominant in the Archean. Many Plinian eruptions produced subaerial deposits on local volcanic islands (18.19,15). Vulcanian eruptions are subordinate, they produce less widespread deposits - examples include the Skead Group (26) and the Lake of the Woods area (27). This eruption type is often the result of less volatile-rich magmas relative to Plinian systems (28) interacting with near-surface water. The deposits are generally less widespread in extent than many Plinian deposits.

Eruption rates of Archean volcanoes can be determined in an approximate and indirect fashion. Sheet flows (25) a greater mean flow thickness than in Phanerozoic analogues (18) and the presence of lava plains (29) in Archean mafic sequences suggest a more rapid eruption rate than in Phanerozoic analogues (30). Phanerozoic ignimbrite systems have volumes in the 10^1-10^2 km^3 range (31) with exceptional examples in the 10^3-10^4 km^3 range (31,32). Phanerozoic felsic volcanoes had a life-span generally not exceeding 1.5 Ma (18) but many Archean felsic edifices apparently existed for 10-20 Ma(18).

The preserved volume of felsic ignimbrites (recalculated to compensate for unerupted material and loss of vitric fines, but ignoring compaction) suggests existance of felsic magma chambers on the order of 10^3 km^3 (21) rivalling those of the largest Phanerozoic systems (28,29). When integrated with data on the lifespan of Archean volcanoes of 10-20 Ma, Archean felsic eruption rates were large, but not as large as those seen in Archean mafic systems.

Volcanological and trace element geochemical data can be integrated to place some constraints upon the size, character and evolutionary history of Archean volcanic plumbing, and hence indirectly, Archean tectonics. The earliest volcanism in any greenstone belt is almost universally tholeitic basalt. Archean mafic magma chambers were usually the site of low pressure fractionation of olivine, plagioclase and later Cpx± an oxide phase during evolution of tholeitic liquids (33 and references therein). Several models suggest basalt becoming more contaminated by sial with time (33,34). Data in the Uchi Subprovince shows early felsic volcanics to have fractionated REE patterns (33)followed by flat REE pattern rhyolites. This is interpreted as initial felsic liquids produced by melting of a garnetiferous mafic source followed by large scale melting of LIL-rich sial (33). Rare andesites in the Uchi Subprovince are produced by basalt fractionation, direct mantle melts and mixing of basaltic and tonalitic liquids(33). Composite dikes in the Abitibi Subprovince (35) have a basaltic edge with a chill margin, a rhyolitic interior with no basalt-rhyolite chill margin and partially melted sialic inclusions. Ignimbrites in the Uchi (16) and Abitibi (36) Subprovinces have mafic pumice toward the top. Integration of these data suggest initial mantle-derived basaltic liquids pond in a sialic crust, fractionate and melt sial. The inirial melts low in heavy REE are melts of mafic material, subsequently melting of adjacent sial produces a chamber with a felsic upper part underlain by mafic magma.
Compositional zonation of the overlying felsic magma develops with time (31), resulting in Plinian eruption through roll over (37) or volatile supersaturation(38).

Numerous arguments suggest widespread volcanism-related subsidence kept pace with the rate of eruption: a) The preservation of felsic sequences rather than the rapid erosion common in Phanerozoic terranes (39) b)Minimum water depth for pyroclastic activity (14) vs preserved stratigraphic thickness of subaqueous pyroclastic units (15)i.e. sections are much thicker than maximum water depth

for eruption - therefore subsidence occurred. c) Lateral extent of 30-50 km for stromatolitic carbonates (40) in the Uchi subprovince, lateral extent of 30-50 km for shallow water silicified evaporites (41) and lateral extent and high eruption rate for shallow water environment mafic plains would have rapidly become subaerial unless subsidence kept pace(18). Isostatic calculations (42,43) suggest lava plain eruptions produce lesser crustal loading than central vent eruptions and less isostatic subsidence. Models involving sialic substrate to lava plain systems produce (42) sufficient subsidence to just maintain volcanic piles at sea level. Therefore we conclude a) subsidence kept pace with volcanism, b)subsidence was regional in extent, c) it is difficult to envision a sagduction style of subsidence (44) producing subsidence over a large area consistent with the great areal extent of the main contributor to the subsidence- the mafic lava plains. Subsidence was more rapid during mafic volcanism slowing during felsic volcanism.

The great volumes of Archean rhyolites and bimodal nature of rift-phase volcanism mitigates against an island arc or back-arc basin analogue where rhyolite is scarce (39 and references therin). Both continental arcs and continental rifts have sufficient volumes of felsic volcanism to compare to greenstone belts. The sediment-filled grabens associated with the Rio Grande Rift (45) offer a possible modern analogue as do the continental intra-arc depressions (39).

REFERENCES

1)Young GM 1978, Geosci. Can.5:140-149.
2)Beakhouse GP 1985, Geol Assoc Can Spec pap 28:169-178
3)Thurston PC and Davis DW 1985, OGS MP 126
4)Dimroth et al 1982, Can J Earth Sci 19:1729-1758
5)Thurston PC 1985,OGS GR 234
6)Bowen RP 1984 OGS OFR 5471
7)Thurston PC and Davis DW OGS MP 126
8)Groves DI and Batt WD 1984 Archaean Geochemistry, Springer 73-98
9)Eriksson KA 1981 Geol Soc Aust Spec Pub 7:235-244
10)Corfu et al 1984 Geol Ass'n Can Abst 9:54
11)Wood J 1980 Prec Res 12:227-255
12)Sutcliffe RH 1985 pers comm
13)Jackson MC 1984 OGS OFR 5534
14)McBirney AR 1963 Bull Volc 26:455-469
15)Ayres LD 1982 Geol Ass'n Can Short Crs Notes 2:1-17
16)Thurston PC Prec Res 12:79-98
17)Ojakangas RW 1985 Geol Ass'n Can Spec Pap 28:23-48
18)Ayres LD and Thurston PC 1985 Geol Ass'n Can Spec Pap 28:343-380
19)Dimroth et al 1983 Can J Earth Sci 20:1374-1388
20)Blackburn et al 1985 GeolAss'n Can Spec Pap 28:89-116
21) Thurston et al 1985 Geol Ass'n Can Spec Pap 28:7-21
22)Trowell NF 1983 OGS GR 221
23)Davis DW 1985 pers comm
24)Davis DW and Trowell NF 1982 Can J Earth Sci 19:868-877
25)Greeley R 1982 J Geophys Res 87:2705-2712
26)Jensen LS 1985 Geol Ass'n Can Spec Pap 28:65-88
27)Car D 1976 Ctr Prec Stud U of Man. 89-96
28)Fisher RV and Schminke H-U 1984 Pyroclastic Rocks, Springer
29)Dimroth E et al 1984 Geol Ass'n Can Spec Pap 28:203-222
30)Wadge G 1982 J Geophys Res 87:4035-4049
31)Smith RL 1979 Geol Soc Amer Spec Pap 180:5-27
32)Lipman PW 1984: J Geophys Res 89: 8801-8841
33)Thurston PC and Fryer BJ 1983 Contr Min Pet 83:204-226
34)Gelinas L and Ludden JN 1984 Phys Earth & Planet Int 35:77-88
35)Gibson H 1983 pers comm
36)Thurston PC 1985 OGS MP 126
37)Rice A 1981 J Geophys Res 86:405-417
38)Sparks RSJ et al 1977 Nature 267:315-318
39)Hildebrand RS and Bowring SA 1984 Geology 12:73-77
40)Hofmann et al 1985 Geol Ass'n Can Spec Pap 28:125-132
41)Fisher GL and Lowe DR GSA Abst 17:1: 583
42)Dimroth E et al 1984 CIMM Spec Vol 34:73-91
43)van Wagoner N et al 1985 McGill U Absts 7
44)Goodwin AM 1982 Can J Earth Sci 19:1140-1155
45)Chapin CE 1979 Rio Grande Rift: Tectonics & Magmatism 1-5

THE DEHYDRATION, REHYDRATION AND TECTONIC SETTING OF GREENSTONE BELTS IN A PORTION OF THE NORTHERN KAAPVAAL CRATON, SOUTH AFRICA; D.D. van Reenen, J.M. Barton Jr., C. Roering, J.C. van Schalkwyk, C.A. Smit (Dept. of Geology, RAU, P.O. Box 524, Johannesburg 2000, S. Africa) J.H. de Beer (NPRL, CSIR, P.O. Box 395, Pretoria 0001, S. Africa) and E.H. Stettler (Geological Survey, Private Bag X112, Pretoria 0001, S. Africa).

High-grade gneiss terranes and low-grade granite-greenstone terranes are well known in several Archaean domains. The geological relationship between these different crustal regions, however, is still controversial. One school of thought favors fundamental genetic differences between high-grade and low-grade terranes while others argue for a depth-controlled crustal evolution. The detailed examination of well-exposed Archaean terranes at different metamorphic grades, therefore, is not only an important source of information about the crustal levels exposed, but also is critical to the understanding of the possible tectonic and metamorphic evolution of greenstone belts with time.

Many features of a metamorphic and deformational transition from a typical greenschist facies granite-greenstone terrane to a high-grade gneiss terrane are illustrated in the crustal section of the northern portion of the Kaapvaal Craton over the 60 km between the Pietersburg greenstone belt and the granulite facies Southern Marginal Zone of the Limpopo Belt. In this section, steeply-dipping, typical greenstone belt lithologies occur at higher and higher grade moving from south to north. In the south, the Pietersburg belt comprises an at least 3450 Ma mafic, felsic and ultramafic volcanic and volcano-sedimentary assemblage (the Pietersburg Group) unconformably overlain by a sedimentary assemblage (the Uitkyk Formation), probably deposited between about 2800 Ma and 2650 Ma. The Pietersburg Group is surrounded by the approximately 3500 Ma tonalitic and trondhjemitic Baviaanskloof Gneiss and is intruded by the approximately 2800 Ma Hout River Gneiss. The belt is apparently intruded by approximately 2650 Ma, largely undeformed, granodioritic plutons. Metamorphic grade within the Pietersburg belt increases from greenschist facies in the southwestern and central parts to amphibolite facies in the northeast, consistent with the regional metamorphic pattern.

North of the Pietersburg belt are situated mafic, felsic and ultramafic volcanic and sedimentary rocks of the Rhenosterkoppies greenstone belt. The Sutherland greenstone belt, composed of mafic and ultramafic and mafic-to-felsic volcanic rocks and sedimentary rocks, occurs to the east. Both belts are surrounded by the Baviaanskloof Gneiss. The ages of the lithologies within the Rhenosterkoppies and Sutherland belts are unknown, but both have been metamorphosed under amphibolite facies conditions.

In the Southern Marginal Zone, highly attenuated and boudinaged granulite facies greenstone belt lithologies (mafic, ultramafic and metapelitic gneisses and banded iron formation) occur within the Baviaanskloof Gneiss. These assemblages are intruded by the approximately 2650 Ma deformed Matok charnockitic-granodioritic pluton while the undeformed Palmietfontein granite was emplaced at about 2450 Ma.

The transition from the lower-grade granite-greenstone terrane to the Southern Marginal Zone is not only refelcted by an increase in the grade of metamorphism but also by an increase in the intensity of deformation. The structural grain of the entire area trends roughly east-northeast with an almost vertically dipping schistosity or gneissosity. In the Southern Mar-

ginal Zone, the distended nature of the granulitic greenstone remnants is in sharp contrast to the more continuous outcrop pattern of the greenstone lithologies to the south. The magnetic lineation patterns of the area change abruptly at the boundary of the granulite facies terrane and seismic velocities increase sharply at the same point, possibly as a result of mantle material being at a higher level.

The crustal behavior of the entire region must have been consistent with the observation that the rocks of the Southern Marginal Zone were depressed into deep crustal levels. This movement probably implies that the lower-grade terranes were depressed in a sympathetic manner. In the Southern Marginal Zone, the maximum prograde conditions (P>9.5 kb and T>800°C) reached during this tectonic event are recorded by the assemblage garnet + hypersthene + quartz + plagioclase +/- biotite +/- kyanite in metapelite. These conditions were followed by rapid, nearly isothermal, decompression between approximately 2700 Ma and 2650 Ma, recorded by decompression textures of cordierite and hypersthene after garnet. P-T conditions of this decompression event were T=800°C with P decreasing to 7.0 kb. The Matok pluton was emplaced toward the end of the isothermal decompression. The southern margin of this dehydrated terrane was then subjected to a regional encroachement of CO_2-rich hydrating fluids before approximately 2450 Ma, the time of emplacement of the Palmietfontein granite. This infiltration produced the retrograde orthoamphibole isograd defined by the reactions: hypersthene + quartz + H_2O = anthophyllite and cordierite + H_2O = gedrite + kyanite + quartz. These reactions occurred at T=650°C to 600°C and a total P less than 6 kb at $PH_2O = 0.2\ P_{total}$. Completely hydrated and recrystallized rocks south of this isograd are characterized by the assemblage anthophyllite + gedrite + kyanite + biotite + quartz + plagioclase. The fluids responsible for rehydration are believed to have been derived from hydrated granite-greenstone lithologies.

Metamorphic assemblages in the area south of the retrograde isograd are still insufficiently documented to delineate isograds but the overall increase in the grade of metamorphism from south to north is illustrated by comparing assemblages in chemically similar granite-greenstone lithologies from different crustal levels:

LITHOLOGIES	PIETERSBURG BELT (Central/South West part)	RHENOSTERKOPPIES AND SUTHERLAND BELTS	SOUTHERN MARGINAL ZONE OF LIMPOPO BELT
Meta-sediments	Qz + Chl + feldsp + sericite ± Carb	Orthoamph + Biot + Qz + Plag + Garn ± Kyan	Hyp + Biot + Qz + Plag + Garn + cord
Mafic Rocks	Ab + Ep + Chl + Act + Qz ± Carb	Hbl + Plag (An 33) + Qz ± Sf ± Diop	Opx + Cpx + Plag (An 49) + Mt ± Hbl ± Qz
Ultramafic Rocks	Trem + Chl ± Talc ± Carb	Trem + Chl ± Ol ± Carb	Ol + Opx + Sp + Ca-Amph

Gravity and resistivity data indicate that the Pietersburg, Rhenosterkoppies and Sutherland greenstone belts are shallow crustal features, rarely exceeding 5 km in depth. This depth is in marked contrast to the thickness of the various lithologic successions measured across the stratification. These observations plus the fact that the crustal thicknesses determined from stratigraphic sections are far in excess of those which can be accepted from the metamorphic data, suggest that major crustal thickening took place in this area by tectonic stacking. A steeply northward dipping shear zone with near vertically plunging mineral lineations is exposed along the Hout River. This shear zone is believed to be associated with the tectonic stacking but, as yet, the sole thrust for this proposed stacking has not been recognized. The hydrating fluids responsible for establishing the retrograde orthoamphibole isograd could have been derived by dehydration of over-ridden lower-grade crustal rocks during and after thrusting.

THE STRATIGRAPHY OF THE STEEP ROCK GROUP, N.W. ONTARIO, WITH EVIDENCE OF A MAJOR UNCONFORMITY

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The Steep Rock Group is exposed 6 km north of Atikokan, 200 km west of Thunder Bay. It is situated on the southern margin of the Wabigoon Belt of the Archaean Superior Province, N. W. Ontario. Reinvestigation of the geology of the Group has shown that the Group lies unconformably on the Tonalite Complex to the east.

This unconformity has been previously suspected, from regional and mine mapping but no conclusive outcrop evidence for its existence has as yet been published.

The strike of the Group, comprised of Basal Conglomerate, Carbonate Member, Ore Zone and Ashrock is generally north-northwest dipping steeply to the southwest. Of the 7 contacts between the Steep Rock Group and the Tonalite Complex, 3 expose the unconformity (The Headland, S. Roberts Pit, Trueman Point), and 4 are faulted.

