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"Relationship between High- and Low-Grade Archean Terranes:
Implications for Early Earth Paleogeography"

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INTRODUCTION

This study was undertaken in the Western Gneiss Terrain (WGT) of the Yilgarn Block, Western Australia. The WGT forms an arcuate belt of Archean gneisses that flank the western margin of the Yilgarn Block. In general the WGT is composed of high-grade orthogneisses and paragneisses which contain supracrustal belts composed largely of siliciclastic metasediments and subordinate iron formation. The platformal nature of the metasedimentary belts and lack of obvious metavolcanic lithologies contrasts with the composition of typical Yilgarn greenstones to the east. Radiometric data from WGT rocks indicates that these rocks are significantly older than Yilgarn rocks to the east (>3.3 Ga) and this has led to the suggestion that the WGT represents sialic basement to Yilgarn granite-greenstone belts.

The Mount Narryer region exposes the northernmost occurrence of high-grade metasediments within the WGT and consists of quartz-rich clastic metasediments at upper amphibolite to granulite grade. Most occurrences of supracrustal rocks in this region comprise isolated lenses within the gneissic basement. However, at Mount Narryer a unique sequence of metaclastics with preserved bedding provide an unusual window into the parentage of similar supracrustal bodies in this region.

Curiosity in this region was originally generated by radiometric data which indicated the relative antiquity of the rocks. A Rb-Sr whole rock isochron on gneissic basement rocks gave an age of 3348 ± 43 Ma, which was interpreted as the time of metamorphism of the gneisses. This work has been amplified and substantiated by subsequent studies of U-Pb systematics in zircons from gneissic rocks which indicate igneous crystallization at 3750 Ma and 3680 Ma followed by metamorphism at .3350

Ma. Ion microprobe analyses of detrital zircons in metasediments from Mount Narryer are consistent with derivation of the zircons from the underlying basement (zircon ages of 3750-3100 Ma) but also indicate the presence of much older basement component of 4100-4200 Ma. The presence of basement zircons in the supracrustal rocks and apparent differences in deformation history led Myers and Williams of the Geological Survey of Western Australia to suggest that the gneisses represent depositional basement to the sediments.

The Mount Narryer region is unique in the WGT for its unequivocal preservation of primary bedding and abundant sedimentary structures, thus providing a rare opportunity for sedimentologic analysis. However, the preservation of primary bedding and stratigraphic facing indicators is perhaps of even greater significance in terms of providing an important fabric element which can be used in structural analysis. The problem of describing the geometry of the strain within rocks and attempting to infer even a preliminary kinematic evolution is greatly aided by preservation of a fabric which precedes the widespread bedding transposition of many supracrustal bodies in the WGT. Fundamental questions regarding the WGT may be addressed by both detailed and regional structural analysis. What is the relationship between basement gneisses and supracrustal assemblages? The boundary between the WGT and Yilgarn is commonly a broad zone of mylonitic deformation. Can the structural history in the WGT provide insight into the origin of this boundary. The structural evolution of the Narryer regin may provide the basis for further work of a more regional aspect.

REGIONAL GEOLOGY

Metasedimentary rocks in the Mount Narryer region of the Western Gneiss Terrain, Yilgarn Block, Western Australia occur as a north-trending lens-shaped zone of low strain, about 2.5 km wide and 20 km long, that is in mylonitic contact with basement gneisses. In the northern end of the belt, the width of the metasediments decreases to a few hundred meters, concommitant with an increase in the intensity of strain and convergence of bounding mylonite zones. The eastern boundary between Mount Narryer metasediments and basement gneisses is marked by a north-trending west dipping zone, up to 200 m wide, of mylonitic quartzite (Elizabeth Spring mylonite zone). The western contact of the metasediments with basement gneiss (Meebeerie gneiss) is a broad zone of distributed shear which extends for several hundred meters into the metasediments but does not appear to affect a similar thickness of gneissic rocks. The overall distribution of strain in the Mount Narryer region is broadly similar to a boudin with intensely-sheared margins and attenuated terminations that enclose a central zone of comparatively low strain.

STRATIGRAPHY AND SEDIMENTOLOGY

The metasedimentary succession at Mount Narryer consists largely of impure quartzites, metaconglomerate, pelitic gneiss and subordinate calcareous gneiss which represent metamorphosed shallow-water sediments. The abundance of primary sedimentary structures within the lower units of this succession (units A-C) indicate that the stratigraphy is right-way-up. Structural deformation has resulted in locally pronounced thinning and thickening of units which obscures primary stratigraphic relationships and thicknesses. In addition the presence of folded

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mylonitic shear zones within the metasediments hints at the possibility of structural duplication which cannot be accurately resolved in the absence of better marker beds. We suspect that the stratigraphic succession is a mix of both structural and as well as depositional interleaving.

A schematic stratigraphic column is shown in Figure 1; thicknesses are averages. Unit A represents an apparent upward-coarsening succession of impure sillimanite-garnet quartzites, feldspathic quartzites and local lenses of calcareous quartzite that grade into interbedded sheets of quartz-pebble conglomerate and impure quartzite. Unit B is composed largely of sillimanite-garnet-cordierite quartzites with well-developed cross-bedding; also present are subordinate feldspathic quartzites, quartz-pebble conglomerates and a single lens of semipelitic gneiss. Unit C is marked by the first appearance of polymitic conglomerates interbedded with cross-bedded garnet-sillimanite-cordierite quartzites. The conglomerates occur as sheets and lenses containing clasts of banded magnetite-chert, vein quartz, garnet-quartz rock and locally sillimanite-rich, vein quartz. The garnet-quartz clasts are typically angular and are interpreted as original pelitic rip-up clasts. Upper Unit C quartzites interfinger with pelitic gneiss and conglomerates (?) of Unit D which are characterized by the occurrence of distinctive "faserkiesel" of intergrown sillimanite and quartz. Unit D is abruptly overlain by sparsely sillimanitic quartzites of Unit E; primary stratification cannot be identified in either of units D or E which occur in the core of the Narryer syncline.

