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WORKSHOP ON THE TECTONIC EVOLUTION OF GREENSTONE BELTS

(Supplement Containing Abstracts Of Invited Talks and Late Abstracts)

HOUSTON, TEXAS

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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

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. <u>Mantle Temperatures</u>

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

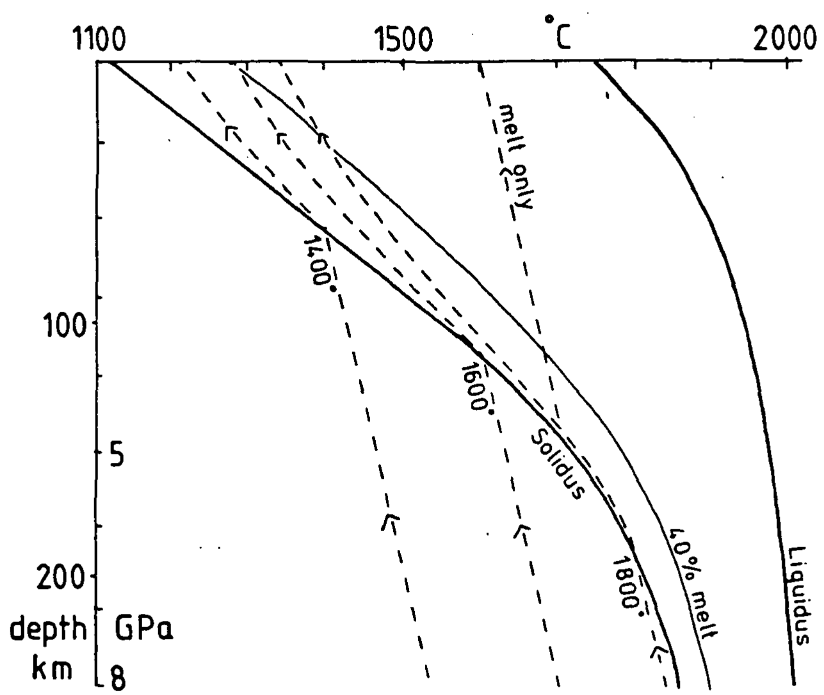

Figure 2. Mantle liquidus and solidus and adiabatic ascent paths calculated with the assumption that melt and solid do not segregate on ascent, after McKenzie and Bickle [23].

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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.

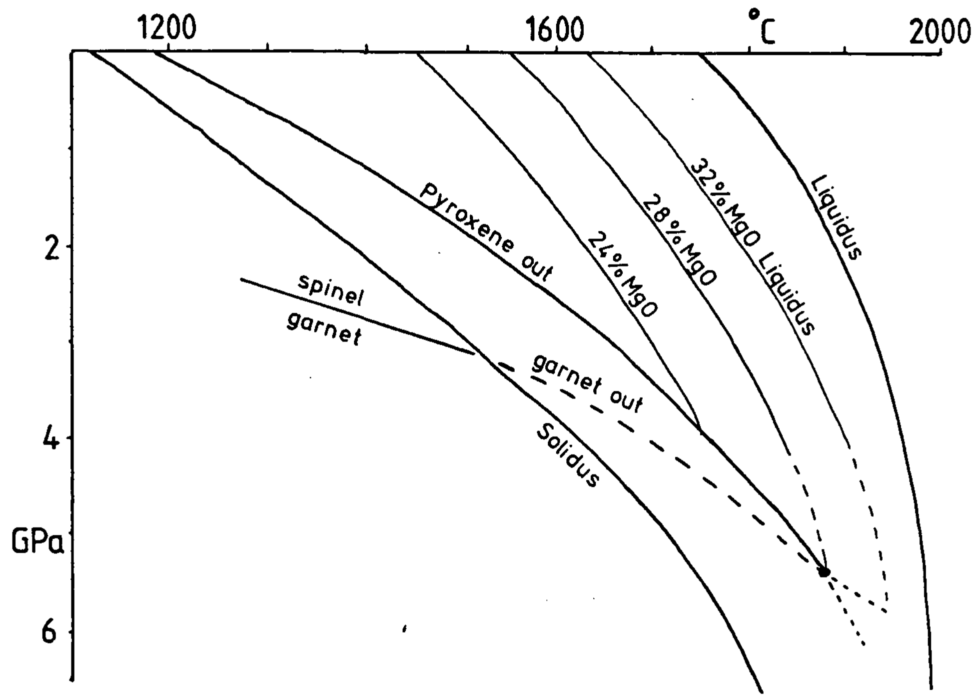

Figure 3. Phase relations for melting mantle-like compositions from experiments on komatiites. Note intersection of garnet melting curve with the pyroxene melting curve is hypothetical.