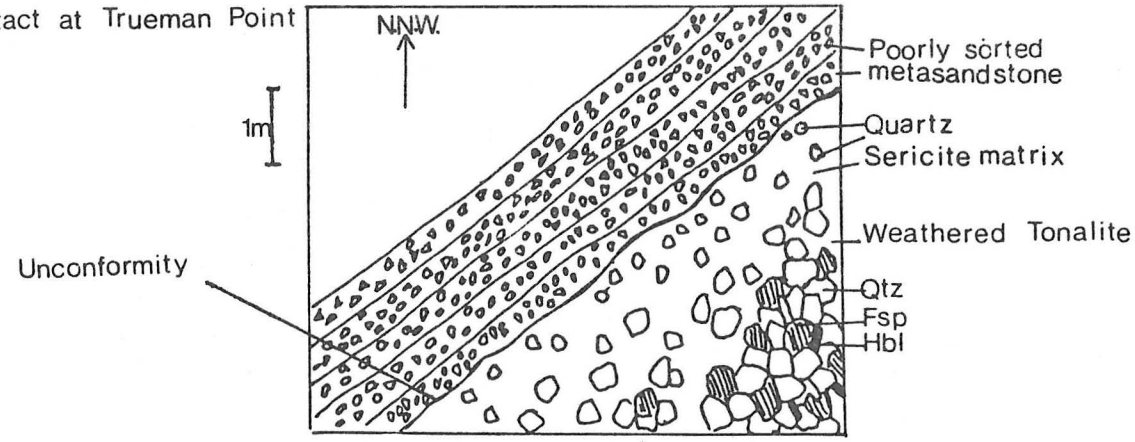
At the Headland poorly sorted metaconglomerate with angular clasts of quartz, tonalite and fine grained mafic material (dykes and remnant xenoliths) overlies mafic tonalite, with no evidence for a fault or an intrusive contact.

At the S. Roberts Pit, poorly sorted metasandstones dip steeply to the west overlying pale greenish white weathered mafic tonalite. The metasandstones pass upwards within 20 cm to massive dark grey carbonate.

At Trueman Point, in an exposure similar to the S. Roberts Pit, coarse angular metasandstone overlies tonalite. However, the contact here is more diffuse with the top metre of the tonalite breaking down to form a regolith of angular quartz grains (1-4mm) in a sericite matrix. This matrix is similar to the overlying metasandstone.

These three outcrops demonstrate unequivocally that the Steep Rock Group was laid down unconformably on the underlying Tonalite Complex, which is circa 3 Ga old (Davis, pers. comm.).

Contact at Trueman Point



THE STRATIGRAPHY OF THE STEEP ROCK GROUP

Wilks, M.E. and Nisbet, E.G.

Overlying the Basal Conglomerate (0-150m) is the Carbonate Member (0-500m) throughout which stromatolites extensively occur.

The carbonate is a laminated dark bluish-grey massive rock with major zones of breccia developed close to fault zones and dykes, which are thought to be feeders for the overlying volcanics.

From a study of 11 good stromatolitic outcrops a crude stratigraphy of the carbonate can be set up. Small scale stromatolites occur throughout the unit, but are best developed near the base. Here simple Stratifera-like stratiform structures having flat to undulatory laminae develop into pseudocolumnar laterally-linked structures. These Irregularia-like structures pass upwards into hemispherical laterally linked stromatolites. Laminae are wavy .5-3.5mm, and the structures are 5-15 cm high and in basal diameter. In places branching walled and unwalled columnar forms occur, with height up to 20 cm.

In the upper part of the Carbonate giant domal stromatolites occur. These range from domed structures typically about 3m in diameter to tabular bodies up to 5m or more long and .75m in stratigraphic height.

Near the top of the unit, small mamillose stromatolites form an egg box fabric with diameters up to 4cm and heights of 1.5cm. Overlying the Carbonate Member is the Ore Zone which Jolliffe[1] divided into a lower Mn Paint Rock and an upper Goethite Member.

The Mn Paint Rock (3%-18% Mn) is an earthy material with a rude varicoloured banding, made up of lumps of goethite, hematite, quartz and chert in a groundmass of the same minerals with calcite, kaolinite, pyrolusite and gibbsite. The contact with the underlying carbonate is extremely irregular, with pinnacles of carbonate protruding into the Paint Rock. This contact has been interpreted as an ancient karst surface by Jolliffe[1].

The Mn Paint passes sharply upwards into the Goethite Member (Mn < .3%) which is predominantly brecciated lump ore of goethite (67%) and haematite (21%) with quartz and kaolinite.

Within the Ore Zone thin layers of Buckshot Ore occur. These layers comprise haematitic pisolites and fragments of haematite set in a lighter aluminous matrix of kaolinite and gibbsite. This material resembles a ferruginous bauxite in both outward appearance and chemical and mineral content.

Overlying the Ore Zone is the Ashrock. The name refers to an ultramafic pyroclastic rock (22% MgO) which makes up to 90% of the unit. Interbedded within this are thin komatiitic basalt (15% MgO) spinifex-textured lava flows.

Within the Goethite Member and Ashrock, pyrite lenses occur. These form discontinuous elongate bodies of massive pyrite closely associated with cherty and carbonaceous beds.

The Ashrock delineates the upper member of the Steep Rock Group.

References:

[1] Jolliffe, A.W. (1955) in Econ. Geol., vol.50, p. 373-398.

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WORKSHOP ON THE TECTONIC EVOLUTION OF GREENSTONE BELTS

(Supplement Containing Abstracts Of Invited Talks and Late Abstracts)

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GREENSTONE BELT TECTONICS - THERMAL CONSTRAINTS: M.J. Bickle (Dept Earth Sciences, University of Cambridge, Cambridge CB2 3EQ, U.K.) and E.G. Nisbet (Dept Geological Sciences, University of Saskatchewan, S7N 0W0, Canada).

Archaean rocks provide a unique record of the early stages of evolution of a planet. Their interpretation is frustrated by the probable unrepresentative nature of the preserved crust and by the well known ambiguities of tectonic geological synthesis. Broad constraints can be placed on the tectonic processes in the early earth from global scale modelling of thermal and chemical evolution of the earth and its hydrosphere and atmosphere. The Archaean record is the main test of such models. It is the purpose of this contribution to outline what general model constraints are available on the global tectonic setting within which Archaean crust evolved, and what direct evidence the Archaean record provides on particularly the thermal state of the early earth.

The distinct tectonic style of Archaean granite-greenstone terrains undoubtedly reflects secular variation in the earth's tectonic processes as a result of chemical and thermal evolution. Since tectonic processes are a direct manifestation of heat loss processes in the earth, changes in the earth's thermal state are likely to be primarily responsible for changes in tectonic style. However, the geological record of tectonic processes is also influenced by the state of chemical evolution of the solid earth and its hydrosphere and atmosphere. As discussed below the basic volcanic dominated nature of greenstone belts is probably as much a consequence of higher mantle temperatures as any specific tectonic setting. Until proved otherwise we must assume that 'greenstone belts' formed in as wide a range of tectonic environments as modern sedimentary sequences. Care must be taken to distinguish features which are due to a specific tectonic environment from those indicative of general tectonic processes in the Archaean earth.

Global Thermal Histories

Calculations of global thermal evolution are based on derivations of relationships between internal temperature and heat loss. Given such a relationship and the present temperature and radiogenic heat producing element distribution within the earth it is possible to calculate temperature distributions in the past with the assumption that the heat loss processes (convection) varied only in rate throughout earth history. Most current models are formulated to satisfy the cosmochemical constraint that present day radiogenic heat production produces about half of the total heat loss and that the earth was hot soon after accretion [e.g. 1]. The main area of uncertainty intrinsic in the modelling is the treatment of convection in a fluid of temperature sensitive and non-Newtonian viscosity. One set of models, the 'parameterised' convection calculations, derives a relationship between internal temperature and heat loss by computing heat loss as a function of viscosity for a series of models run with internally constant but differing viscosities and assuming some form for the viscosity temperature dependence. Implicit in such modelling is the assumption that convection in a variable viscosity fluid can be approximated by a constant viscosity appropriate to a characteristic temperature within the system. However, as first demonstrated by McKenzie and Weiss [2] the assumptions of parametrical convection calculations are not appropriate to convection in variable viscosity fluids. Christensen [3] points out

that it is the lower-temperature higher-viscosity upper boundary layer that dominates convection rates and if heat loss should scale against any internal temperature it will be a temperature within the upper boundary layer rather less than the interior temperature (or correctly interior potential temperature which is mantle temperature normalised along an adiabatic gradient to zero pressure). It is the interior temperature which is used for scaling by the parameterised calculations. The difference may be illustrated by comparison of temperature - heat loss relationships.

The parameterised calculations lead to an expression for the relationship between the Nusselt number Nu (the total heat flux to conducted heat flux) and the Rayleigh number Ra of the form

$$Nu \propto Ra^{\beta}$$

where β is between $\frac{1}{4}$ and 1/3. This relationship determines the temperature sensitivity of the heat flux.

Christensen's calculations with variable-viscosity fluids suggest that values of β around 0.05 are more appropriate over the limited range of the experiments. The real uncertainties are rather greater than this given the possibility of a layered mantle, two scales of convection in the upper mantle, partition of heat loss between oceanic and continental regions and melting with associated density changes with the upper thermal boundary layer. Substantial deviations in tectonic style from modern plate-tectonics could further influence heat loss.

Two important conclusions may be deduced from the thermal modelling.

1. The parameterised calculations with the assumption that mantle viscosity is independent of temperature [4, Fig. 1] probably provide a realistic upper bound on a relationship between temperature and heat loss, indicating that interior temperatures have not changed by more than a few hundred degrees over most of earth history.

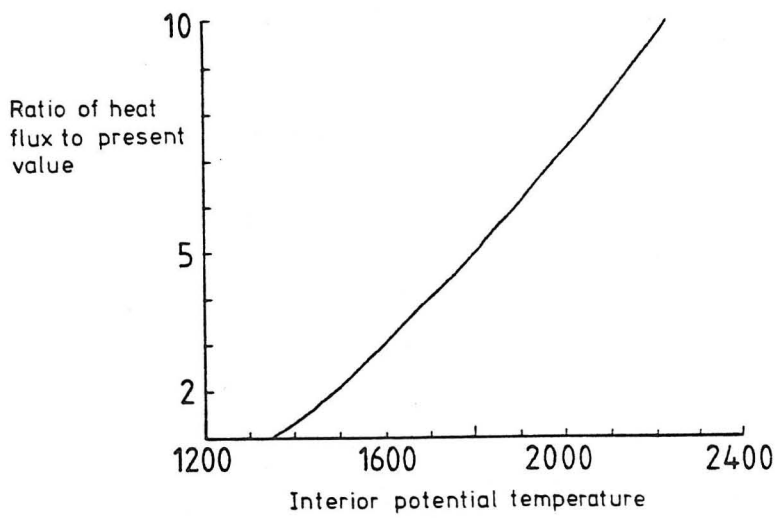

Figure 1. Variation of average upper mantle potential temperature with heat flux. Viscosity independent of temperature from [4]. This represents a plausible upper bound on average mantle temperatures variation.

2. All the models predict that higher internal temperatures result in thinner, higher thermal gradient boundary layers (Plates)[1,3]. Further constraints must come from Archaean geology, which provides evidence on two critical parameters, upper mantle temperatures and continental lithospheric thermal gradients.

1. Mantle Temperatures

The presence in Archaean greenstone belts of komatiitic lavas more magnesian than any younger lava is one of the few distinctive features of the Archaean and prime evidence that mantle temperatures were higher. To quantify the difference we need to know (1) the eruption temperature of komatiites and (2) the relationship between komatiite eruption temperatures and mantle temperatures. The first question has provoked surprisingly little discussion given its significance [e.g. 5,6]. Liquidus temperatures of komatiitic lavas are proportional to MgO content but this may be increased by olivine accumulation. Glassy, near phenocryst free lavas [7], and relict forsterite-rich olivine compositions have been taken to indicate liquids at least as magnesian as 27-30% MgO [5] although this is disputed [6]. Alternatively excess H_2O or alkalis have been suggested as fluxes lowering liquidus temperatures [e.g.8]. The latter is potentially testable through the temperature dependence of Ni olivine:liquid partition coefficients although such systematic tests have not been made. Even so eruption temperatures of ~1500°C (25% MgO) to ~1600°C (30% MgO) are 100-200°C hotter than any more recent lava.

The relationship between komatiite temperature and mantle temperature is more problematic. Adiabatically upwelling mantle cools along substantially higher thermal gradients (higher dT/dP) above the solidus as a result of the latent heat of melting (Fig. 2). If komatiites represent ~50% melts at high

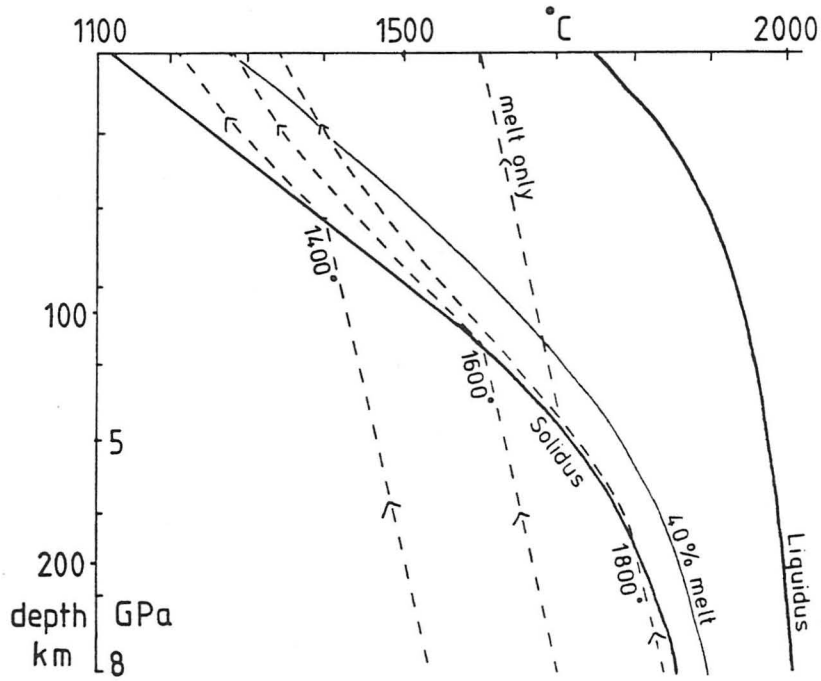

Figure 2. Mantle liquidus and solidus and adiabatic ascent paths calculated with the assumption that melt and solid do not segregate on ascent, after McKenzie and Bickle [23].

level with an olivine residue then a 1600°C komatiite must be derived from a mantle in excess of 2000°C from depths >300km where we are essentially ignorant of mantle solidus-liquidus temepratures. Such temepratures substantially exceed the upper bound of mantle temperatures derived from global thermal modelling. Alternatively it has been suggested that eutectic melts at high pressure shift to komatiite compositions [9]. Available phase equilibrium data suggests this might be in the region 50-100 kbar [Fig.3]. If so komatiites might be derived from mantle temperatures of 1800°C-1900°C, a potential temperature of 1700-1800°C, and 400-500°C hotter than present day average mantle. If komatiites are derived from anomalously hot upwelling convective instabilities the potential temperatures of such regions are 200°C-300°C hotter than mantle in present day thermal plumes.