Cross-stratification in arenaceous beds and primary clast shapes in conglomerates are remarkably well-preserved particularly in Units B and C despite deformation and granulite facies metamorphism (Fig. 2).

Cross-bedding is commonly defined by concentrations of red-brown metamorphic garnets and hematite along foresets. Excellent three-dimensional exposures of crossbeds show that they range in shape from tabular planar to trough with generally unimodal foreset orientations. The complete lack of flattening or distortion of the crossbed shapes by penetrative deformation is emphasized as this bears on structural history of the region. The lack of penetrative deformation is also shown by the conglomerates which, between the mylonite zones, show essentially no evidence of strain except for local rigid-body rotation of some clasts into the plane of weak cleavage.

The lenticularity of stratigraphic units is the result of both depositional and structural processes. Detailed mapping of conglomerate beds and associated arenaceous sediments in Unit C demonstrates a primary lenticularity (Fig. 3) reflecting a channel and bar-type geometry common in many fluvial systems. In contrast, lenticular semipelitic units, such as the lens in Units B and the faserkiesel gneisses of Unit D, are the result of tectonic thickening and thinning. This is indicated by the internal strain within these units which characteristically have well-developed penetrative fabrics and small-scale folds, reflecting the greater ductility of these impure lithologies. In addition, some lenticular packets are bounded by mylonitic shear zones and are therefore clearly structural in origin.

STRUCTURE

Kinematic analysis of mylonites and documentation of the geometry of strain within the Mt. Narryer rocks suggests that this belt represents the remnants of a mid-crustal ductile fold-and-thrust belt. Early deformation was partitioned into the metasediments resulting in mylonitic detachment from basement and synmetamorphic imbrication along ductile thrusts. Progressive simple shear led to formation of a coaxial recumbent anticline-syncline pair which folded early mylonites and rotated into parallelism with the X direction of the finite strain ellipsoid, as inferred from lineations in mylonites. A set of late mylonites contain kinematic indicators suggestive of east-directed thrusting; this shear sense is inferred to have persisted throughout evolution of the Narryer belt. Folding of basement and cover or fault-bend rotation led to steepening of mylonitic shear zones and formation of the synformal geometry of this belt. Strain partitioning into the mylonite zones is considered to have spared most of the primary sedimentary structures from deformation. If this sense and style of deformation is present on a regional scale it may have significant bearing on the nature of the mylonitic boundary between the Western Gneiss Terrain and granite-terrain of the Yilgarn Block.

E
garnet-sillimanite quartzite

D
"Faserkiesel" pelitic gneiss with quartzite interbeds at base

C
crossbedded garnet-sillimanite quartzite and polymict conglomerate (Co)

B
crossbedded garnet-sillimanite quartzite (Bq) with minor feldspathic quartzite and quartz pebble conglomerate and (tectonic) lens of pelitic gneiss (Br)

A
interbedded feldspathic quartzite (Af), calcareous quartzite (Ap), sillimanite-garnet quartzite (Aq) and quartz pebble conglomerate (Ao)

DÉCOLLEMENT
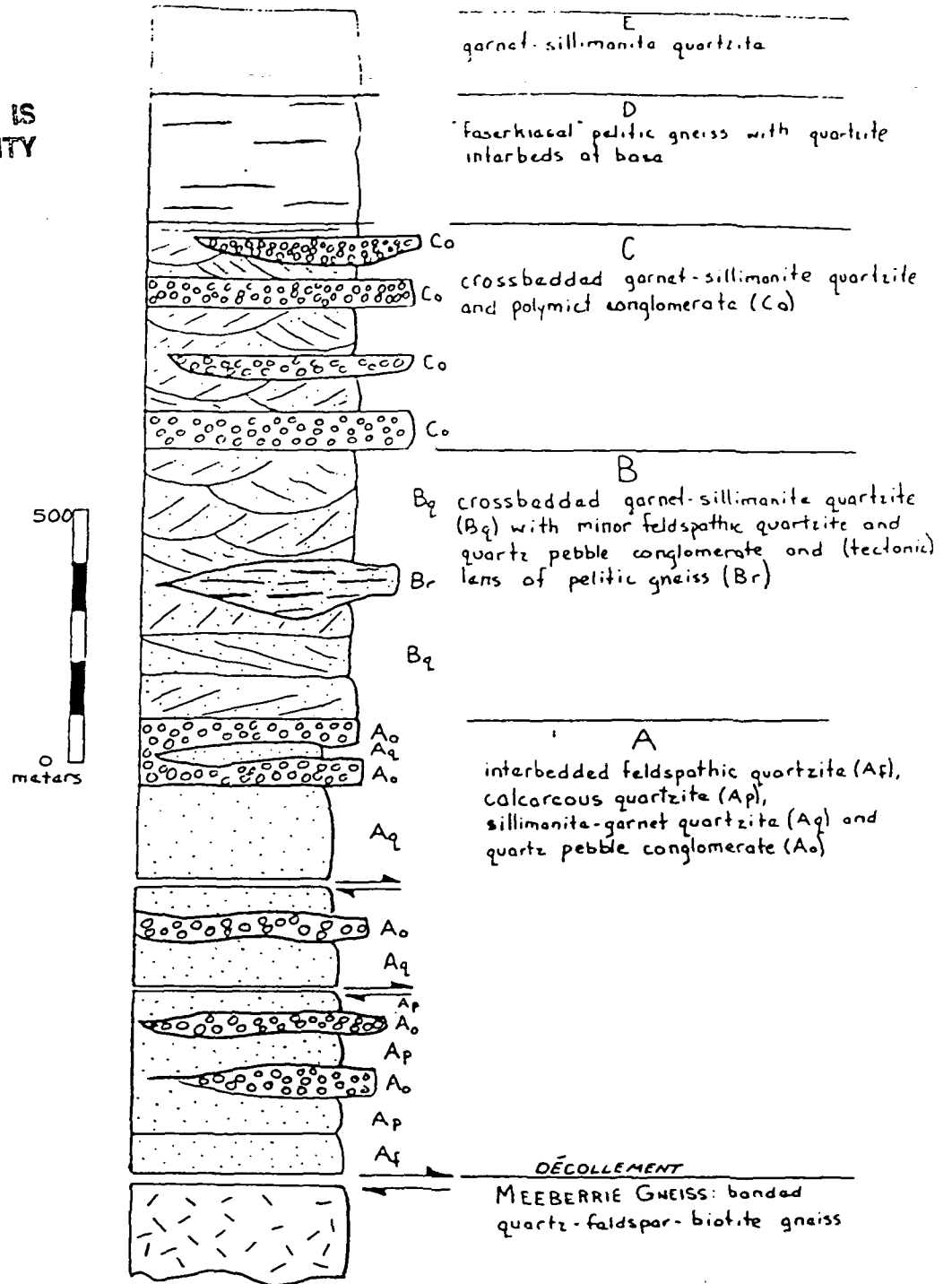MEEBERRIE GNEISS: banded quartz-feldspar-biotite gneiss