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

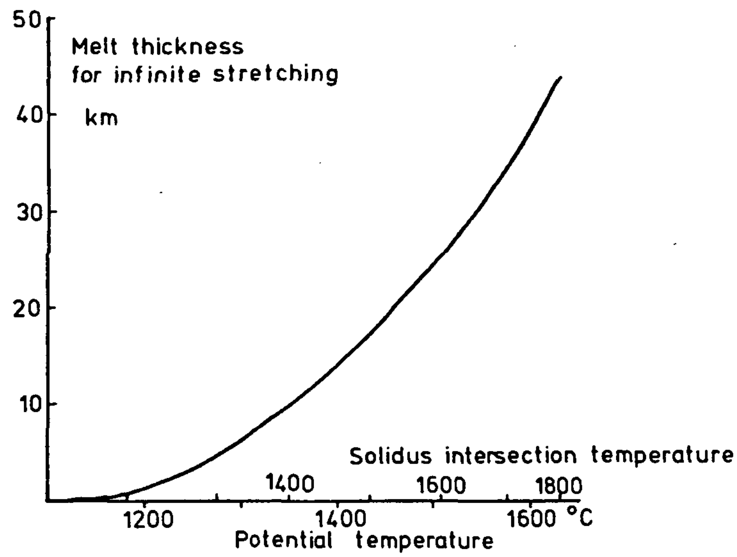

Figure 4. Melt thickness as a function of mantle temperature for infinite stretching (oceanic ridge case) after McKenzie & Bickle [23].

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.

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ABSTRACT FORM - FA...

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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 postdepositional 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.

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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.

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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 staurolitekyanite-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).

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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.

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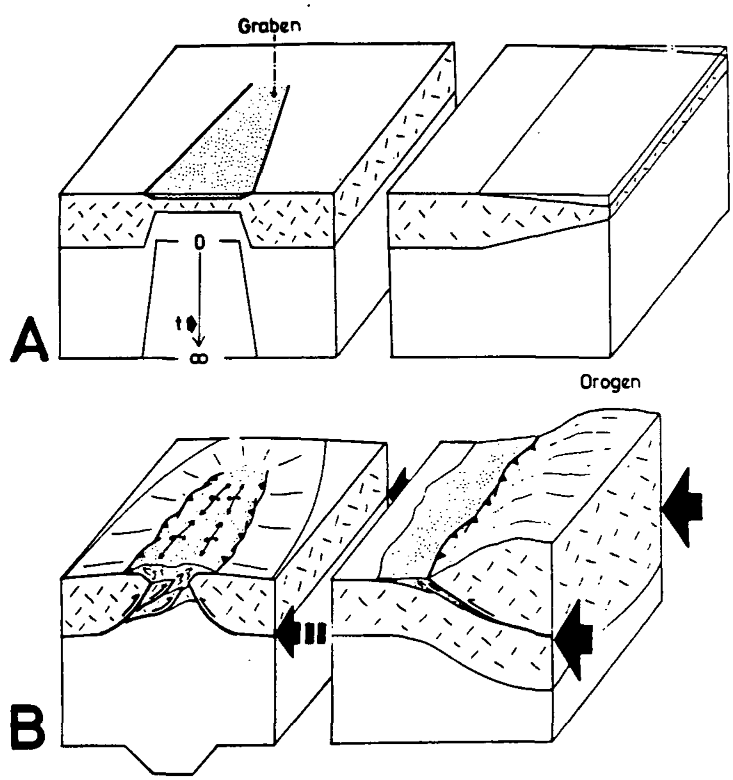