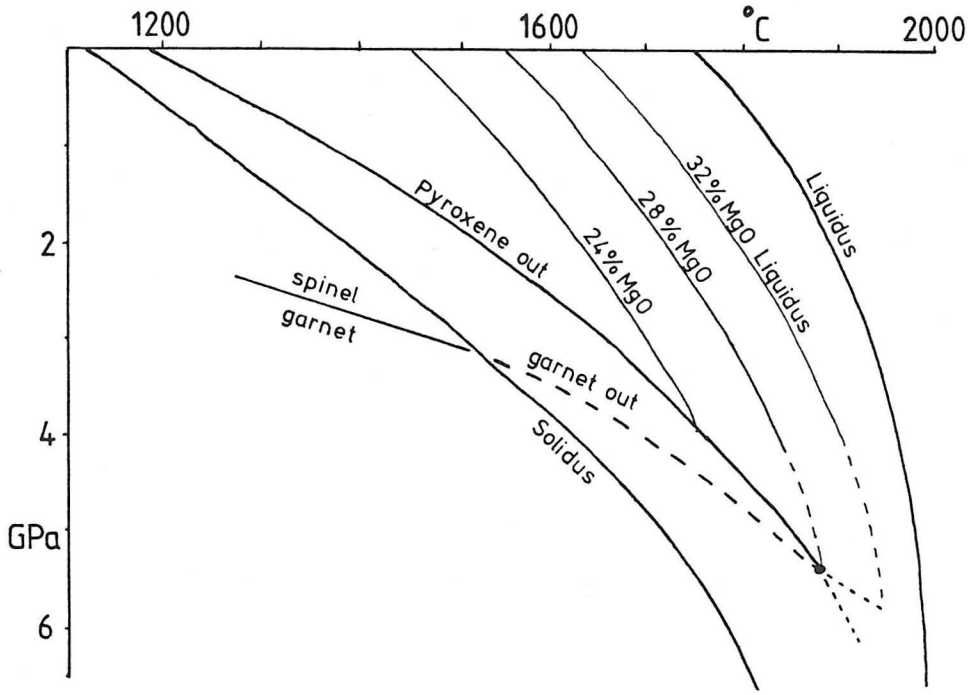

Figure 3. Phase relations for melting mantle-like compositions from experiments on komatiites. Note intersection of garnet melting curve with the pyroxene melting curve is hypothetical.

The chemistry of komatiites is not obviously reconcilable with their being small degrees of eutectic melts. Incompatible element concentrations are surprisingly uniform and are consistent with komatiites being ~50% melts of plausible mantle materials [10,11]. Small degrees of melt would be expected to be substantially enriched in incompatible elements although partition coefficients at the pressures of komatiite genesis are unknown and substantial modifications to komatiite chemistry by wall rock interaction might be expected during their ascent [12].

Komatiite genesis is therefore problematic. However, even the most conservative estimates of komatiite eruption temperatures (a 25% MgO 1500°C lava) implies mantle potential temperatures ~200°C hotter than at present and a 30% MgO, 1600°C lava is inferred to imply mantle potential temperatures ~400°C greater than today. One further complication is the possibility that at high pressure the komatiite melt density exceeds that of solid mantle. If the inversion in density is associated with a change in sign of the pressure

derivative of the potential temperature on the melting curve existence of a stable magma ocean at depth is probable [13]. The implications of such a magma ocean for global thermal and chemical evolution are profound.

2. Crustal Thermal Gradients

Metamorphic pressures and temperatures record anomalous thermal conditions in tectonically active crust. If sufficient is known about the tectonic setting of the metamorphism it is possible to invert the perturbed thermal conditions to infer steady state lithospheric thermal gradients [14]. Models for such inversion are mostly based on the thermal time constant over lithospheric thicknesses being rather greater than that of tectonic events ($\sim<50$ Ma). Given the possibility of magmatic or fluid heat transfer, such models tend to put upper bounds on lithospheric thermal gradients.

Archaean metamorphic conditions exhibit as wide a range of thermal gradients as modern orogenic provinces. High thermal gradients may at least locally be associated with magmatic advection of heat [15]. The lower thermal gradient, higher P/T metamorphism has attracted most interest as it places limits on the magnitude of lithospheric thermal gradients. The widespread 8-10 Kb, 700°C-900°C conditions recorded by gneiss terrains [16] imply background gradients little different from those in modern continental lithosphere. However, Morgan [17] suggests that these metamorphic conditions are buffered by crustal melting and heat flow in these regions is underestimated. Comparable high P/T metamorphism is known from upper-greenschist and amphibolite facies Archaean terrains [15,18-20] although it is less well documented. This is inconsistent with high heat flow through the underlying crust and not explicable as buffered by melting.

The inference from the metamorphic conditions of relatively low lithospheric thermal gradients has received substantial support from the observation of the formation and preservation of Archaean age diamonds [21]. These imply lithospheric thicknesses of ~150-200 km and mantle heat flux as low as 20 mWm^{-2}.

The observation that greenstone belts may have formed or been preserved in continental crust with relatively low thermal gradients has far-reaching implications for Archaean tectonics. Study of the metamorphism and its tectonic setting in greenstone belts would seem to be one rather neglected area of greenstone tectonics.

Implications on Global Thermal Evolution

The evidence for a significantly hotter mantle implied by komatiites is irreconcilable with the evidence for a thick cool continental lithosphere if the lithosphere behaved as its modern counterpart. There is good evidence from the depth-age relationships of oceanic lithosphere and sedimentary basin evolution that Phanerozoic oceanic and continental lithosphere behaves as a simple thermal boundary layer. To preserve a similar or greater thickness of Archaean lithosphere requires some additional process to stabilise the continental lithosphere. Morgan [17] suggests that increasing the concentration of radiogenic heat production might achieve this. It might but thermal gradients over such enriched lithosphere would have to be at least as high as those over correspondingly thin but unenriched lithosphere. An alternative mechanism is that the stabilisation results from density changes on melting [e.g. 22]. One consequence of a higher temperature mantle is that melting would start at much greater depths (Fig. 2)(~115 km for a 1600°C mantle versus

~60 km for the present day ~1300°C mantle). The depleted zone is comparatively less dense than unmelted mantle although whether the relatively small changes are sufficient to stabilise the lithosphere against convective instabilities is open to question. The mechanism of stabilisation of Archaean continental lithosphere and the formation and preservation of Archaean diamonds is a key question. It has implications both for Archaean tectonic interpretations as well as subsequent global evolution given the significance of the continental lithosphere to continental tectonics.

There is one further significant tectonic implication of a hotter mantle. The amount of melt produced by upwelling mantle is proportional to mantle temperature [Fig. 4; 23]. With a 1600°C mantle any tectonic activity such as

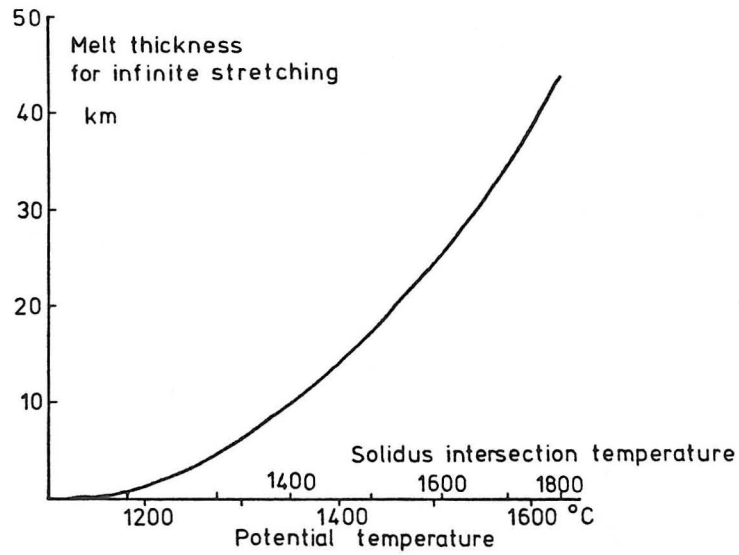

Figure 4. Melt thickness as a function of mantle temperature for infinite stretching (oceanic ridge case) after McKenzie & Bickle [23].

crustal extension which led to mantle upwelling would produce significant magma. It seems probable that the basalt dominated nature of both Archaean greenstone and late Archaean cratonic supracrustal sequences is a reflection of mantle temperature and not necessarily of a special tectonic setting. The extrusion of thick dense basaltic volcanics in supracrustal sequences may be an important factor in the development of the characteristic tectonic style of granite-greenstone terrains.

Archaean Tectonic Regimes

The prime assumption of all the global scale thermal models is that heat loss processes changed only in rate. One hotly debated point is whether plate tectonics or some alternative tectonic scheme operated during the Archaean. For example, Richter [1] has suggested that once convecting mantle penetrated the melt region below continental lithosphere the surface tectonic regime would be dominated by vertical recycling rather than horizontal

motions. This scheme does not explain the preservation of the early Archaean crustal relicts for which some special survival mechanism must be proposed. Perhaps the best evidence for major horizontal (plate) motions lies in the linear tectonic belts characteristic of the larger Archaean terrains (Superior Province, Yilgarn Block) and the evidence for large scale overthrust nappe tectonics in the high-grade gneiss belts. Other geological evidence is open to interpretation. For example, the significance of the calc-alkaline-like granite suites, possible analogies between some greenstone belt mafic sequences and ophiolites and the tectonic state of greenstone belts (allochthonous or authochthonous) are all disputed. One additional line of evidence does strongly suggest division of the Archaean earth into continental and oceanic-type regions. The heat loss through the Archaean continental regions inferred from metamorphic thermal gradients is too low by an order of magnitude to be representative of heat loss from the Archaean earth [24,25]. The extra heat is plausibly lost through oceanic-like regions as is the case today. This would involve substantial melting and recycling of volcanic crust.

References:

[1] Richter, F.M. (1985) Earth planet. Sci. Lett. 73, 350-360.
[2] McKenzie, D.P. and Weiss, N. (1980) Geophys. Jour. Royal astr. Soc. 42, 131-174.
[3] Christensen, U.R. (1985) Jour. Geophys. Res. 90, 2995-3007.
[4] Richter, F.M. (1984) Earth planet. Sci. Lett. 68, 471-484.
[5] Bickle, M.J. (1982) in: Komatiites, N.T. Arndt and E.G. Nisbet (eds.).
[6] Elthon, D. (1986) Komatiite genesis in the Archaean mantle with implications for the tectonics of Archaean greenstone belts (abstract). In Workshop on the Tectonic Evolution of Greenstone Belts. Lunar and Planetary Institute, Houston, in press.
[7] Pyke, D.R., Naldrett, A.J. and Eckstrand, O.R. (1973) Geol. Soc. Am. Bull. 84, 955-978.
[8] Allègre, C.J. (1982) in: Komatiites, N.T. Arndt and E.G. Nisbet (eds.).
[9] O'Hara, M.J., Saunders, M.J., Mery, E.L.P. (1975) in: L.M. Ahrens et al. (eds.), Phys. Chem. Earth 9, 577-604.
[10] Bickle, M.J., Hawkesworth, C.J., Martin, A., Nisbet, E.G., O'Nions, R.K. (1976) Nature, 263, 577-580.
[11] Nesbitt, R.W. and Sun, S.S. (1976) Earth planet. Sci. Lett. 31, 433-453.
[12] Huppert, H. and Sparks, R.S.J. (1985) Earth planet. Sci. Lett. 74, 371-386.
[13] Nisbet, E.G. and Walker, D. (1982) Earth planet. Sci. Lett. 60, 105-113.
[14] England, P.C. and Bickle, M.J. (1984) Jour. Geology 92, 353-367.
[15] Bickle, M.J. and Archibald, N.J. (1984) Jour. Metamorphic Geol. 2, 179-203.
[16] Perkins, D.P. and Newton, R.C. (1981) Nature 292, 144-146.
[17] Morgan, P. (1986) Thermal implications of metamorphism in greenstone belts and the hot asthenosphere - thick continental lithosphere paradox (abstract). In Workshop on the Tectonic Evolution of Greenstone Belts. Lunar and Planetary Institute, Houston, in press.
[18] Boak, J.L. and Dymak, R.F. (1982) Earth planet. Sci. Lett. 59, 155-176.
[19] Wells, P.R.A. (1979) Jour. Petrology 20, 187-226.
[20] Bickle, M.J., Morant, P., Bettenay, C.F., Boulter, C.A., Blake, T.S. and Groves, D.I. (1985) Geol. Assoc. Canada Spec. Pap. 28, 325-341.
[21] Richardson, S.H., Gurney, J.J., Erlank, A.J. and Harris, J.W. (1984) Nature 310, 198-202.

[22] Jordan, T.H. (1979) in F.R. Boyd and H.O.A. Meyer (eds.) Am. Geophys. Union, 1-14.
[23] McKenzie, D.P. and Bickle, M.J. (in preparation).
[24] Bickle, M.J. (1978) Earth planet. Sci. Lett. 40, 301-315.
[25] Burke, K. and Kidd, W.S.F. (1978) Nature 272, 240-241.

THE ROCK COMPONENTS AND STRUCTURES OF ARCHEAN GREENSTONE BELTS: AN OVERVIEW; Donald R. Lowe and Gary R. Byerly, Department of Geology, Louisiana State University, Baton Rouge, LA 70803.

Much of our understanding of the character and evolution of the earth's early crust derives from studies of the rocks and structures in Archean greenstone belts. Our ability to resolve the petrologic, sedimentological, and structural histories of greenstone belts, however, hinges first on an ability to apply the concepts and procedures of classical stratigraphy. Unfortunately, early Precambrian greenstone terranes present particular problems to stratigraphic analysis, some of which we would like to discuss here. We would also argue that many of the current controversies of greenstone belt petrogenesis, sedimentology, tectonics, and evolution arise more from our inability to develop a clear stratigraphic picture of the belts than from ambiguities in its interpretation.

We will here consider four particular stratigraphic problems that afflict studies of Archean greenstone belts: (a) determination of facing directions, (b) correlation of lithologic units, (c) identification of primary lithologies, and (d) discrimination of stratigraphic versus structural contacts.

(a) Facing Directions: Determination of facing directions in greenstone belt sequences is often difficult because of the absence of useful facing indicators throughout great thicknesses of section and because we do not sufficiently understand the origins of many structures and textures in Archean sedimentary rock types to be able to use them as facing indicators. Thick sequences of massive volcanic rocks, banded black and white cherts, black cherts, and banded iron formation are inevitably rather stingy in yielding familiar facing indicators whereas thick turbiditic units, layers of graded accretionary lapilli, and sands containing large-scale cross-stratification are particularly user-friendly in this regard. Facing directions in banded cherty units are most readily determined from fluid escape features, particularly pockets of druzy quartz, which originate as pockets of trapped fluid, usually directly beneath early-lithified white chert bands. Geopetal accumulations of debris in cavities, cracks, and at the bases of early-formed breccias and the preferential development of stalactitic dripstone in stratiform cavities (the development of both stalactitic and stalagmitic dripstone is also common, but stalagmites alone are extremely rare) are also widespread and useful as facing indicators in cherty successions. In all cases where supporting evidence is available in adjacent sedimentary units, we have found pillow geometry and drain-out cavities, where developed, to be reliable facing indicators in tholeiites.

Small-scale cross-laminations, load structures, and individual graded detrital layers must be approached with caution because nearly identical features can form facing upward or downward. Pillows, where present in komatiitic sequences, generally lack useful facing information. The recent trend to quantify the reliability of facing estimates (e.g. 95% confidence) is misleading inasmuch as the principal errors in determining facing directions originate not through statistical ambiguities in the structures themselves but from their misidentification by the investigator.

(b) Correlation: The correlation of stratigraphic units within poorly exposed, structurally complex, highly altered Archean terranes represents a major challenge to unravelling greenstone belt stratigraphy and evolution. The absence of useful guide fossils and the paucity of unique, recognizable

time markers, such as distinctive ash beds, makes this task difficult relative to similar studies in Phanerozoic terranes. Recent precise zircon age dating in the Canadian belts is aiding in resolving gross problems of stratigraphy, but will do little for detailed correlation.

In the early Archean Barberton and Pilbara belts, we have found a number of features particularly useful in correlation: (1) lithologically and texturally distinctive layers of airfall and/or turbiditic accretionary lapilli, (2) individual airfall ash beds in sequences of orthochemical and biogenic deposits, (3) airfall spherule layers, (4) distinctive sequences of non-facies controlled deposits, and (5) rare, facies-related units and sequences. Least reliable are distinctive successions of environmentally or petrogenetically controlled lithologies that can be repeated many times within individual sections as sedimentary environments and magmatic systems come and go. Even continuous, traceable lithologic units cannot serve as unambiguous time markers unless there is independent evidence that they are not diachronous.