Figure 1:   Schematic lithostratigraphic section of the metasedimentary
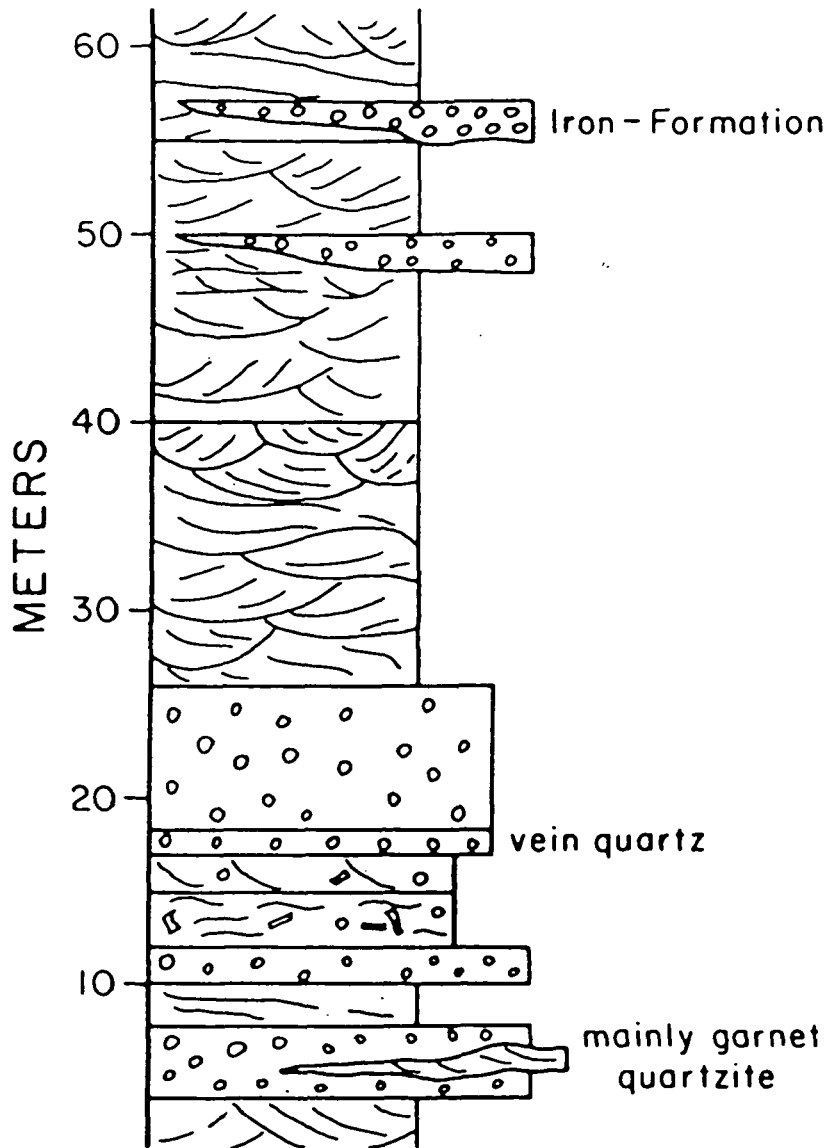   sequence at Mt. Narryer.