Figure 1. (from ref. 6) Illustration of how rifts within a continent (such as the Dneipr-Donetz rift) have been affected by neighboring continental collisions (as the Dneipr-Donetz structure responded to collisions in North Dobrudja in the Early Jurassic). Observation has shown that folding and thrusting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

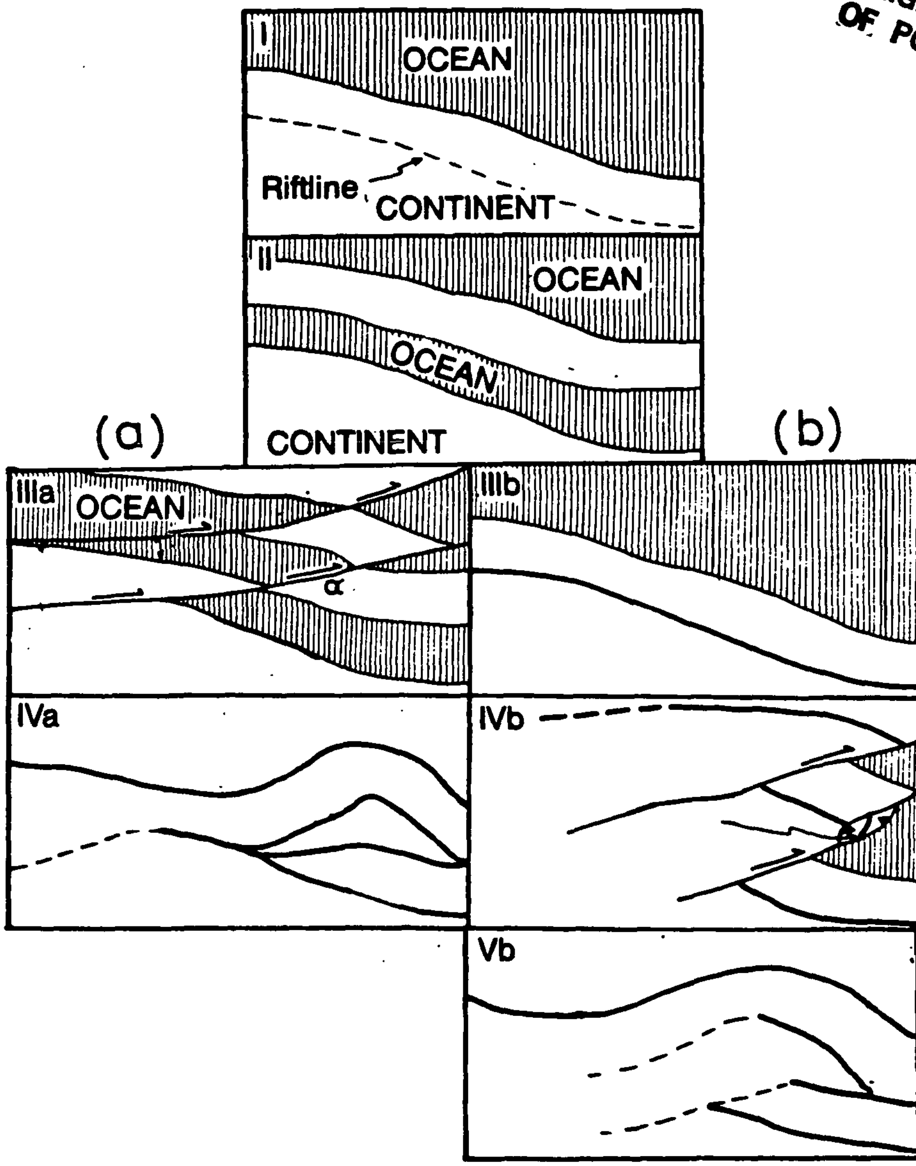

Figure 2. (from ref. 6) A possible origin for some greenstone belts. Rifting (I) takes a continental fragment out into an ocean (II). Major strike-slip motion (IIIa) is depicted as preceding collision between slivers of the continental fragment and the main continent (IVa). As an alternative suturing 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.

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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.

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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.

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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.