(c) Primary Lithologies: Perhaps as much as any other problem, our inability to decipher primary lithologies has hampered the development of a clear picture of greenstone belt make-up and evolution. It has long been recognized that early alteration is pervasive throughout greenstone belts. This alteration was for many years considered part of the post-accumulation metamorphic history of these belts. More recently, however, the trend has been to attribute alteration to relatively high-temperature exhalative to shallow-subsurface hydrothermal processes (1, 2) or to low-temperature metasomatism, perhaps related to the circulation of surficial waters through the rock sequences (3).

Interpretation of the primary MgO contents and petrogenesis of komatiites, role of calc-alkaline and subduction-related volcanism, presence or absence of volcanic cycles, distribution of felsic lavas, nature of metamorphism and metasomatism, provenance of detrital sediments, composition of early surface waters, and sedimentology of cherty units have all been stymied to some extent by uncertainties in the composition of the original sedimentary and volcanic layers. A number of relatively recent studies have shown clearly that (i) many specific units previously interpreted to be silicic volcanic rocks are actually silicified mafic to ultramafic lavas (e.g. 2, 3), (ii) many of the "classic" mafic-to-felsic volcanic cycles are non-existent (4) although large-scale volcanic cyclicity seems to be widely developed (5), (iii) calc-alkaline volcanics, as well as komatiites, are abundant in some belts but poorly represented in others, (iv) some belts exhibit a more-or-less continuous spectrum of rock compositions from komatiitic to rhyolitic whereas others are strongly bimodal or trimodal; (v) evaporitic sediments, especially gypsum, were widespread and abundant constituents of shallow-water Archean greenstone-belt sedimentary deposits (6), (vi) relatively few, if any, cherty layers represent primary silica precipitates (7), and (vii) there may be important lithologic and tectonic differences between early and late Archean greenstone belts (7).

Many of the remaining ambiguities in the alteration histories of these rocks originate because most studies of alteration are focused on identifying the role or evaluating the influence of one particular style or setting of alteration. Clearly, some silicification and carbonatization began concurrently with deposition and involved essentially surface waters at surface temperatures. The abundance of cherts in shallow-water sequences but their paucity in deeper-water units (7) suggests that early post-

depositional fluctuations in water chemistry (e.g. deposition in marine but early flushing by meteoric waters) may have been an important control on silicification. Later large-scale recrystallization and replacement almost certainly occurred both through low-temperature processes, similar to those affecting modern oceanic crust, as well as during local higher-temperature, hydrothermal and black-smoker-type metasomatism and mineralization. The widespread presence of epidote and resetting of isotopic systems, such as Ar-Ar, clearly argue for still later regional metamorphism, and the localization of silicification along some joints and fractures indicates continued alteration under fully post-tectonic and post-metamorphic conditions. Future studies must provide unambiguous criteria for distinguishing stages and environments in this prolonged alteration history, many of which may leave similar mineralogical and textural records.

(d) Stratigraphic vs. Structural Contacts: Greenstone belt sequences are characteristically highly deformed, typically showing polyphase deformation and structural repetition through faulting and folding. One of the principal problems facing structural, stratigraphic, and tectonic synthesis of greenstone belts lies in distinguishing between structural and stratigraphic contacts in areas of poor exposure and in the near-absence of unambiguous tools for relative age determination and correlation. Whereas it was once fashionable to regard thick, apparently intact, uniformly facing successions of volcanic and sedimentary rocks in greenstone belts as forming coherent stratigraphic sections, often in excess of 15 km in thickness, the present tendency is often to infer that such sequences, at least on this planet, are composite, formed by the tectonic repetition of considerably thinner stratigraphic sections.

The problem, now as previously, is the field recognition of faults, particularly stratiform faults, such as thrusts. In the Barberton belt, for instance, there are large areas, particularly in upper parts of the succession, within which nearly stratiform thrust faults are present and can be easily recognized using conventional means: (1) truncated and offset stratigraphic units and folds, (2) unambiguously repeated stratigraphic sequences, (3) the development of mylonitic and brecciated zones along fault planes, and (4) the formation of drag folds in units adjacent to the faults. However, throughout most of the classic sections of the Onverwacht Group in the southern part of the belt, major faults identifiable by such conventional criteria are absent. Although it has been suggested that most of the apparent 12-km thickness of the Komati, Hooggenoeg, and Kromberg Formations is an artifact of isoclinal folding of a much thinner sequence (2), studies of facing directions throughout the section do not bear out this interpretation (3). Arguments have also been advanced (2, DeWit, this meeting) that chrome-mica-bearing alteration zones at the tops of komatiitic units within this sequence represent stratiform shear zones with displacements of perhaps 1-10 km. Unfortunately, however, these units display none of the usual characteristics of faults (such as cross-cutting relationships) and are developed only at the tops of komatiitic flows (never at the tops of tholeiitic of felsic units). They exhibit cataclasis and schistosity only where cross-cut by clearly later, through-going faults or where present in areas where all units show penetrative deformation. In most sections, these rocks display well-preserved, unsheared primary spinifex and cumulate textures. Inferences that these zones represent faults must at some point be based on a systematic consideration of their characteristics, including clear enumeration of features indicating an

origin through faulting and the means of determining displacement.

Although it is clear that our ability to unambiguously differentiate structural and stratigraphic contacts in greenstone belts without fossils or rather fortuitous combinations of features will remain limited, the use of conventional criteria cannot be abandoned entirely. The possibility that thick, stratigraphically intact sequences are present in greenstone belts must remain as a working hypothesis until internal faults or folds can be identified based on clearly defined and well-understood criteria.

As noted above, it is our assessment that much of the controversy surrounding greenstone belt tectonics and evolution originates not from ambiguities in the genesis of rocks and structures in greenstone belts but from ambiguities in what those rocks and structures are and were. Future resolution of these controversies will rest more on careful, systematic studies of individual aspects of greenstone belts than on broad-brush syntheses or non-systematic collections of observations. A clear example of the success of the systematic approach is the role detailed geochronological studies have played in resolving the evolution of the late Archean Canadian belts. These studies (e.g. 5) have confirmed the existence of large-scale volcanic cycles within the Canadian greenstone belts and the existence of stratigraphic sections up to 10 km thick.

The results of any attempted overview of the similarities and differences among Archean greenstone belts depend significantly on how the term "greenstone belt" is defined. Presently used definitions (8) range from exceedingly broad (supracrustal successions in which mafic volcanic rocks are predominant) to relatively narrow (those requiring specific components, such as ultramafic or komatiitic lavas, and the increasingly common, largely implicit definition equating greenstone belts and ophiolites). Based on consideration of features common to most of the greenstone belts discussed in the present set of abstracts, we offer the following definition:

> Greenstone belt - an orogen made up largely of mafic to ultramafic volcanic rocks and their pyroclastic equivalents and epiclastic derivatives, showing intense macroscale deformation but regionally low grades of thermal alteration, and extensively intruded by penecontemporaneous or slightly younger granitoid plutons.

Virtually all terranes commonly considered as greenstone belts are encompassed by this definition, including many Phanerozoic examples. A critical aspect of this definition, and one that requires careful consideration, is that the terms "greenstone belt" and "ophiolite" are not synonymous. Rather, as in Phanerozoic orogens, ophiolites or ophiolite-like sequences may be components of greenstone belts.

Even with the restrictions imposed by this or most other definitions, greenstone belts constitute a highly diverse family of terranes. Some include an essentially continuous spectrum of komatiitic, tholeiitic, and calc-alkaline lavas, such as many belts in the Superior Province; others show a strongly bimodal volcanic suite (Barberton). Some are dominated by eruptive rocks (Superior Province, eastern Pilbara Block, and Barberton), others by sedimentary units (Slave Province and many Indian belts). The volcanic sequences in older greenstone belts (Barberton and eastern Pilbara) accumulated under shallow-water, anorogenic platform conditions; those in

most younger belts represent deep-water, tectonically active settings (7). Additional differences have been noted by other investigators (9, 10). These differences encompass nearly as much variability as represented by the spectrum of modern orogens. A possible implication of this diversity is that greenstone belts may represent tectonic settings as varied as those represented by modern orogenic belts.

The results of most modern studies of greenstone belts suggest that close scrutiny of individual belts usually allows identification of lithologically and structurally analogous modern terranes and, by inference, tectonic settings. There is an emerging consensus, for instance, that the petrologic, structural, and geochronological characteristics of large parts of the Superior Province indicate that it is an assembly of late Archean volcanic arcs formed along convergent plate boundaries that were basically similar to volcanic arcs and convergent boundaries today (Card, this volume). An important dissenting view, however, is expressed by David and others (this volume). Parts or all of the volcanic sequences of other Archean belts have been interpreted to represent oceanic or simatic crust formed at spreading centers.

Using a similar argument, the more-or-less regular vertical stratigraphic succession in greenstone belts, including lower volcanic and upper sedimentary stages, is grossly similar to the stratigraphic sequences in many modern orogens. If a genetic similarity is indicated, then it may be expected that individual greenstone belts include rocks formed in an evolutionary spectrum of tectonic settings. Perhaps, under ideal conditions of preservation, these may range from cratonic rift and/or ocean floor settings near the base to volcanic arc and, in some instances, cratonic or peri-cratonic settings at the top.

At the same time, if we look closely at individual greenstone belts, many features can be identified that are not present in their younger analogs. These include the common presence of extensive komatiitic lavas, banded iron formation, ocean-crust-like sequences (ophiolites) in excess of 10 km thick, and regionally extensive shallow-water sedimentary units deposited in anorogenic simatic settings. Some of these features, such as banded iron formation, reflect differences in modern and Archean systems that are probably unrelated to tectonics. Others, such as unusually thick ocean-crust sequences and widespread shallow water simatic platforms, may reflect important differences between Archean and Phanerozoic tectonic systems, if not in fundamental character then in local expression.

Future resolution of many of the outstanding controversies of greenstone belt evolution rests in detailed systematic studies of (i) individual properties of individual greenstone belts (structural style, alteration, sedimentology, petrology), (ii) differences among Archean greenstone belts, and (iii) similarities and differences between Archean belts and younger, apparently analogous terranes.

REFERENCES: (1) Barley, M. D. (1984) Geol. Dept. Univ. West. Aust. 9, 23-36; (2) DeWit, M. J., et al. (1982) Econ. Geol. 77, 1783-1802; (3) Lowe, D. R., et al. (1985) Precam. Res. 27, 165-186; (4) Byerly, G. R., et al. (1983) LPSC XIV, 84-85; (5) Krogh, T., et al. (1985) G.A.C. Spec. Paper 28, 358; (6) Fisher, G. L., and Lowe, D. R. (1985) G.S.A. Abs. 17, 583; (7) Lowe, D. R. (1982) Precam. Res. 17, 1-29; (8) Condie, K. C. (1981) Archean Greenstone Belts, Elsevier, Amsterdam; (9) Windley, B. F. (1977) The Evolving Continents, Wiley, Chichester; (10) Groves. D. I., and Batt, W. D. (1984) in Archean Geochemistry, Springer, New York, 73-98.

GREENSTONE BELTS: THEIR BOUNDARIES, SURROUNDING ROCK TERRAINS, AND INTERRELATIONSHIPS

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Introduction

Greenstone belts are an important part of the fragmented record of crustal evolution, representing samples of the magmatic activity that formed much of Earth's crust. Most belts developed rapidly, in less than 100 Ma, leaving large gaps in the geological record. Surrounding terrains provide information on the context of greenstone belts, in terms of their tectonic setting, structural geometry and evolution, associated plutonic activity, and sedimentation.

Tectonic Setting

Major controversy exists as to whether greenstone belts were deposited in oceanic, or marginal oceanic (1-3) or on rifted or thinned sialic crust (4-8). Archean volcanic sequences have much in common with Cenozoic volcanic arcs in terms of linear arrangements, rock types, and sequences, including calc-alkalic volcanic cones built on basal, subaqueous tholeiitic flows. Life spans are 5 to 20 Ma for individual volcanoes and 50 to 100 Ma for individual greenstone belts; some granite-greenstone terrains have several volcano-plutonic cycles differing in age by 200-300 Ma. Associated sediments consist of thin sequences of iron formation, chert, carbonate, and shale, and aprons of immature volcanogenic turbidites. Significant differences include the relative abundance of komatiites, the bimodal nature of some Archean sequences compared to the dominantly andesitic Cenozoic volcanoes, and the paucity of shelf sediments in Archean belts.

Direct evidence of oceanic settings for greenstone belts is rare. A well-preserved ophiolite sequence of Early Proterozoic age is reported from the Kainuu area of Finland (Kontinen, A., written communication, 1985) and a dismembered Archean ophiolite sequence has been interpreted in the southern Wind River Range (9). Neither is evidence for a dominantly continental setting compelling. Although sialic basement to the 2.7 Ga greenstone belts of the Slave and Superior Provinces of Canada has been recognized or inferred at several localities (4,10-13), most granitoid rocks are intrusive into, or in tectonic contact with, the volcanic rocks. Plutonic rocks, commonly with remnants of still-older supracrustal sequences, formed the basement to some volcanic piles, in a continental, micro-continental, or dissected arc setting.

A minor but significant component of Late Archean greenstone belts of the Superior Province is alkaline volcanic rocks, commonly associated with coarse alluvial-fluvial sediments, that unconformably overlie the major volcanic-plutonic successions, only a few Ma older (14-16). These sequences have many similarities to shoshonites formed in recently stabilized arcs (17).

Relationship of Greenstone Belts to Surrounding Terrains

In addition to rare unconformable relationships, fault, intrusive, and conformable depositional contacts characterize greenstone belt margins. Structure within greenstone belts is highly variable in both style and intensity of deformation. Common

features include sinuous, bifurcating folds, steep foliation and lineation and internal shear zones. Deformation may result from several causes, including: 1) tectonic emplacement of the belt (18-21); 2) diapiric rise of external and internal granitoid bodies (18,22-24); and 3) regional compression and/or transpression (25-27). In Slave Province stratigraphic onlap relationships between overlying greywacke-shale sequences and underlying volcanic rocks are common. This contrasts with the Superior Province, where belts of sedimentary rock, fault-bounded for the most part, alternate on a 50-150 km scale with major volcanic-plutonic belts.

As well as discrete fault contacts that form many belt boundaries, complex intercalation of volcanic and plutonic or sedimentary rocks by thrusting has been recognized in widespread locations (19,28-31). Thrusting at infrastructural levels may be an important process in high-grade gneissic terrains (32). Transcurrent displacements of at least several tens of kms have been estimated along some subprovince boundaries in the Superior Province (27,33,34), leading to the suggestion that greenstone and sedimentary subprovinces are accreted blocks. (27, 47, 59)

Plutonic Terrains. Plutonic rocks are particularly abundant in Archean volcano-plutonic terrains where they surround and intrude greenstone belts. Lithologically, these include variably xenolithic tonalite gneiss and more homogeneous bodies ranging from diorite to granite and syenite. Many syn-to post-kinematic plutons were emplaced during early magmatic and late diapiric stages spanning time intervals of ca 20 Ma (35). External plutons are generally similar in composition and age to plutons within belts. Although some plutonic rocks are older than and may represent basement to supracrustal sequences, contacts are generally intrusive or tectonic; precise zircon dating in Superior Province has demonstrated that many tonalite-diorite plutons are coeval with the volcanic hosts (13,36,37). Plutons of granodiorite-granite composition commonly post-date the youngest volcanic rocks and major tectonism by 5-25 Ma. Abbott and Hoffman (38) accounted for voluminous Archean tonalitic magmatism by tapping of low-temperature melts from large volumes of hydrous oceanic lithosphere consumed in shallow subduction zones. The equally voluminous granodiorite-granite magmatism may be the result of lower-crustal melting induced by thickening during collisional or accretionary events. (47).

Plutonic terrains east and west of the Kolar Schist belt have been interpreted as distinct continental fragments, sutured along the schist belt (39). Collisional processes between Precambrian blocks have not been substantiated paleomagnetically (40).

Metasedimentary Belts. Large tracts of metasedimentary rock, predominantly greywacke and shale deposited in turbidite sequences, are distinguished from the iron formation-chert-carbonate-shale successions commonly associated with greenstone belts. Metasedimentary belts, commonly metamorphosed to amphibolite facies gneiss and migmatite, constitute a significant supracrustal component of many Archean terrains, most notably the Slave and Superior Provinces of Canada.