Figure 2: Measured stratigraphic column through part of Unit C illustrating the use of conventional sedimentology despite deformation and granulite facies metamorphism.
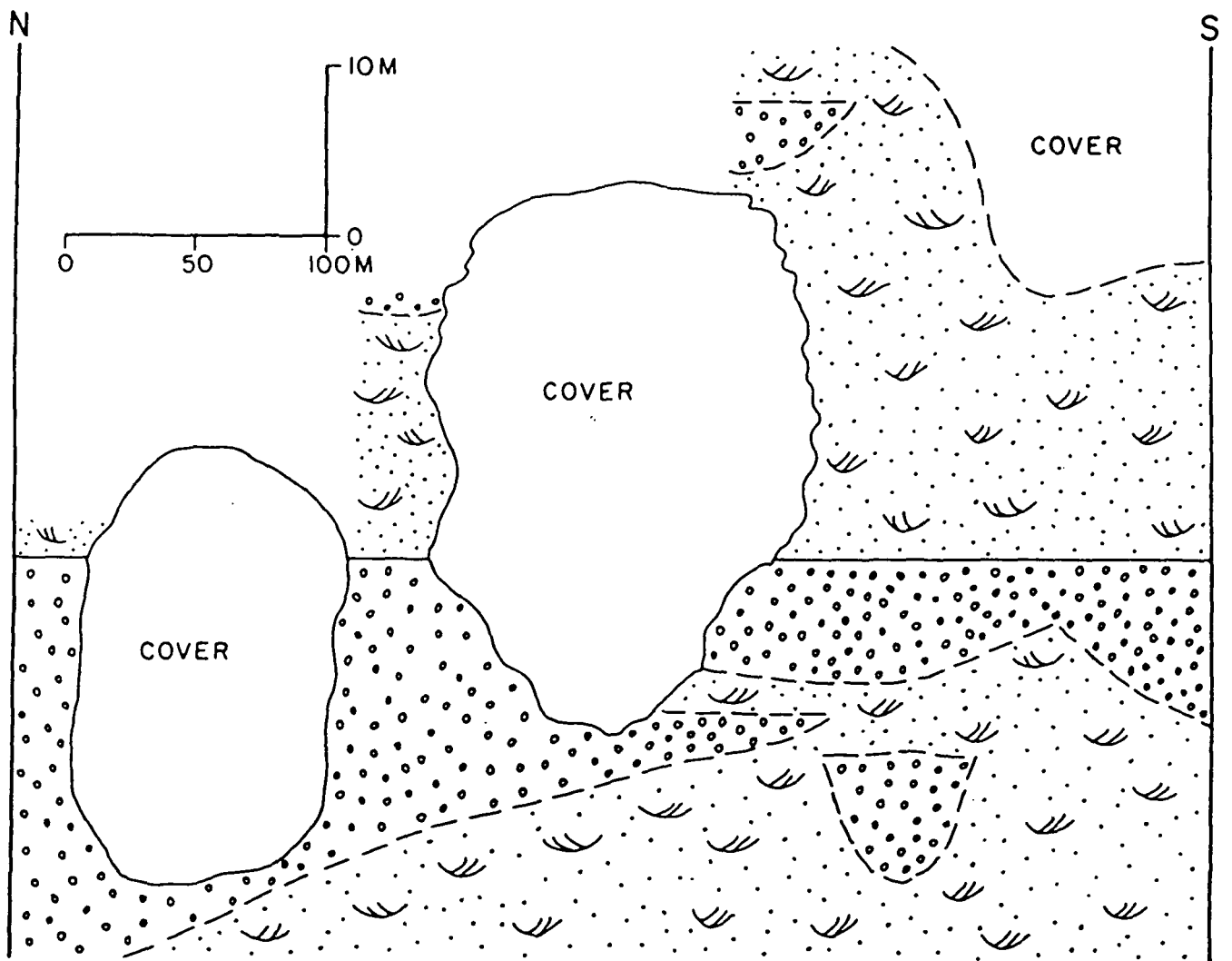
Figure 3: Cross-section of part of Unit C showing syndepositional lenticularity
of conglomerates within sandstones. This association is characteristic
of braided fluvial channel and bar deposition.

BASIN ANALYSIS IN METAMORPHOSED AND STRUCTURALLY-DISTURBED TERRANES: PROBLEMS AND PERSPECTIVES WITH EXAMPLES FROM THE EARLY ARCHEAN OF SOUTHERN AFRICA AND WESTERN AUSTRALIA

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Early Archean (> 3.2 b.y. old) sediments are present in high-grade metamorphic and granite-greenstone terranes in southern Africa and Western Australia. Analysis of these sediments places constraints on early crustal evolution and some, or all, of source area composition and location, depositional environments, basin configuration and tectonic setting depending on the degree of preservation of the rocks.

Pre-greenstone (> 3.6 b.y. old) gneisses in the high-grade Limpopo Province and West Yilgarn Block represent basement to cover rocks which are exclusively of sedimentary origin and accumulated between 3.6 and 3.3 b.y. ago. A thick sequence (~ 2.0 km) of conglomerates and cross-bedded aluminous gneisses (wacke) in the West Yilgarn is interpreted as a probable rift deposit. The Limpopo cover rocks are devoid of primary sedimentary structures and consist of metaquartzite with detrital zircon (original quartz arenite), marble (limestone), metapelite (mudstone) and aluminous gneiss (possible wacke). This quartz arenite-carbonate association implies a stable tectonic setting and the best analogues may be younger passive-margin deposits. Metapelites in the Limpopo Province have a complex geochemistry indicating a mixed provenance which included differentiated sialic crust.