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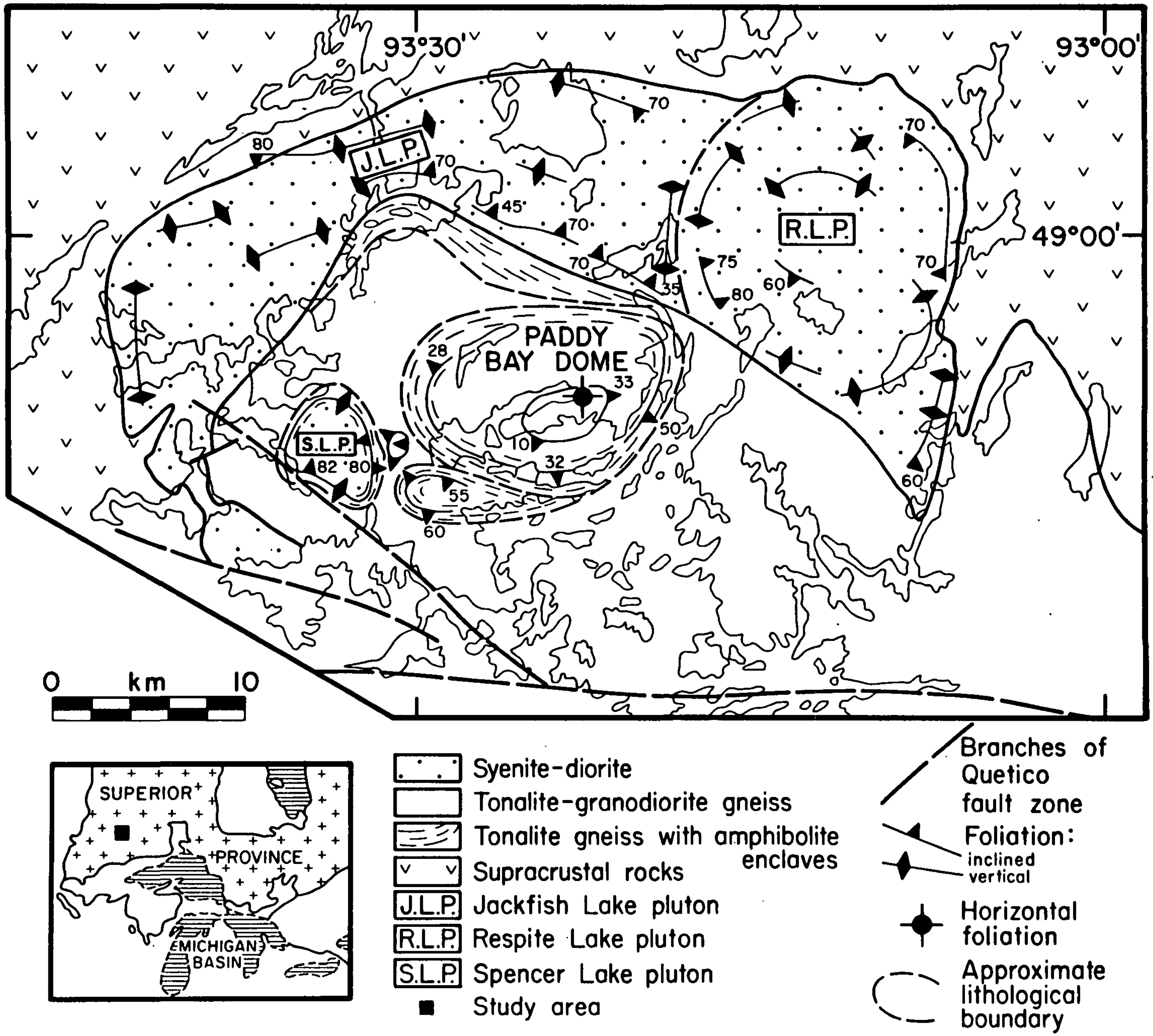
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J.L.P.
R.L.P.
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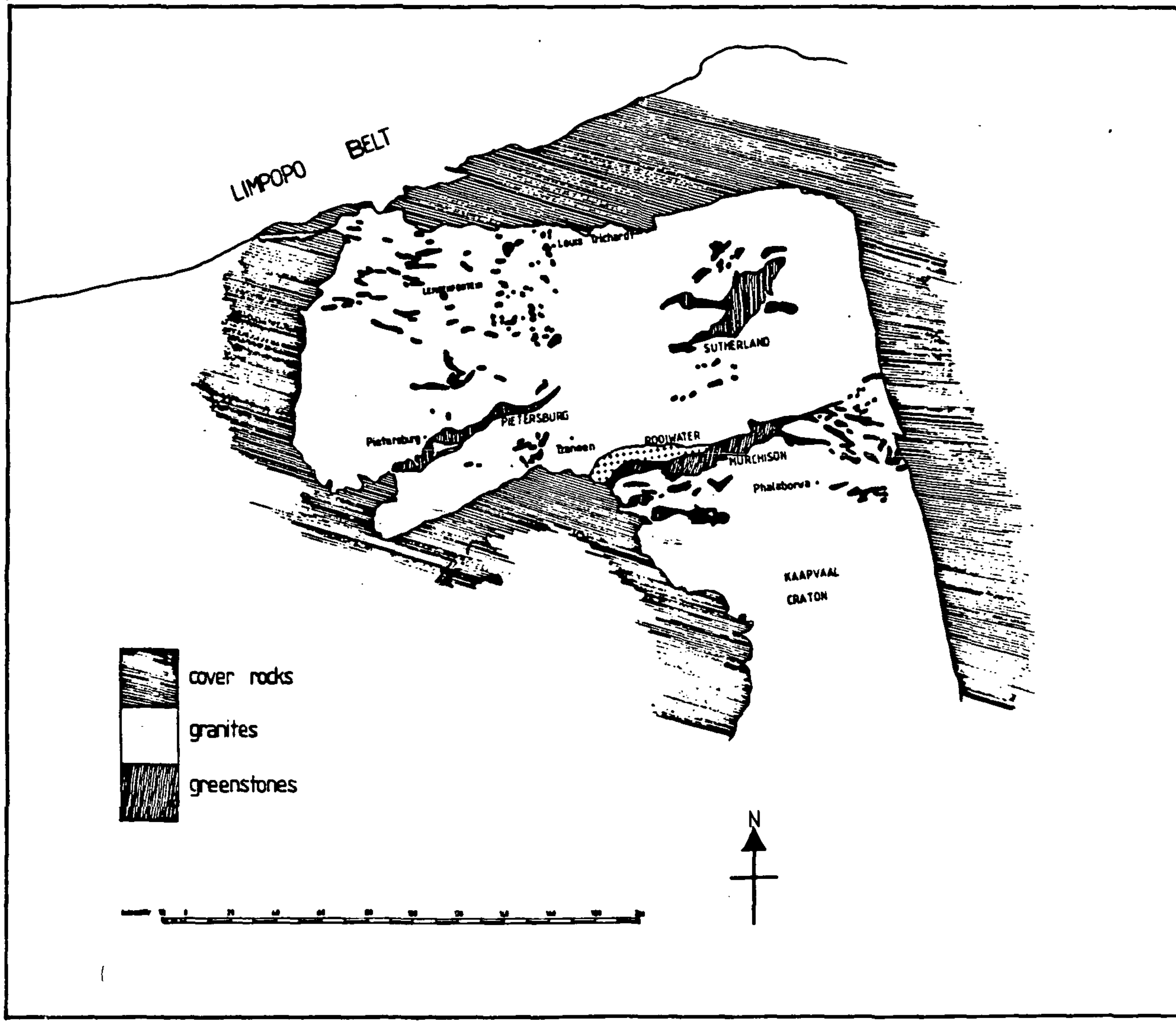
PADDY BAY DOME
0 km 10
SUPERIOR
PROVINCE
MICHIGAN BASIN
Syenite-diorite
Tonalite-granodiorite gneiss
Tonalite gneiss with amphibolite enclaves
Supracrustal rocks
J.L.P. Jackfish Lake pluton
R.L.P. Respite Lake pluton
S.L.P. Spencer Lake pluton
Study area
Branches of Quetico fault zone
Foliation: inclined vertical
Horizontal foliation
Approximate lithological boundary



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