Turbidites make up some 80% of the supracrustal sequences within the Slave Province (70). Deposition of sediments of felsic volcanic and plutonic derivation (41), is thought to be broadly coeval with eruption of marginal volcanic sequences of about 2670 Ma age (10), possibly in response to regional extension (42). The turbidites have alternatively been interpreted (20) as trench-fill deposits in a prograding accretionary complex. Sialic basement of 3 Ga age (43,44), recognized at several locations, has

been variably interpreted as continuous pre-greenstone sialic crust or as microcontinental fragments. Low-pressure regional metamorphism results from the rise of thermal domes (45), possibly associated with the intrusion of plutons.

Three major linear metasedimentary belts separate granite-greenstone terrains of the Superior Province (46,47): the English River, Quetico and Pontiac belts. Although volcanic rocks are rare or absent from the turbiditic sequences, a felsic volcanic (48) or mixed volcanic and plutonic provenance (49) is inferred. Sedimentary sequences are generally in fault contact with adjacent terrains and increase in metamorphic grade from low at the margins to high (migmatite to low-P granulite) in axial regions, where plutons, particularly peraluminous monzogranites, are abundant. It is apparent that these belts developed as elongate sedimentary basins collecting detritus from adjacent volcanic-plutonic highlands and were later subjected to deformation, axial plutonism and high-level metamorphism.

The oldest detrital zircons in metasedimentary belts are commonly derived from ancient terrains either not yet recognized, at great distance from sediment deposition, or destroyed, buried or allochthonous subsequent to the erosional event. Examples include 4.2 Ga zircons in the 3.5 Ga Mt. Narryer quartzite (50), 3.1 Ga zircons in the 2.7 Ga Pontiac belt (51), and 3.8 Ga zircons in the 3.7 Ga Nulliak quartzite (52).

Relationship Between Low and High-Grade Terrains. High-grade terrains form large parts of some Archean cratons and have variable relationships to adjacent greenstone belts. Characterized by upper-amphibolite to granulite-facies metamorphic grade in mainly intrusive rock types, high-grade terrains have been interpreted as either lateral equivalents of greenstone belts, in a different tectonic environment (53,2), or as the deeply-eroded roots to greenstone belts (54). Geobarometry is a useful tool in distinguishing between alternative interpretations in specific areas. Recognition of geological and geophysical criteria of crustal cross-sections (55) may also guide interpretation.

Examples of both lateral and vertical transitions from low to high-grade terrains are documented in the Superior Province. A lateral relationship has been inferred for the high-grade Quetico metasedimentary belt and adjacent low-grade Wabigoon and Wawa metavolcanic-plutonic belts. Volcanic rocks were deposited 2750-2695 Ma ago (13,26). Coeval turbiditic metagreywackes of the Quetico belts, about 2744 Ma old (56) have an axial high-temperature, low pressure zone of schist, migmatite, S-type granites and local granulite (58-60), suggesting a major thermal anomaly at high structural levels. Different tectonic settings and evolution are proposed for the low- grade volcanic (arc) and high-grade metasedimentary (marginal basin) terrains. Differences in structural style between belts can be attributed to variable levels of exposure (60) or mechanical character.

Evidence of dextral transpressional deformation characterizes the Wawa-Quetico-Wabigoon boundary region. This includes: 1) assymetric folds and other kinematic indicators in the northern Wawa (26), Quetico (60) and southern Wabigoon (27) belts, and 2) conglomerate and alkaline volcanic deposits associated with strike-slip faults (27,26). The event is bracketed between 2695 and 2685 Ma by zircon dates (13).

Adjacent high and low-grade Archean terrains have been interpreted, by analogy with the Cenozoic Rochas Verdes complex (2), as deeply-eroded arcs and adjacent back-arc basins respectively.

Vertical relationships between low and high-grade regions have been interpreted in the intracratonic Kapuskasing uplift (61,62) and marginal Pikwitonei region (63) of the Superior Province, as well as in the Kaapvaal Craton (64). An uninterrupted oblique cross-section through the Michipicoten greenstone belt to lower crustal granulites is exposed across a 120-km-wide transition in the southern Kapuskasing uplift. Well-preserved metavolcanic and metasedimentary rocks of the greenstone belt, metamorphosed to greenschist facies at 2-3 kbar, are intruded and underlain by some 10-15 km of tonalitic rocks which increase in structural complexity from homogeneous plutons to contorted gneisses with increasing depth. Lowermost in the section is a heterogeneous granulite complex, at least 10 km thick, of interlayered supracrustal (15%) and intrusive (85%) rocks recording metamorphic conditions of 700-800°C, 7-8 kbar (66). The crustal slab was emplaced onto low-grade rocks of the Abitibi belt on the Ivanhoe Lake thrust (66) some 2 Ga ago.

In the Pikwitonei region, distinctive rock types including iron formation, pillow basalt, calc-silicates and anorthosite can be traced along strike from the low-grade Sachigo Subprovince into Pikwitonei granulites (63). Supracrustal rocks step up in metamorphic grade across faults (67) as intrusive rocks become more abundant. Metamorphic pressure increases within the granulites from 7 to 12 kbar (68) toward the western boundary, the Nelson Front. Both the Kapuskasing and Pikwitonei structures have diagnostic features of crustal cross-sections including gradients of metamorphic grade and pressure, high proportions of intrusive rock types and paired gravity anomalies.

REFERENCES

1) Glikson, A.Y., 1979, Earth Sci. Rev. 15, 1-73. 2) Tarney, J., Dalziel, I.W.D. and DeWit, M.J., 1976, in The Early History of the Earth, 131-146. 3) Windley, B.F., 1981, The Evolving Continents. 4) Baragar, W.R.A. and McGlynn, J.C., 1976, Geol. Surv. Can. Paper 76-14. 5) Ayres, L.D. and Thurston, P.C., 1985, Geol. Assoc. Can. Sp. Pap. 28, . 6) Hunter, D.R. et al., 1986, this volume. 7) Wilks, M.E. and Nisbet, E.G., 1986, this volume. 8) Fripp, R.E.P., 1986, this volume. 9) Harper, G.D., 1986, this volume. 10) Henderson, J.B., 1981, in Precambrian Plate Tectonics Elsevier, 213-235. 11) Joliffe, A.W., 1955, Econ. Geol., 50, 373-398. 12) Clark, G.S., Bald, R., Ayres, L.D., 1981, Can. J. Earth Sci., 18, 94-102. 13) Davis, D.W., Corfu, F. and Krogh, T.E., 1986, this volume. 14) Cook, D.L. and Moorehouse, W.W., 1969, Can. J. Earth Sci., 6, 117-132. 15) Shegelski, R.J., 1980, Precambrian Res., 12, 331-347. 16) Brooks, C. et al., 1982, Can. J. Earth Sci., 19, 55-67. 17) Joplin, G.A., 1968, Geol. Soc. Austr. J., 15, 275-294. 18) St-Onge, M.R. et al., 1986, this volume. 19) deWit, M.J., 1986, this volume. 20) Hoffman, P.F., 1986, this volume. 21) Kusky, T.M., 1986, this volume. 22) Schwerdtner, W.M., Morgan, J. and Stott, G.M., 1985, in Geol. Assoc. Can. Sp. Pap. 28, 117-123. 23) Schwedtner, W.M., 1986, this volume. 24) Dixon, J.M. and Summers, J.M., 1985, Can. J. Earth Sci., 20, 1843-1861. 25) Park, R.G., 1981, Precambrian Res., 14, 31-47. 26) Corfu, F. and Stott, G.M., 1985 (abstr.), Inst. Lake Superior Geology, 31, 15-16. 27) Poulsen, H.,

GREENSTONE BELTS: THEIR COMPONENTS AND STRUCTURE.

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Although in common geological usage there is considerable ambiguity over the definition of greenstone belts which are historically regarded as long and narrow in shape, Archaean in age and composed of volcanic and sedimentary sequences at greenschist facies. This definition remains true for many of what are commonly regarded as greenstone belts but others differ significantly, particularly in shape and metamorphic facies. For this reason the term 'succession' is prefered for greenstones which are not particularly linear. In the following discussion it is our intention to maintain 'greenstone' as a useful term and for that reason we specifically aim to exclude high-grade supracrustal gneiss terrains such as those of the central zone of the Limpopo belt and early Precambrian supracrustal sequences such as the 3Ga Pongola, the 2.7Ga Witwatersrand and the 2.4Ga Ventersdorp from any definition of greenstone successions. We also aim to include all commonly accepted greenstone successions. The following points are of relevance to the definition of greenstone belts:

1. Most commonly accepted greenstone successions are of Archaean age but a few younger belts have been reported from Wisconsin, USA (1) and northern Quebec, Canada (2).
2. Although many greenstone successions are long, linear and narrow (e.g. Pietersburg and Murchison, Kaapvaal craton) many others have more irregular shapes (eg. Bulawayan, Zimbabwean craton and Pilbara, Western Australia). The word 'belt' therefore is inappropriate for some greenstone successions.
3. Volcanic rocks are ubiquitous components whereas sediments may be of secondary importance. The volcanics frequently include komatiitic rocks. Intrusive igneous rock units such as layered complexes, dykes and sills may be present.
4. Greenstone successions occur at metamorphic conditions from sub-greenschist to granulite facies and the colour prefix, refering to the greenschist facies, is unfortunate.
5. Deformation intensity within the greenstone successions is variable.
6. Greenstone successions are always intimately associated with and surrounded by trondhjemite-tonalite-granodiorite-granite granitoids.

We tentatively suggest the following definition:
Greenstone successions are the non-granitoid component of granitoid-greenstone terrains. Volcanic rocks are an essential component, some of which are usually komatiitic. Sedimentary rocks are commonly present and igneous intrusive units may exist. The greenstone successions are linear to irregular in shape and where linear they are termed belts. The greenstone successions may occur at all metamorphic facies and are heterogenously

deformed. Most greenstone successions are Archaean in age.

Greenstone successions comprise a wide variety of rocks, dominated by volcanics, which are usually altered and deformed. Alteration of volcanic and other rock types is manifested by hydration with variable silicification (3), carbonate-isation (4) or silica loss (5) as well as isochemical metamorphism. Alteration itself is temporally and spatially variable, Smith and Erlank (6) have described possible early sea floor alteration of komatiitic rocks from Barberton and carbonate-isation in Murchison is patchy and syn- to post-tectonic. This alteration constrains identification of original rock-types and the use of whole rock chemistry. This restriction added to the problems of equating area of surface outcrop with rock volume means that estimates of greenstone lithological proportions must be treated circumspectly. However, greenstone successions commonly comprise the following primary lithologies: komatiitic, mafic and felsic volcanics, cherts, banded iron formations, shales, graywackes and quartz arenites. Less commonly limestones (including stromatolites), arkose, ultramafic and mafic layered complexes, quartz-feldspar porphyries and quartz tholeiite dykes are present.

The identification of the environment of emplacement of greenstone igneous rocks is highly problematic. Subvolcanic intrusions exhibit many features almost indistinguishable from true lavas. Skeletal crystal growths, commonly grouped under the all-embracing term of 'spinifex', are an important textural form in these rocks and these textures, in abundance, are restricted to Archaean greenstone successions. These textures are indicative of rapid crystal growth under supersaturated conditions (7) and need not be restricted to lava flows. In fact, the inordinatly thick cumulate zones associated with some spinifex-bearing rock-types preclude these being lava flows in the currently accepted sense and the non-genetic term 'cooling unit' has been used to describe these layered rocks which may represent lava flows or subvolcanic intrusions. The recognition of crescumulate type crystal growth and rhythmically developed spinifex units indicate a variety and complexity of mechanisms which have given rise to these textures and criteria should be established to permit the environment of emplacement to be determined more precisely. Symmetry of structures and spinifex textures encountered in some units may be indicative of dyke emplacement.

Until recently, greenstone research was largely oriented towards deducing a unifying model, subsequently heterogeneity has become the key-word. In essence, greenstone belts are of different ages and formed in different tectonic situations. Groves and Batt (8) recognise both younger and older greenstone successions in Western Australia in two distinct environments, determined on the basis of volcanic constituents, sedimentary facies, mineral deposits and tectonic style, to which they gave a

genetic interpretation as rift-phase or platform-phase greenstones. Whereas this is a major development in understanding Australian greenstones the division of other greenstone successions into rift- and platform-phase is tenuous, particularly for those of the Kaapvaal craton. The Murchison greenstone belt, for instance, has characteristics of both rift- and platform-phase greenstones.

The Barberton greenstone belt, comprising the lower komatiitic to felsic units of the Onverwacht Group and overlying deep water sediments of the Fig Tree Group, probably represents a rift-phase (8) and the overlying Moodies Group with shallow water quartzites and banded iron formation is typical of a platform-phase greenstone belt. However, herewithin lies an important observation on greenstone successions: the environment of formation can vary within a greenstone. This variation may be due to either:

1. A progressive evolution in environment. (Eriksson (9) has described the Fig Tree to Moodies group evolution of the Barberton greenstone belt in terms of an evolving back-arc, or passive continental margin.)
2. The superposition of different environments which are temporally separate and manifested in the field by an unconformity.

or 3. Some or all of the units are allochthonous and represent spatially and/or temporally diverse environments now tectonically juxtaposed.

Another aspect of the heterogeneity is the recognition of both continental and oceanic environments. The Mberengwa greenstone belt of Zimbabwe rests unconformably on granitic rocks (10, 11, 12). Basement has also been inferred to exist beneath other greenstone belts in Australia, Canada and India (13, 14, 15). Major layered igneous complexes such as Dore Lake (16) and the Rooiwater, Murchison greenstone belt (17), are a significant component of some greenstone belts. These complexes have minor ultramafic components, anorthosite-gabbro layers, magnetitite layers and a highly differentiated and sodic granite. These complexes are analogous to bodies such as the Bushveld and are intrusions in a continental environment.

In contrast to the continental environment of some greenstone successions no proven continental basement exists at the base of the Barberton greenstone belt and the Onverwacht Group may be partially of oceanic origin (18). In addition, some ultramafic complexes may also be ophiolitic (19). De Wit and Stern (20) have recognised a possible sheeted-dyke complex in the Onverwacht group. Support for the obducted oceanic origin for some greenstone rocks comes from the recognition of podiform alpine-type chromites at Shurugwi (Zimbabwe) (21, 22) and at Lemoenfontein (Kaapvaal craton) (23). These have textural and chemical characteristics similar to those recognised in ophiolitic complexes of Phanerozoic age.

GREENSTONE BELTS: THEIR COMPONENTS AND STRUCTURE
Vearncombe, J.R. et al.

Historically greenstone structures were regarded as simple synformal belts between sub-circular rimming granitoid domes. This relationship has given rise to genetic interpretations that greenstone belts are pinched-in synformal keels between domal or diapiric granitoids or between granitoid domes which are the result of interference folding (24). Unfortunately the paucity of detailed structural observations and accurately determined stratigraphic successions mean that few of the assumed synforms are proven.

In the Kaapvaal craton the Murchison, Pietersburg, Sutherland, Rhenosterkoppies, Amalia and Muldersdrift belts lack a gross synformal structure. At Barberton the greenstone succession comprises several synformal structures separated by steep reverse faults (25). De Wit (26) and Lamb (27) have recently described thrusts, some of which emplace Onverwacht volcanics over Moodies sediments. The suggestion of Anhaeusser (28) that deformation structures within the Barberton greenstone belt can mostly be related to granitic diapirism is at variance with the observed thrust structures and evidence presented by Ramsay (25), Roering (29) and Burke et al. (30) who note deformation structures prior to granite intrusion, intrusive granite contacts oblique to deformation structures and an absence of deformation structures within the greenstone directly related to those in the surrounding granitoids.

We suggest that whereas broadly synformal belts may exist this is not a characteristic of greenstone belts. Many of the intrusive granitoids are undoubtedly domal but intervening greenstone belts are not necessarily synformal and the role of diapirism in controlling the structure of greenstone successions may be over-emphasised.