Coeval with sedimentation in the Yilgarn and Limpopo regions at ca. 3.5 b.y. ago, predominantly ultramafic and mafic volcanism in the Barberton and Pilbara greenstone terranes took place in an anorogenic, oceanic environment distant from any continental influence. Intercalated sediments are mostly of airfall pyroclastic origin and have been interpreted as both shallow- and deep-water deposits. Evaporites and biogenic sediments, including rip-up algal mats and stromatolites, associated with the volcanics may support the shallow-water interpretation. The greenstone volcanics are overlain by terrigenous sediments which, in the Barberton Mountain Land, display an evolutionary pattern similar to Phanerozoic foredeeps. Basin deepening due to crustal loading from the south accompanied the onset of foredeep development. At this stage, basinal iron-formation, chert and mudstone of the basal Fig Tree Group accumulated conformably on the underlying volcanics. Overlying sediments of the Fig Tree Group consist of ~ 2 km of submarine-fan greywackes and mudstones. Shoaling of the basin is indicated by the transition to the overlying braided-alluvial and shallow-marine sediments of the Moodies Group (~ 3 km thick). Changing REE patterns and relative abundances of Ni, Cr, Th and Ur in the sediments of the Fig Tree and Moodies Groups indicate a significant ultramafic-mafic volcanic component diluted upward in the stratigraphic sequence by an increasing granitic contribution. A similar trend is reflected in sandstone petrography, in which the ratio of extrusive to intrusive rock fragments decreases upwards in the Fig Tree Group whereas the Moodies Group sediments are enriched in quartz and potassium feldspar. A persistent chert component in both units suggests that the tectonic setting of the basin is analogous to that of Phanerozoic recycled orogens. The distribution of facies in the Fig Tree and Moodies Groups, in conjunction with paleocurrent, petrographic and geochemical data, indicate progressive unroofing of a southerly source terrane consisting of the older volcanics with intercalated sediments, and a 3.5-3.3 b.y.-old gneiss complex. Uplift in the source area was by crustal shortening; this deformational front subsequently advanced northwards to involve the greenstone belt in thrust-nappe tectonics by 3.2 b.y. ago.

ARCHEAN SEDIMENTARY ASSOCIATIONS: A REVIEW.

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Archean sedimentary associations are developed in high-grade metamorphic, granite-greenstone and cratonic terranes. Pre-greenstone (> 3.6 Ga) gneisses in the Limpopo Province and West Yilgarn Block represent basement to ca 3.6-3.2 Ga high-grade supracrustal rocks (Association 1). The supracrustals are exclusively of sedimentary origin, consisting in the Limpopo Province of metaquartzite (original quartz arenite), marble (limestone) and metapelite (mudstone), and in the West Yilgarn of conglomerate and garnet-sillimanite gneiss (original wacke) with well-preserved sedimentary structures. The quartzite-carbonate association implies a stable-shelf setting whereas the conglomerates and wackes are of braided-alluvial origin and may have accumulated in a rift depository. Coeval (ca 3.5-3.3 Ga) volcanism in 'the Barberton and Pilbara greenstone terranes took place in an oceanic environment distant from any continental influence. Intercalated sediments (Association 2) are mainly of airfall pyroclastic and less commonly epiclastic origin. Reworking of clastic material occurred in subaerial, and subaqueous environments both above and below wave base. Associated evaporites and biogenic sediments, including stromatolites and rip-up algal mats, support a relatively shallow-water origin for the volcanism and sedimentation. The greenstone volcanics in both regions are overlain by terrigenous clastic sediments (Association 3) recording the onset of continental-margin sedimentation. In the Barberton Mountain Land sedimentation is constrained to 3.4-3.3 Ga with influx of extrabasinal detritus a response to crustal shortening and uplift of a 3.5 Ga gneiss terrane. The synorogenic deep- and shallow-water sediments are visualized as a flysch-molasse couplet. Sedimentary rocks in 3.0-2.5 Ga greenstone terranes (Association 4) in Canada and the Yilgarn Block are dominated by reworked-pyroclastic and terrigenous turbidites, iron-formations and massive sulfide deposits. Stromatolites and shallow-water sediments are rare to absent. The association of lithologies is interpreted to be mainly of deep-water origin. Cratonic sediments (Association 5) of similar age to the younger greenstone association are developed on the Kaapvaal Province and the southern Pilbara Block. Included in this association are rift-related volcanics and terrigenous and biochemical sediments, and epicratonic braided-alluvial and tidal-shelf terrigenous, biochemical and orthochemical sediments.

Rare earth element patterns in Archean high-grade metasediments

and their tectonic significance

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CONTENTS

Abstract
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Metasediments from two contrasting types of Archean high-grade terrains are interpreted as being derived from distinct tectonic settings. The Kapuskasing structural zone, Canada, represents the deep roots of a typical greenstone belt, whereas the Limpopo Province, southern Africa and Western Gneiss Terrain, Australia, mainly consist of shelf sediments deposited on a granitic basement and then buried to the depths required for granulite facies metamorphism. Upper amphibolite to granulite facies paragneisses from the Kapuskasing structural zone have REE patterns similar to those of greenstone belt sediments, except where partial melting has occurred, forming restites with Eu enrichment and melts with Eu depletion. Except in this latter instance, metamorphism has not affected the primary REE patterns. REE patterns in Archean upper amphibolite to granulite facies metasediments from the central Limpopo Province and Western Gneiss Terrain show wide differences, ranging from patterns which resemble those in post-Archean terrigenous sediments, to typical Archean sedimentary rock patterns. The diversity in REE patterns for these shallow shelf metasediments is interpreted as resulting from derivation from local provenances. Samples with "post-Archean" patterns, displaying Eu depletion, are interpreted as being derived from K-rich granitic plutons which were portions of small, stable early Archean terrains, precursors of the widespread late Archean-Proterozoic episode of major cratonic development.

REE AS INDICATORS OF CRUSTAL EVOLUTION

Rare earth elements (REE) are relatively insoluble elements, with short residence times in the ocean. Accordingly, they are transferred almost quantitatively from upper crustal sources to terrigenous sedimentary rocks. The observed uniformity of REE patterns in post-Archean terrigenous sedimentary rocks, in contrast to the diversity of source rock patterns, is interpreted to indicate that the processes of erosion, transportation and deposition provide thorough mixing and averaging of these differing source patterns. Thus the REE patterns preserved in the sedimentary record provide an overall average of the composition of the upper crustal rocks exposed to erosion.

Post-Archean sedimentary rocks

Post-Archean terrigenous sediments display a depletion in Eu, relative to the other REE, when normalized to chondritic REE abundances. Since mantle derived igneous rocks (the ultimate source of the continental crust) rarely display either relative depletion or enrichment in Eu and since no upper crustal reservoir enriched in Eu has been identified, the depletion of Eu in the upper crust has been ascribed to intracrustal processes, whereby K-rich felsic igneous rocks, depleted in Eu, are produced. Petrological and isotopic evidence favour intracrustal melting, rather than fractional crystallization, to produce large volumes of K-rich granitic rocks (e.g. Wyllie, 1977; Allegre and Ben Othman, 1980). In this model, Eu^{2+} is retained in plagioclase, in the lower or middle region of the crust (Taylor and McLennan, 1985). Such reservoirs have not been positively identified. The ubiquitous depletion of Eu in upper crustal terrigenous sediments is thus considered due to the domination of the present composition of the upper crust by K-rich granites, granodiorites and other

felsic igneous rocks produced by intracrustal melting.

Archean sedimentary rocks and the significance of the Archean-Proterozoic boundary

Studies of REE patterns in low-grade Archean sedimentary rocks from Western Australia, South Africa, Canada, and Greenland reveal a significant difference from that observed in the post-Archean record (see recent review by McLennan and Taylor, 1984). There is more diversity in the patterns, which may be flat, MORB-like, or steeply LREE-enriched, typical of Na-rich plutonic rocks (tonalites, trondhjemites, with occasional Eu enrichment) and felsic volcanics. These patterns are attributed to local provenance effects. Mostly the patterns in the sedimentary rocks are intermediate, approximating to 1:1 mixtures of the extreme basic and felsic patterns. Significant Eu anomalies are absent. This evidence is consistent with an Archean upper crust dominated by the bimodal basalt : tonalite-trondhjemite-dacite igneous suite. Archean sedimentary rocks sampled are typically located in greenstone belts, but their petrography and chemistry indicate that they sample a wider provenance. For example, the metasediments at Kambalda, localized in a basic-ultrabasic volcanic sequence, contain a substantial Na-granitic component (Bavinton and Taylor, 1980). Since the presence of 10% or more of K-rich felsic igneous rocks with Eu depletion would be reflected in a corresponding depletion of Eu in the sediments, the general absence of such effects has been taken as evidence that K-rich granites and granodiorites were scarce (< 10% exposure) in the upper crustal rocks being sampled. These observations have led to a model of crustal growth in which the Archean upper continental crust is dominated by the bimodal basalt : tonalite-trondhjemite-dacite suite, mostly of mantle derivation, whereas

the post-Archean upper crust is dominated by K-rich granites, granodiorites and associated volcanic rocks, the products of intracrustal melting (Taylor and McLennan, 1985).

Major intracrustal melting events, preceded on time scales of 10^8 years by a massive addition of material from the mantle, are thus thought to occur over a period of 500-700 Ma towards the end of the Archean. This event is proposed to constitute the major episode in the growth of the continental crust, about 75% having been formed by about 2500 Ma ago, at the commencement of the Proterozoic Era.

Archean high-grade terrains

Archean sedimentary rocks found in high-grade metamorphic terrains have received little attention in formulating these models, partly because little geochemical work has been carried out in such regions. The limited REE data that have been reported from Greenland (Boak et al. 1982; Dymek et al. 1983; McLennan et al. 1984), Montana - Wyoming (Mueller et al. 1982; Gibbs et al., 1986) and India (Naqvi et al. 1983) indicate considerable variability, suggestive of the dominating influence of local provenance effects. There is growing evidence that many high-grade terrains contain metasedimentary rocks deposited in a different tectonic setting from those of the well studied greenstone belts.

In this paper, we study REE data for metasedimentary rocks from two contrasting types of Archean high-grade terrains. These are: 1) the Kapuskasing structural zone of Central Ontario, Canada, and 2) the Limpopo Province of southern Africa and Western Gneiss Terrain of Western Australia. The former region constitutes a high-grade equivalent of

typical late Archean greenstone belts, whereas the latter regions are examples of high-grade areas containing sedimentary rocks deposited in a cratonic environment.

GEOLOGICAL SETTINGS

Kapuskasing Structural Zone (KSZ), Canada

Rocks of the Kapuskasing structural zone (Fig. 1) represent highly metamorphosed equivalents of the widespread low-grade Archean granite-greenstone terrains. The KSZ transects the Superior Structural Province of central Ontario, extending from the eastern shores of Lake Superior northeast to the southern shores of James Bay. Along the southern portion in the Chapleau-Foleyet region, granulite facies gneisses are in fault contact with rocks of the Abitibi greenstone belt to the east and grade into lower grade rocks of the Michipicoten greenstone belt to the west, over a distance of about 120 km. This gradation from granulite through greenschist facies rocks is interpreted to represent an oblique cross section through about 20 km of Archean crust (Percival and Card, 1983). Detailed zircon geochronologic work by Percival and Krogh (1983) reveals the following tectono-thermal history prior to overthrusting: deposition of the volcanic-sedimentary sequence of the Wawa and Abitibi greenstone belts and intrusion of plutons - 2750-2696 Ma ago, metamorphism and tectonism 2700-2685 Ma ago, and post tectonic plutonism 2685-2668 Ma ago.

Pressures and temperatures of metamorphic equilibration increase gradually from west to east across the KSZ. Percival (1983) recognizes two high-grade metamorphic zones. (1) The garnet-clinopyroxene-plagioclase zone, characterizing the transition from amphibolite to granulite facies,

extends from the Ivanhoe Lake cataclastic zone in the east, westward into tonalitic rocks of the Wawa subprovince and (2) The orthopyroxene zone, which characterizes the granulite facies, occurs in four patches within the eastern and northern portion of the KSZ. Mineral pair thermobarometry yields temperatures > 800°C in the east, ranging to under 600°C in the west, with pressures ranging from 5.4 to 8.4 kbar (Percival, 1983).