In deducing the overall large-scale structural characteristics of greenstone successions the following general observations may be relevant:

1. Contacts with the surrounding granitoids can be either tectonic (31) or intrusive with dykes and veins of granitic rock in the greenstone belts and a static high T/low P metamorphism near the greenstone contact with the granitoids suggesting contact metamorphism by igneous intrusion.
2. Geophysical evidence from a number of belts suggests they are shallow with vertical depth extents rarely more than 10km and usually less than 5km (32, 33), figures considerably less than the proposed stratigraphic thicknesses of these belts. This shallow depth extent suggests no simple rotation of the usually upright greenstone belt but instead a truncation which may be a major decollement zone, recumbent syntectonic granite or a late intrusive contact.
3. Recumbent fold structures and possible thrusts are relatively common and have been described from greenstone successions of the Zimbabwean craton (34, 35), of the Kaapvaal craton (25, 26, 27),

of the Western Australia shield (36, 37) and the north American shield (38).
4. Greenstone successions occur as either linear belts or as irregular shaped units comprising arcuate arms.
5. Late-deformation structures and the present disposition of primary layering structures in the greenstone successions are usually upright.

Greenstone successions are composed of deformed and metamorphosed (including metasomatised) rocks. However despite the obvious difficulties, many authors have proposed stratigraphies for greenstone belts, but some have deduced total stratigraphic thicknesses dramatically in excess of those predicted by currently accepted models for basin formation (39, 40). Greenstone successions such as Barberton with 17 to 23km (41), Pietersburg with 21.4km (41) and Abitibi with over 30km (42) or up to 45km (43) total stratigraphic thickness contrast with both thinner sequences from other greenstone and non-greenstone early Precambrian supracrustal sequences such as the Witwatersrand. It is the greenstone successions with large stratigraphic thicknesses which are invariably at sub-greenschist or greenschist facies and without the high grades of metamorphism that would be expected at the base of these sequences. These thicknesses represent one of the challenging problems in greenstone geology.

Possible explanations for the large stratigraphic thicknesses are as follows:
1. They are an artifact of combining separate sections into a composite section or are oblique sections.
2. That incorporated within the greenstone belt and incorrectly interpreted as part of the stratigraphy are layered igneous complexes, sills and tectonically rotated dykes.
3. The stratigraphic sequences are in fact related to two or more spatially superimposed but temporally separate and essentially unrelated events. In the Barberton greenstone belt granite cobbles in a Moodies Group conglomerate have yielded zircons giving ages of 3.15Ga (44) contrasting with ages of 3.54Ga (45) for the stratigraphically lower Onverwacht volcanic rocks. A major phase of granite emplacement separates these two dates and a major unconformity may exist at the base of the Moodies Group.
4. They are not true stratigraphic sections but are structurally repeated by imbricate thrusting and/or folding. To achieve significant structural repetition by thrusting, folding or both requires major recumbent tectonics on or above a decollement plane.

Whilst explaining large stratigraphic repetition the recumbent thrust-fold model also predicts metamorphic conditions at the base of the pile initially at high P/low T and with thermal relaxation to medium pressure facies. Bickle et al. (46) have reported such rocks from the Yilgarn and similar staurolite-

kyanite-bearing rocks occur in the Murchison greenstone belt. However the very large apparent stratigraphic thicknesses with associated sub-greenschist or greenschist metamorphism remain unexplained by horizontal thrust-nappe tectonics. These may however be explained by repetition above a flat decollement in an imbricate stack with associated folding. In this situation the stratigraphy is turned on end and multiply repeated but the structure remains shallow. Zones of cyclic repetition should be investigated to determine if the cyclicity is real or the result of imbricate stacking. Examples of this type of structural stacking resulting in repetition are provided by Coward et al. (35) from Matsitama, Zimbabwean craton, Botswana and Martyn (37) from the Kalgoorlie area in the Norseman-Wiluna greenstone belt (Western Australia).

References

(1) Schultz, K.J. & LaBerge, G.L. this volume. (2) St-Onge, M.R., Hoffman, P., Lucas, S.B., Scott, D.J. & Begin, H.J. this volume. (3) Paris, I., Stanistreet, I.G. & Hughes, M.J. 1985, J. Geol. 93, 111-130. (4) Pearton, T.N. 1980, The geochemistry of carbonate & related rocks of the antimony line, Murchison greenstone belt, with particular reference to their genesis and to the origin of stibnite mineralization. Ph.D. thesis, Univ. Witwatersrand. (5) de Wit, M.J. this volume. (6) Smith, H.S. & Erlank, A.J. 1982, in Arndt, N.T. & Nisbet, E.G. Komatiites, 347-397. (7) Donaldson, C.H. 1982, in Arndt, N.T. & Nisbet, E.G. Komatiites, 211-244. (8) Groves, D.I. & Batt, W.D. 1984, in Kroner, A. Archaean geochemistry, 73-98. (9) Eriksson, K.A. 1980, Precambrian Res. 12, (10) MacGregor, A.N. 1951, Trans. geol. Soc. S. Afr. 54, xxvii-lxxi. (11) Wilson, J.F. 1973, Phil. Trans. Roy. Soc. Lond. A273, 389-411. (12) Bickle, M.J., Martin, A. & Nisbet, E.G. 1975, Earth Planet. Sci. Lett. 27, 155-162. (13) Archibald, N.J., Bettenay, L.F., Binns, R.A., Groves, D.I. & Gunthorpe, R.J. 1978, Precambrian Res. 6, 103-131. (14) Baragar, W.R.A. & McGlynn, J.C. 1976, Geol. Surv. Canada Pap. 76-14, 20pp. (15) Chadwick, B., Ramakrishnan, M. & Viswanatha, M.N. 1981, J. geol. Soc. Ind. 22, 557-569. (16) Allard, G.O. 1970, Geol. Soc. S. Afr. Sp. Publ. 1, 477-491. (17) Vearncombe, J.R., Walsh, K.L. & Barton, J.M. Jr. in prep. (18) Anhaeusser, C.R. 1973, Phil. Trans. Roy. Soc. Lond. A273, 359-388. (19) Wuth, M. 1980, The geology and mineral potential of the Oorschot-Weltevreden schist belt south-west of Barberton - eastern Transvaal, M.Sc. Thesis, Univ. Witwatersrand. (20) De Wit, M.J. & Stern, C.R. 1980, Extended abtracts 2nd. Int. Archaean Symp. 85-86. (21) Cotterill, P. 1969, The chromite deposits of Selukwe, Rhodesia. Econ. geol. Monogr. 4, 154-186. (22) Stowe, C.W. 1984, in Kroner, A. & Greiling, R. Precambrian Tectonics Illustrated, 41-56. (23) Smit, A.C. 1984, Trans. geol. Soc. S. Afr. 87, 303-314. (24) Snowden, P.A. 1984, in Kroner, A. & Greiling, R. Precambrian Tectonics Illustrated, 135-145. (25) Ramsay, J.G. 1963, Trans. geol. Soc. S. Afr. 66, 353-398. (26) De Wit, M.J. 1982, J. structural geol. 4, 117-136. (27) Lamb, S.H. 1984, in

Kroner, A. & Greiling, R. Precambrian Tectonics Illustrated, 19-39. (28) Anhaeusser, C.R. 1984, in Kroner, A. & Greiling, R. Precambrian Tectonics Illustrated, 57-78. (29) Roering, C.R. 1967, Economic geol. Res. Unit. Rpt. 35, Univ. Witwatersrand. (30) Burke, K., Dewey, J.F. & Kidd, W.S.F. 1976, in Windley, B.F. The Early History of the Earth, 113-130. (31) Spray, J.G. 1985, J. structural geol. 7, 187-203. (32) de Beer, J.H., Stettler, E.H., Barton, J.M. Jr., van Reenen, D.D. & Vearncombe, J.R. this volume. (33) Thomas, M.D., Losier, L., Thurston, P.C., Gupta, V.K., Gibb, R.A. & Grieve, R.A.F. this volume. (34) Stowe, C.W. 1974, J. geol. Soc. Lond. 130, 411-425. (35) Coward, M.P., Lintern, B.C. & Wright, L.I. 1976, in Windley, B.F. The early History of the Earth, 323-330. (36) Platt, J.P. 1980, Tectonophysics 65, 127-150. (37) Martyn, J.E. this volume. (38) Poulsen, K.H., Borradaile, G.J. & Kehlenbeck, M.M. 1980, Can. J. Earth Sci. 17, 1358-1369. (39) McKenzie, D. 1978, Earth Planet. Sci. Lett. 40, 25-32. (40) McKenzie, D., Nisbet, E. & Sclater, J.G. 1980, Earth Planet. Sci. Lett. 48, 35-41. (41) SACS, 1980, Stratigraphy of South Africa. Handbk. 8., Geol. Surv. S.A. 690pp. (42) Jensen, L.S. 1985, Geol. Assoc. Canada Sp. Pap. 28, 65-87. (43) Ayers, L.D. & Thurston, P.C. 1985, Geol. Assoc. Canada Sp. Pap. 28, 343-380. (44) van Niekerk, C.B. & Burger, A.J. 1978, Spec. Publ. geol. Soc. S. Afr. 4, 99-106. (45) Hamilton, P.J., Evensen, N.M., O'Nions, R.K., Smith, H.S. & Erlank, A.J. 1979, Nature, 179, 298-300. (46) Bickle, M.J., Morant, P., Bettenay, L.F., Boulter, C.A. Blake, T.S. & Groves, D.I. 1985, Geol. Assoc. Canada Sp. Publ. 28, 325-341.

GREENSTONE BELTS ARE NOT INTRACONTINENTAL RIFTS. WHAT THEN ARE THEY?

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Hundreds of intracontinental rifts ("elongate depressions [within continents] overlying places where the lithosphere has ruptured in extension" ref. 1) with ages between 3.0 and 0 Ga have been recognized on earth (2,3,4). Compressional features are either absent or insignificant in the vast majority of these rifts. Prominent compressional features are reported from only a very few rifts. (Notably: the Benue trough (5) the Dneipr-Donetz rift (Fig. 1) (6) the Southern Oklahoma rift (7) and the rift occupying East Arm of Great Slave Lake (8)).

Intense compression is the rule in greenstone belts and preservation of regional extensional structures is rare. (Abstracts at this meeting). Whatever greenstone belts are they do not satisfy the definition of intracontinental rifts.

Wilson (9) showed that a common fate of intracontinental rifts is to develop into oceans and that oceans are likely to close. Mountain belts mark places where oceans have closed. In contrast to intracontinental rifts both mountain belts and greenstone belts are dominated by compressional structures. Pursuing Wilson's idea I therefore suggest that it might be useful for students of greenstone belts to test the hypothesis that: "Greenstone belts are mountain-belts marking places where OCEANS have closed". Ocean closing is a complicated process (ref. 1) and some of the regional complexities that may be recorded in greenstone belts are indicated in Fig. 2.

There is a possibility that students of greenstone belts are confusing each other because some who describe greenstone belts as intracontinental rifts may be consciously concentrating on an early episode in greenstone belt evolution and recognize that the belts have a later oceanic and collisional history. I suggest that this practise is confusing and is rather like describing Ronald Reagan as a movie actor and ignoring more significant later episodes in his career.

References

1. Basaltic Volcanism Study Project (1981). Basaltic Volcanism on the Terrestrial Planets. Pergamon Press, Inc., New York, 1286 pp.

2. Morgan, P. and Baker, B.M. (1983). Processes of continental rifting. Tectonophysics, 94, 680 pp.

3. Burke, K., Kidd, W.S.F. and Kusky, T.M. (1985). The Pongola structure of south-eastern Africa: The World's oldest preserved rift? Journ. of Geodynamics, 2, 35-49.

4. Burke, K., Delano, L., Dewey, J.F., Edelstein, A., Kidd, W.S.F., Nelson, K.D., Sengor, A.M.C. and Stroup, J. (1978). Rifts and sutures of the World. Contract Report NAS5-24094. Geophysics Branch, ESA Division, Goddard Space Flight Center, Greenbelt, Maryland. 238 pp.

5. Benkhelil, J., Dainelli, P., Ponsard, J.F., Popoff, M. and Saugy, L. (1986) The Benue trough in press AAPG special publication Sediments in Rifts ed. Warren Manspeizer.

6. Sengor, A.M.C. (1985). Die Alpiden und die Kimmeriden: Die verdoppelte Geschichte der Tethys. Geol. Rundsh., 74, 181-213 pp.

7. Ham, W.E., Denison, R.E. and Merritt, C.A. (1964). Basement rocks and structural evolution of Southern Oklahoma. Oklahoma Geol. Surv. Bull., 95, 302pp.

8. Hoffman, P.E. (1980). Wopmay Orogen: a Wilson cycle of early Proterozoic age in the northwest of the Canadian Shield. In The Continental Crust of the Earth and its Mineral Deposits (D.W. Strangway, ed.). p.523-549. Geol. Assoc. Canada Spec. Paper 20, Toronto, Ontario.

9. Wilson, J.T. (1968). Static or Mobile Earth: the current scientific revolution. In Gondwanaland Revisited, p.309-320. Amer. Philos. Soc. Proc. vol. 112.

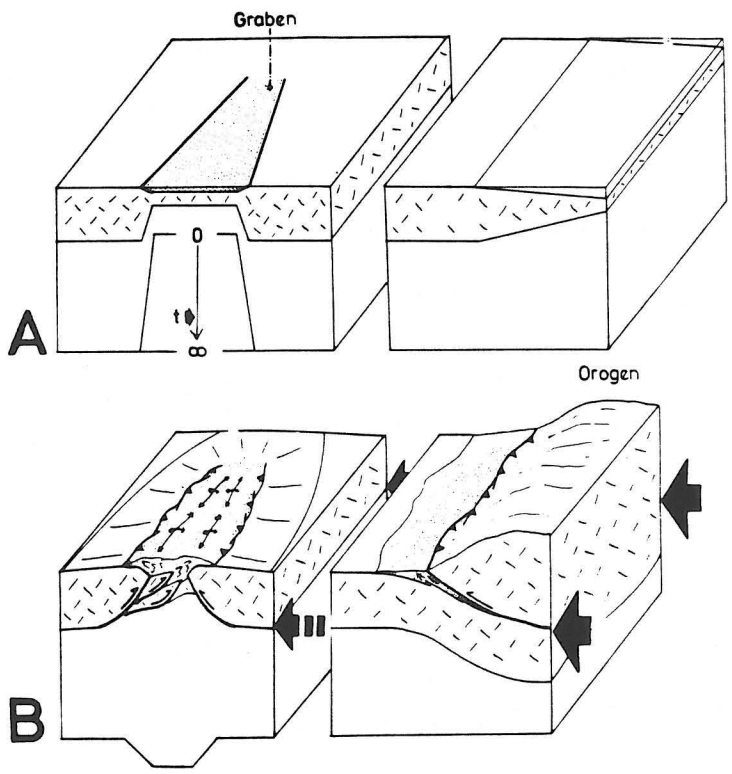

Figure 1. (from ref. 6) Illustration of how rifts within a continent (such as the Dneipr-Donetz rift) have been affected by neighboring continental collisions (as the Dneipr-Donetz structure responded to collisions in North Dobrudja in the Early Jurassic). Observation has shown that folding and thrusting developed in this environment is much less intense than that with which we are becoming familiar in greenstone belts.