The Limpopo Province, southern Africa

In contrast with the KSZ, the Limpopo Province (Fig. 1) forms an example of those Archean high-grade terrains which consist predominantly of metasedimentary rocks, and lack the extensive metavolcanic sequences typical of greenstone terrains. The Limpopo Province is bounded to the north and south by typical Archean granite-greenstone terrains. The metasedimentary rocks studied here, part of the Beit Bridge Complex, overlie Sand River gneisses (3786±61 Ma (Rb-Sr), Barton et al., 1983), and are intruded by the Messina layered intrusion (3270±105 Ma (Pb-Pb) Barton et al., 1983). The Beit Bridge supracrustals are thus early Archean in age. Metamorphic pressures and temperatures are estimated at about 10 kbars and over 800°C (Horrocks, 1983).

Lithological associations present in the Beit Bridge Complex have been studied by Eriksson and Kidd (1985, 1986). Quartzites up to 100 m thick are interpreted to be of detrital sedimentary origin, on account of high concentrations of heavy minerals (zircon, rutile), rather than as recrystallized cherts (Fripp, 1983). They are interlayered in places with amphibolites, interpreted as basic sills and dikes. Widespread biotite-garnet-cordierite-sillimanite gneisses resemble shales in composition and are interpreted as metapelites (Brandl 1983, Fripp 1983).

Associated magnetite quartzites (metacherts) represent metamorphosed iron formation. Units of marble and calc-silicate gneiss up to 30 m thick are interpreted as metasedimentary carbonates or calcareous-pelites. The protolith of the fourth association, the gray gneisses, is interpreted as arkose (Eriksson and Kidd, 1986). These authors have estimated the lithologic proportions of the primary sedimentary rock types in the Beit Bridge Complex as: mudstone: 50%; quartz arenite: 17%; arkose: 17% and limestone and calcareous mudstone: 17%.

Such a sequence is clearly different from the Archean greenstone belt terrains and represents a distinct early Archean tectonic setting. It resembles the widespread quartz arenite-carbonate association which characterizes shallow cratonic shelf environments in the Proterozoic and Phanerozoic (Eriksson and Kidd, 1986). The Beit Bridge supracrustals are thus considered to constitute an Archean precursor of later widespread cratonic shelves.

The Western Gneiss Terrain (WGT), Western Australia

The Western Gneiss Terrain is situated along the westernmost margin of the Archean Yilgarn Block, Western Australia. This terrain is composed mainly of banded, locally migmatitic, granitic (s.l.) gneiss with interlayered metasedimentary and mafic lithologies (Gee et al., 1981). Tonalitic rocks, the dominant rock type in many Archean terrains, are only minor components in this region (Gee et al., 1981). Metamorphic grade is generally upper amphibolite, but granulite facies assemblages are locally developed (e.g., orthopyroxene-garnet assemblages in some metapelites).

The samples investigated in this study come from the northern portion

of the Western Gneiss Terrain, in the vicinity of Mount Narryer (Fig. 1). Major rock types in this region are old (>3380 Ma) adamellitic and syenogranitic gneisses, fragments of a disrupted layered mafic intrusion, and younger metasediments (Myers and Williams, 1985; Kinny et al, 1986). Like the Limpopo metasediments, the lithologic assemblages in the Mt. Narryer region are characteristic of shallow shelf depositional environment: quartzite, conglomerate, pelite, carbonates and banded iron formation (Gee et al., 1981).

The depositional age of the metasediments is not precisely known, but zircon geochronology on the Mt. Narryer quartzites and metapelites constrains it to be between 3250 Ma (the age of the youngest detrital zircons) and ca. 2800 Ma (the age of metamorphic rims on zircons) (Kinny, 1986; Kinny et al., 1986). As sediments would be expected to contain a sample of detrital zircons from any exposed igneous rock, the large time interval between 3250 and 2800 Ma shown by zircon ages in the metasediments indicates either a lack of igneous activity during this period or deposition of the sediments soon after 3250 Ma. The latter hypothesis is considered more likely.

ANALYTICAL RESULTS

Sample selection

A description of samples and sample localities is given in the Appendix. It is always difficult to recognize with certainty, metapelitic lithologies in high-grade metamorphic terrains. Lithological associations, such as interlayered metaquartzites and carbonates, the presence of

graphite, high Al_2O_3 content or the occurrence of aluminous minerals such as garnet, sillimanite and cordierite and evidence for production of the rock through weathering processes from major element data (e.g., high Al_2O_3/Na_2O ratios) all may be used as evidence for a sedimentary protolith. None of these characteristics alone is diagnostic.

In the the Limpopo and WGT suites, sedimentary parentage is clear on the grounds of mineralogy, composition and lithologic association. A sedimentary origin for samples from the Kapuskasing structural zone is somewhat less obvious because of the lack of sedimentary structures or associations. The generally high Al_2O_3 content (up to 18.2%) and rare occurrence of graphite do, however, point to a sedimentary origin.

Two samples, KSZ-12 and KSZ-13 are from the same outcrop and, on the basis of field evidence, are clearly affected by partial melting.

Methods

Rare earth elements, Th, U and some other trace elements were determined by spark source mass spectrometry using the method described by Taylor and Gorton (1977). Other elements were determined by a combination of electron microprobe on fusion glasses, X-ray fluoresence spectrometry, atomic absorption spectrometry and inductively coupled argon plasma spectrometry. The details of which elements were determined by which methods are provided in the tables of analytical data (Tables 1-5).

Rare Earth Element data

The results are given in Table 1 for the six Limpopo and three Western Gneiss Terrain metasediments and five Kapuskasing paragneisses.