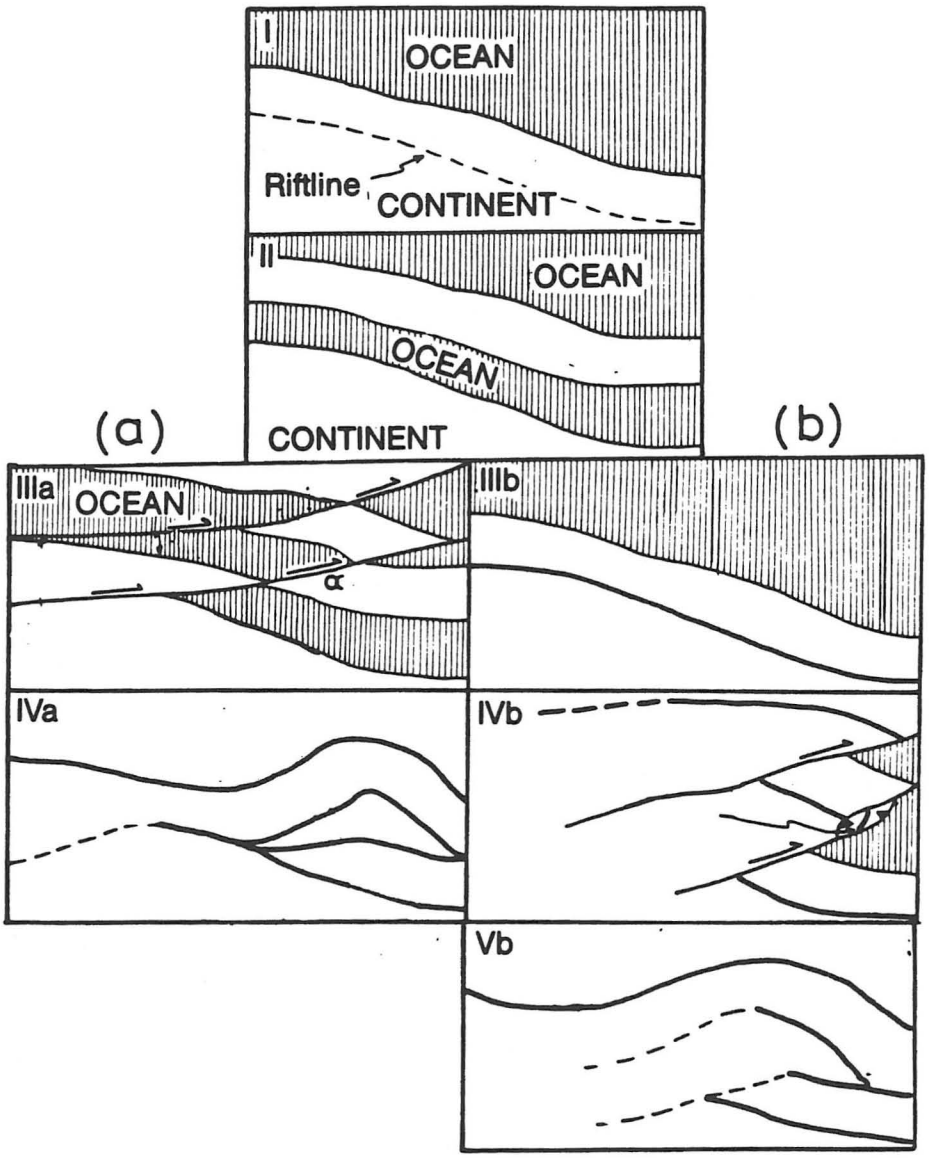

Figure 2. (from ref. 6) A possible origin for some greenstone belts. Rifting (I) takes a continental fragment out into an ocean (II). Major strike-slip motion (IIIa) is depicted as preceding collision between slivers of the continental fragment and the main continent (IVa). As an alternative suturing may take place (IIIb) before major strike-slip motion (IVb). In either case the preserved suture zones may end abruptly at strike-slip faults and late rotation may preserve puzzling polarities (Vb).

THE WESTERN WABIGOON SUBPROVINCE, SUPERIOR PROVINCE, CANADA: LATE ARCHEAN GREENSTONE SUCCESSION IN RIFTED BASEMENT COMPLEX.

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The Wabigoon Subprovince, interposed between the predominantly metasedimentary-plutonic and gneissic English River and Quetico Subprovinces to the north and south respectively, exposes Archean greenstone and granitoid rocks for a strike length of greater than 700 km. Based on predominating rock types, the western part of the subprovince is divided into two terranes: the northwestern Wabigoon volcano-sedimentary and plutonic terrane (NWW) and the Wabigoon Diapiric Axis terrane (WDA)(1).

NWW in Ontario extends southwesterly from Savant Lake to Lake of the Woods. Organized searches for older and younger age limits for the evolution of this terrane, yield reliable zircon U-Pb ages for supracrustal strata that span from 2755 Ma to 2711 Ma, although most ages are between 2720 Ma and 2734 Ma (2,3,4,5,6,7). The lowermost volcanic sequence in the western part of NWW is bimodal Mg-rich tholeiitic basalt and rhyodacite at Thundercloud Lake (2755 Ma); later, at 2734 to 2718 Ma, bimodal Fe-rich tholeiitic basalt and rhyodacite (Dash Lake) is attended by bimodal basalt and tonalite plutonism. This stage overlaps with intermediate to felsic calc-alkaline volcanism (Kakagi Lake). The latest volcanism in the sequence at 2711 Ma is dacite at Stephen Lake (3,7) which is conformable with the subjacent Kakagi Lake strata and as such gives an upper limit for the age of major tectonism affecting the supracrustal rocks.

WDA is a 400 km long by 75 km wide domal structure which consists of 1) gneissic tonalitic to granodioritic rocks forming domes and lesser massive segregations, 2) crescentic dioritic to granitic plutons occurring at or near the contact between the gneiss domes and the Wabigoon supracrustal rocks, and 3) later plutons of diorite to granite (1,8,9). U-Pb geochronology indicates that at least some of the eastern part of the terrane, which extends from Steep Rock Lake in the south to Caribou Lake in the north, has some old (approx. 3.0 Ga) gneissic and supracrustal rocks (10). The western part of WDA, so far has not yielded old ages; gneissic to massive tonalitic rocks have intrusive ages of 2720-2725 Ma (3,7,9). At least some of the gneissic tonalite forming the domes in the western part of WDA have ages similar to, and in the field are gradational with, tonalite plutons intruding NWW. A sphene U-Pb age of 2674 Ma for gneissic tonalite with a zircon U-Pb age of 2723 Ma suggests that the gneissification was a late event involving the resetting of the sphene age but that the age of intrusion was retained by the zircon. The crescentic and later plutons dated so far have ages near 2700 Ma (3,7,9) and do not have regional foliation thus providing an approximate lower limit for the age of major tectonism in the terrane.

NWW is interpreted to have formed during rifting of a basement complex that underlies the adjacent English River

Subprovince (11) and the western part of NWW and WDA. The complex is approximately 3.0 Ga old and perhaps older. The rifting started with mafic magmatism which evolved to be bimodal basalt-rhyodacite. Tonalite intrusions accompanying the bimodal volcanism caused little or no deformation of the adjacent supracrustal rocks (12). Much of the contemporaneous calc-alkaline sequence may be from mixing of basalt and tonalitic magmas. The age of major deformation in the supracrustal rocks may be bracketed by the age of the uppermost (and conformable) Stephen Lake dacite at 2711 Ma and the age of the posttectonic plutons at approximately 2700 Ma. Heating of the lower crust by ponding of mafic magma caused most of the deformation of both the younger Wabigoon 'rift' sequence and the basement complex; WDA is the scar of maximum crustal diapirism transecting the new and old crust.

REFERENCES:

1) Edwards GR and Sutcliffe RH 1980 Geol Ass'n Can Abst 5 :50
2) Davis DW et al 1982 Can Jour Earth Sci 19:254
3) Davis DW and Edwards GR 1982 Can Jour Earth Sci 19:1235
4) Davis DW and Trowell NF 1982 Can Jour Earth Sci 19:868
5) Edwards GR and Davis DW 1984 Ont Geol Surv Misc Pap 121:222
6) Davis DW and Edwards GR 1985 Ont Geol Surv Misc Pap in press
7) Davis DW and Edwards GR 1986 Can Jour Earth Sci in press
8) Schwerdtner WM et al 1979 Can Jour Earth Sci 16:1965
9) Sutcliffe RH 1980 MSc Thesis U of Toronto Unpubl
10) Davis DW and Jackson 1985 Ont Geol Surv Summary Field Work in press
11) Krogh TE et al 1976 Can Jour Earth Sci 13:1212
12) Edwards GR 1985 PhD Thesis U of Western Ontario Unpubl

A SIMPLE TECTONIC MODEL FOR CRUSTAL ACCRETION IN THE SLAVE PROVINCE: A 2.7-2.5 Ga "GRANITE-GREENSTONE" TERRANE, NW CANADA

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A prograding (direction unspecified) trench-arc system is favored as a simple yet comprehensive model for crustal generation in a 250,000 km^2 "granite-greenstone" terrane (1). The "greenstone" belts are seen as synformal remnants of a formerly continuous complex of tectonically accreted seamounts, remnant arcs, aseismic ridges, submarine plateaus and microcontinents. (Off-ridge volcanism was important in the Archean ocean because the ability of increased plate accretion to dissipate the estimated heat flux from the earth's interior was limited by the buoyant resistance to subduction of very young lithosphere.) The bathymetric highs, veneered atop by chemical sediments and aproned by indigenous clastics, were buried by kms of orogenic turbidites upon entry into the trench. Landward of the trench axis, the previous bathymetric highs and overlying trench turbidites were structurally detached (and foreshortened) from the underlying or surrounding "ophiolitic" crust and mantle, which were then subducted. The accretionary complex was later massively intruded by late-to post-tectonic plutons of the prograding magmatic arc, volcanic levels of which have been eroded away. As in Cenozoic arcs, variation in plutonic suites may be attributable to buoyant and non-buoyant subduction. The regional pattern of anastomosing "greenstone" belts may reflect the interference of first-order, NW and NNE striking, pinched synforms, spaced 70-120 km apart.

The model accounts for the evolutionary sequence of volcanism, sedimentation, deformation, metamorphism and plutonism, observed throughout the province. It accounts for both unconformable (trench inner-slope) and subconformable (trench outer-slope) relations between the volcanics and overlying turbidites. It admits the existence of relatively minor amounts of "pre-greenstone" basement (microcontinents) and "syn-greenstone" plutons (accreted arc roots). It predicts a variable age gap between "greenstone" volcanism and trench turbidite sedimentation (accompanied by minor volcanism). It also predicts systematic regional variations in age spans of volcanism and plutonism. An efficient test of the model would be a regional Sm-Nd study of the late plutons, predicting "syn- to post-greenstone" model ages for bulk crust-mantle separation.

Previous models (1,2), interpreting the "greenstone" belts as continental rifts, do not account for the observed deformation and metamorphism, nor for the myriad of late- to post-tectonic plutons, the ages of which cluster 40-100 Ma younger than the dated "greenstones" (3). They fail to explain the general absence of rift-type clastics in the lower volcanics and predict the inverse stratigraphic sequence from that observed (ie. subsidence and trench-type sedimentation preceding submarine volcanism, as the lithosphere progressively attenuates). They are incompatible with existing isotopic evidence (4,5) for massive crust-mantle separation following "greenstone" volcanism, and with evidence from detrital zircon dating (6) that the preponderance of turbidite source rocks were significantly younger than the "greenstone" volcanism.

Implications of the model will be illustrated with reference to a new 1:1 million scale geological map of the Slave Province (and its bounding 1.9 Ga orogens) compiled by the author as preparation for the "Decade of North American Geology" volume on the Canadian Shield.

(1) Henderson, J.B. (1981) in Precambrian Plate Tectonics, A. Kröner, ed., Elsevier, 213-236. **(2)** Easton, R.M. (1985) in Geol. Assoc. Can. Spec. Pap. 28, 153-168. **(3)** Green, D.C., and Baadsgaard, H. (1971) J. Petrology. 12, 177-217. **(4)** Robertson, D.K. and Folinsbee, R.E. (1974) Can. J. Earth Sci. 11, 819-827. **(5)** McCulloch, M.T. and Wasserburg, G.J. (1978) Science 200, 1003-1011. **(6)** Schärer, U. and Allegre, C. (1982) Can. J. Earth Sci. 19, 1910-1918.

GEOLOGICAL EVOLUTION OF THE PIETERSBURG GREENSTONE BELT, SOUTH AFRICA, AND ASSOCIATED GOLD MINERALIZATION; M.G. Jones, Department of Geology, Imperial College, London and M.J. de Wit, Lunar and Planetary Institute, Houston, Texas.

This poster presents current thoughts based on preliminary field work carried out as part of a Ph.D. project, the aim of which is to integrate the polyphase history of gold mineralization seen in the area with the geochemical and tectonic evolution of the greenstone belt as a whole.

Gold mineralization is found in four distinct regional geological settings;

1. A first phase of gold mineralization was associated with early low grade metamorphism and metasomatism of a 'greenstone basement' sequence of serpentinites (metaperidotites). These are generally intrusive into a series of BIF units, ferruginous shales and cherts. There are also associated extrusive tholeiitic metabasalts and ocellular-bearing komatiitic basalts. The regional hydration which characterizes this early metamorphism resulted in major chemical alteration of the basement and large scale fluid movement, with migration of Fe, and Mg ions, SiO_2 and possibly gold. Early shear zones (possibly represented by a now flat-lying carbonate-fuchsite-gneiss horizon) may have facilitated this fluid movement.

2. The basement sequence is unconformably overlain by a 'cover' of coarse clastic sandstones and conglomerates which contain basement-derived detritus. The conglomerates are often well sorted and graded and may represent coarse turbidites. Placer-type pyrite and BIF clasts, both containing minor gold values, are present in these cover rocks and hence a second period of gold mineralization (reworking) is envisaged.

3. The older rock sequences and gold mineralization above were all affected by a regional deformation event and it is the associated structural traps which contain the most significant gold occurrences seen in this greenstone belt. A well developed upright cleavage with a predominantly NE-SW strike and three major composite shear zones (each containing a number of tectonic breaks) are the main manifestations of this deformation. Strain analysis in the shear zones has been carried out using ocelli from the pillowed komatiitic basalts. The measurements indicate that close to or within the shear zones the finite strain ellipsoid results from a minimum of 50-70% flattening across the cleavage and 100 - 180% extension along the main stretching lineation seam.

Antitaxial and composite extension veins have been recognized. The veins contain fibrous crystals of quartz and calcite which plunge parallel to the stretching lineation (as defined by stretched conglomerate and breccia clasts lying in the cleavage plane). The veins are thus syn-kinematic with this main deformation event. The orientation of the quartz fibres is parallel to the incremental extension growth direction of the dilational veins and so the stretching lineation is parallel to the kinematic movemnt direction (approx. NW-SE when rotated to the horizontal). The veins are formed by the crack-seal fibrous growth mechanism and semi-quantitative strain analysis indicates clearly that the incremental strain ellipse (in the X-Y plane) did not change orientation significantly during the deformation event.

Field evidence indicates that the shear zones were thrusts (SE over NW) with both vertical and lateral components of movement. One of the shear zones, the Synmansdrift shear zone is marked by an unusual chaotic breccia which consists of white and brown-red banded chert and BIF clasts, identical to the BIF from the basement, set in a red ferruginous shale-like matrix. The clast content and size vary abruptly both across and along strike and there is no well-defined bedding. The lithology can be traced for 7km along strike and may be upto 100m in thickness but its upper (southern) boundary is ill-defined as it grades over tens of metres into conformably overlying but often highly deformed red shales and sandstones. Hence the upper contact appears to be sedimentary although this has yet to be confirmed by lithogeochemistry. The lower contact is clearly tectonic and an L-S tectonite fabric is well-developed. As well as small clasts, the lower half of the breccia also contains extremely large (upto 100m long x 20m wide) BIF inclusions which themselves are 'clasts' within the lithology. These larger inclusions can be clearly seen to be tectonically ground-up by a 'spalding-off' process which produces the smaller, often euhedral, breccia clasts.

As a whole the unit constitutes a tectono-sedimentary melange which is envisaged to have formed as a sedimentary wedge above a low dipping shear zone (thrust) <u>during</u> horizontal shortening across the region. Large scale movement of Fe ions, $SiO_2 \pm Au$ occurred.

Gold mineralization is found in quartz $\pm$ tourmaline veins associated with various structural traps e.g. fold hinges and minor shear planes including ultracataclasites. In the vicinity of these traps pressure solution and metamorphic segregation features are common which indicate fluid movement and possible gold mobility from the deformed sediments (and possibly the basement rocks) into the traps. This fluid migration may have occurred early with respect to the deformation with the resultant veins being subsequently slightly deformed and tectonically displaced probably later but within the same deformational event.

4. A later porphyroblastic overprint of gold-bearing arsenopyrite is seen locally within the shear zones as well as porphyroblasts of ephesite (a lithium-bearing brittle mica) and andalusite. These features seem to indicate a later period of gold mineralization and 'static' metamorphism probably related to granitic and pyroxenitic intrusions which provided a heat source (and possibly fluids) for element mobility and mineralization within the already deformed volcano-sedimentary pile.

SYNSEDIMENTARY DEFORMATION AND THRUSTING ON THE EASTERN MARGIN OF THE BARBERTON GREENSTONE BELT, SWAZILAND, Lamb, S.H., Research School of Earth Sciences, Victoria University of Wellington, Private Bag, Wellington, New Zealand.

Mapping on the eastern margin of the 3.6-3.3 Ga Barberton Greenstone Belt, NW Swaziland, has revealed a tectonic complex which is more than 5 km thick (Lamb, 1984a). The area consists of fault bound units made up of three lithological associations. Some of these have been affected by four phases of deformation (D1-D4). Fold structures (F1-F4), foliations (S1-S4), and lineations are associated with the deformation.

The oldest rocks consist of metaigneous rocks (talcose schists, serpentinite, and quartz-chlorite-sericite schists) interleaved with silicified fine grained sediments (cherts). These make up the Onverwacht Group, though deformed (D1) and intruded by meta-ultramafic rocks. Onverwacht Group cherts locally pass conformably into a circa 1.8 km thick sequence of siltstones, shales, BIF, with sandstone and conglomerate layers, forming the Diepgezet Group. The lower part of the Diepgezet Group is interpreted as submarine fan deposits, and can be correlated with sequences in South Africa referred to as both the Moodies and Fig Tree Groups (Lamb and Paris, in prep). The Diepgezet Group is overlain unconformably, with angular discordances of up to 90 degrees, by at least 1.8 km of coarse clastics (Malalotsha Group). These are interpreted as fluvial and marginal marine deposits. In certain localities the Diepgezet Group passes up conformably into the Malalotsha Group through a sequence of coarse sediments which have been left undifferentiated (Mal/Diep Group). Parts of the Malalotsha Group can be correlated with the Moodies Group.

Three pronounced angular unconformities occur within the basal 1000m of the Malalothsha Group. Malalotsha Group sediments are both folded by, as well as unconformably overlying, D2 fold structures which deform the Diepgezet and Onverwacht Groups. Folded fault zones (D1) juxtaposing the Diepgezet and Onverwacht Groups are also unconformably overlain by the Malalotsha Group. Faults associated with the F2 folding (flexural slip faults) offset Malalotsha Group sediments, but are also unconformably overlain by younger Malalotsha Group sandstones and conglomerates. In sequences where the Malalotsha Group is transitional with the Diepgezet Group, a progressive change is observed in the clast content of the sandstones. Chert grain dominated sandstones within the Diepgezet Group pass up into sandstones made up mainly of single crystal quartz grains. Clasts representing all the underlying stratigraphy, as well as parts of the gneissic terrain (potassium poor granitoids) are found in Malalotsha Group conglomerates. Palaeocurrents within the basal Malalotsha Group indicate polymodal sediment transport directions. This, combined with evidence for rapid sediment thickness changes and facies variation, suggest that these sequences were deposited in tectonically controlled (and actively deforming) basins. However the overall tectonic setting is not clear, though the sediments were clearly deposited in a compressional regime.

The sedimentary sequences described above are now found within thrust sheets up to a kilometre thick. These are bounded by thrust faults, subparallel to bedding, which juxtapose different parts of the stratigraphy. One of these thrusts emplaces part of the Onverwacht Group on top of the Malalotsha Group, with a displacement of more than 10 km. The Onverwacht Group here contains a low angle foliation (S2) subparallel to the bounding fault. The thrust faults are considered to be a later expression of the D2 deformation, which is seen as syn-sedimentary deformation structures within the thrust sheets. The D2 deformation caused shortening in northerly and westerly directions.

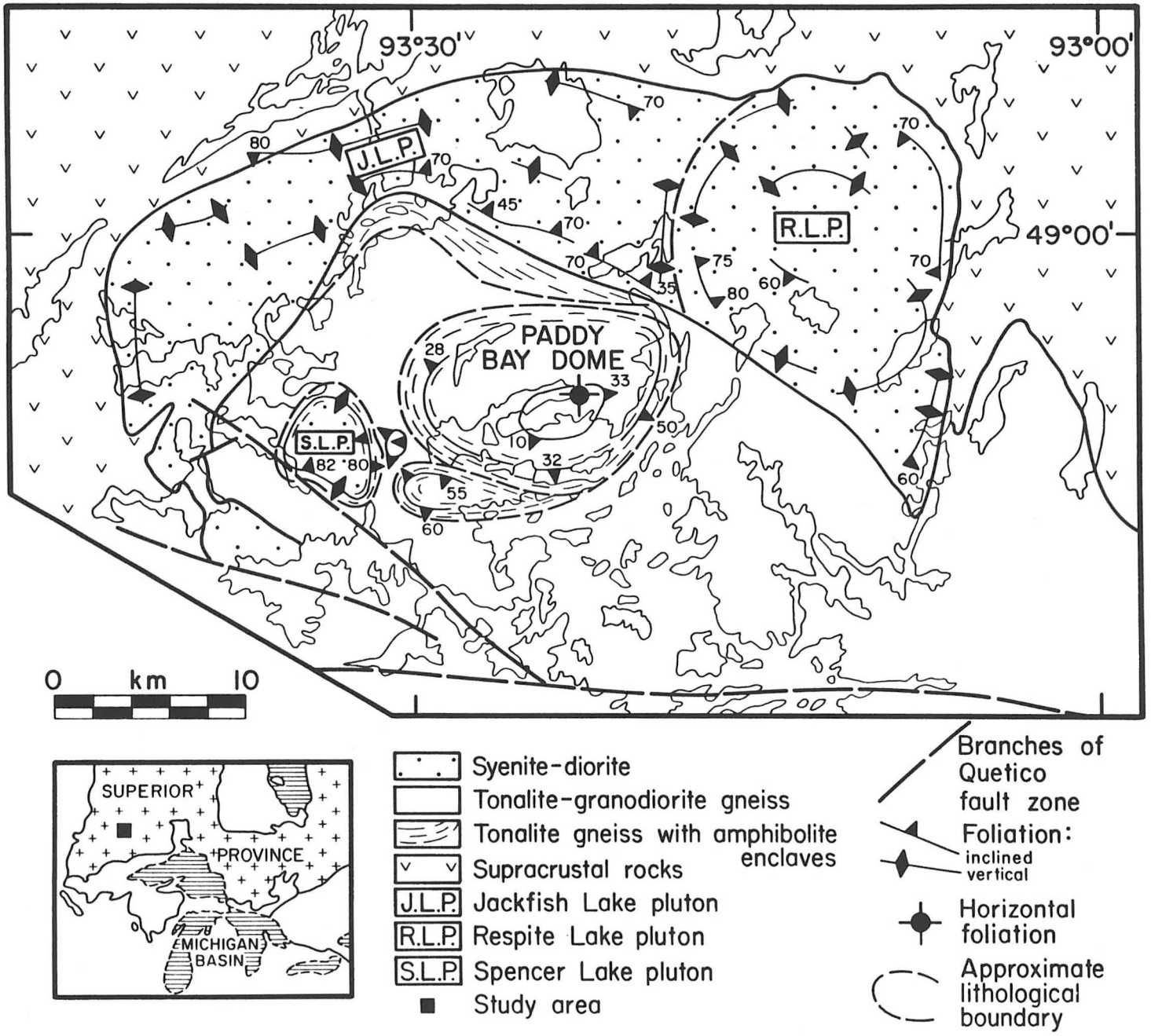
The thrust sheets and their internal structures have been refolded by tight

kilometre scale north trending folds, which plunge south at 20-40 degrees (F3). The folds contain a pronounced axial planar cleavage defined in places by a muscovite schistosity. The cleavage is most intense near and within marginal granitoids which were probably intruded c. 3.0 Ga (part of the Mpuluzi batholith, Barton 1981). Earlier fold structures have been tightened up, intensifying an axial planar cleavage fabric in F2 folds (S2/S3). The contact with intrusive granitoids on the western margin of the study area (Steynsdorp pluton, which may be c. 3.4 Ga, Barton 1981) contains a pronounced foliation which cuts across intrusive contacts. This is interpreted as an S3 foliation which contains an intersection and/or stretching lineation plunging at 20-40 degrees NE. The apparent domal pattern of foliations in the marginal parts of the Steynsdorp pluton is interpreted as both the result of F3 folding of an earlier foliation (S2) and also the imprint of an S3 foliation. Elongation lineations in sediments within the greenstone belt may be a result of subvertical extension during the D3 shortening (e.g. Jackson and Robertson, 1983).

The above structures have been refolded by heterogeneous southeast trending folds (F4) with the local development of an L4 crenulation lineation.

It has been suggested (Lamb, 1984a,b) that the high level syn-sedimentary D2 deformation and subsequent development of a thrust complex was related to coeval deformation and metamorphism (Jackson, 1984) in the Ancient Gneiss Complex of southern Swaziland. D2 in the study area predates the c. 3.0 Ga Mpuluzi batholith. It is not clear what the relation was between D2 and an early D1 deformation, which occurred during the evolution of the Onverwacht Group rocks (de Wit, 1982; pers. com.). It is likely to be close as a continuous depositional sequence is preserved between the Onverwacht and Malalotsha Groups. The correlation of clastic sequences in the southern part of the greenstone belt with those in the study area, indicates that the D2 deformation was diachronous with variable structural trends. The presence and position of unconformities show that NW-SE shortening (D2b) and the deposition of the Malalotsha Group in the study area post-dates the deposition of the Moodies Group and N-S shortening (D2a) observed in the southwestern part of the greenstone belt (de Wit et al., 1983). It is however not clear to what extent the D2b shortening has reworked and translated structures which formed in D2a. Subsequent D3 deformation (coeval with the intrusion of the Mpuluzi batholith, c.f. Jackson and Robertson, 1983) has had a considerable effect on structures in the study area, continuing the shortening (E-W) on the eastern margin of the greenstone belt.

References: (1) Barton, J.J. Jr., 1981. In: Anhauesser, C.R. (ed.), Barberton Excursion Guide, Geol. Soc. S. Afr. (2) de Wit, M.J., 1982. Gliding and overthrust nappe tectonics in the Barberton greenstone belt. J. Struct. Geol., 4: No. 2, 117-136. (3) de Wit, M.J., Fripp, R.E.P., and Stanistreet, I.G., 1983. Tectonic and stratigraphic implications of new field observations along the southern part of the Barberton greenstone belt. Spec. Publ. Geol. Soc. S. Africa., No. 9: 21-29. (4) Jackson, M.P.A., 1984. Archaean structural styles in the Ancient Gneiss Complex of Swaziland, southern Africa. In: Kroner, A. (ed.), Precambrian Tectonics Illustrated. E. Schweizerbart, Stuttgart, West Germany. (5) Jackson, M.P.A., and Robertson, D.I., 1983. Regional implications of Early-Precambrian strains in the Onverwacht Group adjacent to the Lochiel Granite, North-West Swaziland. Spec. Publ. Geol. Soc. S. Africa, No. 9: 45-62. (6) Lamb, S.H., 1984a. Geology of part of the Archaean Barberton Greenstone Belt, Swaziland. Unpublished Ph.D thesis, University of Cambridge, England. (7) Lamb, S.H., 1984b. Structures on the Eastern Margin of the Archaean Barberton Greenstone Belt. In: Kroner, A. (ed.), Precambrian Tectonics Illustrated. E. Schweizerbart, Stuttgart, West Germany.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R.L.P.
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PADDY
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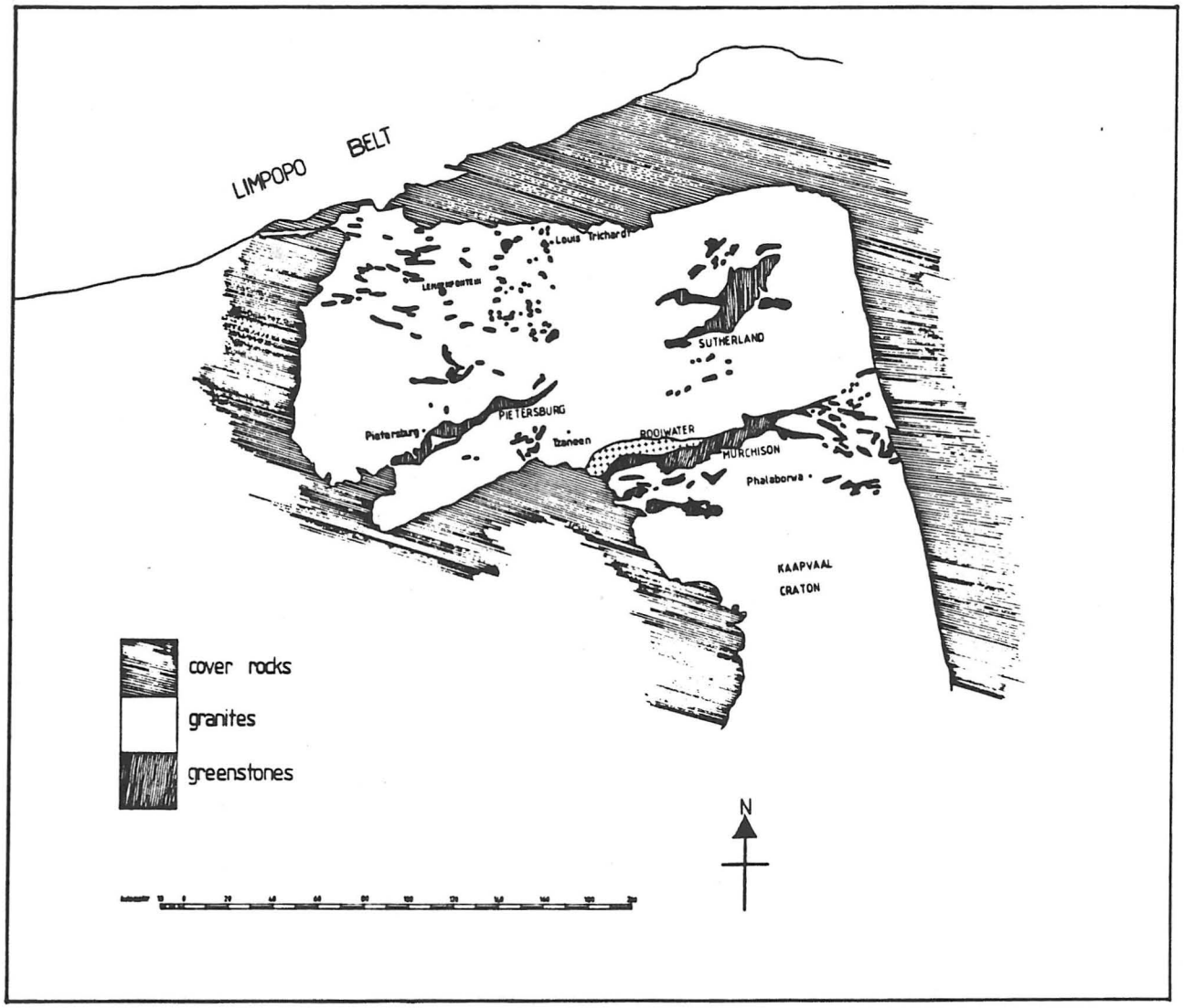
SUPERIOR
PROVINCE
MICHIGAN
BASIN
Syenite-diorite
Tonalite-granodiorite gneiss
Tonalite gneiss with amphibolite enclaves
Supracrustal rocks
J.L.P. Jackfish Lake pluton
R.L.P. Respite Lake pluton
S.L.P. Spencer Lake pluton
Study area
Branches of Quetico fault zone
Foliation:
inclined
vertical
Horizontal foliation
Approximate lithological boundary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ROOIWATER
MURCHISON
Phalaborwa
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CRATON
cover rocks
granites
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