The Kapuskasing samples, where unmodified by partial melting, have REE patterns typical of low-grade Archean sediments (Figs. 2, 3). Two samples (KSZ-1 and KSZ-8) possess variably enriched LREE ($(La/Yb)_N$ = 26 and 5, respectively), with no Eu anomaly (Fig. 2). The two samples which have undergone partial melting (Fig. 3) have complementary REE patterns. The restite, KSZ-13, is LREE enriched ($(La/Yb)_N$ = 13) with a large positive Eu anomaly and the melt, KSZ-12, is extremely LREE enriched ($(La/Yb)_N$ = 49) with a slight negative Eu anomaly (Eu/Eu* = 0.93). If these two patterns are combined in 50:50 proportions, the resultant REE pattern closely matches that of unmelted paragneiss KSZ-1 (Fig. 2). Paragneiss sample KSZ-11 is LREE enriched ($(La/Yb)_N$ = 16) and has a large positive Eu anomaly. These features resemble those of KSZ-13, the paragneiss restite, suggesting that KSZ-11 may also be a restite.

The Limpopo metasediments provide an interesting contrast. These samples show a wide variation in REE patterns (Table 1, Fig. 4). Two samples have patterns which resemble those of post-Archean terrigenous sediments (PAAS, NASC, ES, Taylor and McLennan, 1985; Fig. 4a), with significant depletion in Eu (Eu/Eu* values are 0.56 and 0.65). A third sample (LP-30) has a peculiar V-shaped pattern (Fig. 5), which may be due to garnet concentration through metamorphic differentiation. This is discussed in more detail in the following section. The other three Limpopo metasediments display quite different patterns (Fig. 4b). Sample LP-20 has no Eu anomaly and a high La/Yb ratio, resembling the Archean Yellowknife and Pilbara sediments (Jenner et al. 1981; McLennan et al. 1983), suggesting it is derived from a source dominated by felsic igneous rocks (McLennan and Taylor, 1984). The REE pattern of sample LP-28 resembles the patterns of many low-grade Archean metasediments, with no Eu anomaly, but is much flatter, consistent with a preponderance of basaltic source

material (McLennan and Taylor, 1984). Sample LP-14, a ferruginous metachert, is generally similar to LP-28, although with very low total REE abundances due to dilution with quartz. This sample is probably part of an iron-formation sequence. REE patterns of Archean iron-formations may, in some cases, reflect hydrothermal inputs (Fryer et al., 1979) and accordingly are difficult to interpret. If crustal sources are being reflected in the REE data, the slightly steeper, more LREE-enriched pattern, compared to LP-28, would indicate a more felsic provenance, with no significant Eu-anomaly.

The three Western Gneiss Terrain samples have comparatively uniform HREE patterns (Fig. 6). All are characterized by LREE enrichment and substantial negative Eu anomalies with $Eu/Eu^* = 0.46-0.71$. HREE patterns are flat at about 6-9 times chondritic levels. There is variability in the degree of LREE enrichment with $(La/Sm)_N = 2.8-9.1$; the silica-rich samples (impure metaquartzites) having the greatest degree of LREE enrichment. These REE patterns, in general, resemble those in typical post-Archean sedimentary rocks.

Other data

The other geochemical data are presented in Tables 2-5. The major element data are particularly interesting (Table 2; Fig. 7). The Kapuskasing samples are all characterized by low Al_2O_3/Na_2O ratios and low K_2O/Na_2O ratios. Such data indicate a general absence of significant weathering and/or a Na-rich provenance. These characteristics are typical of the majority of Archean terrigenous sedimentary rocks (McLennan and Taylor, 1984). Apart from the metachert (LP-14), the Limpopo samples fall into two distinct groups. One group (LP-20, LP-28) has low Al_2O_3/Na_2O and

K_2O/Na_2O, similar to the Kapuskasing samples described above. These are also the samples with the typical Archean-like REE patterns. The other group (LP-4, LP-11, LP-30), with negative Eu anomalies, are characterized by high Al_2O_3/Na_2O ratios and K_2O/Na_2O ratios, indicative of fairly severe weathering in the source rocks (Nesbitt and Young, 1984). Samples from the Western Gneiss Terrain are similar to this latter group but with variable K_2O/Na_2O. The low CaO, particularly in the WGT samples, also indicate weathering processes affected the source rocks. The data are also consistent with a K-rich source composition.

Most of the samples analyzed have high Th/U ratios (>4) which are normally associated with post-Archean recycled upper crustal sources (e.g. Taylor and McLennan, 1985). However, because of mobility of Th and U during high grade metamorphism, and the possiblity of accessory phase control on Th and U abundances in the WGT quartzites (see below), the Th/U ratios cannot be regarded as a primary feature.

A number of the Limpopo samples and the metapelite from the WGT have high Cr and Ni concentrations (Table 5), with Cr values up to 570 ppm and Ni values reaching 220 ppm. Although Archean shales commonly have elevated Cr and Ni concentrations related to a mafic component in their provenance, anomalously high Cr and Ni abundances, with associated high Cr/V (>4) and Ni/Co (>7) ratios have been noted for shales from the Fig Tree and Moodies Group from South Africa, as well as some other Archean sequences (Danchin, 1967; Taylor and McLennan, 1985). The Limpopo and WGT samples tend to have Cr/V and Ni/Co ratios which are typical of most shales and accordingly the elevated Cr and Ni abundances are not anomalous.

INTERPRETATION OF ANALYTICAL RESULTS

Metamorphic Effects

Whole-rock compositions can be affected by granulite facies metamorphism in two ways: 1) by loss of elements to a fluid phase and 2) by partial melting of the protolith. In general, K, Rb, Cs, U, and Th are most susceptible to depletion during granulite facies metamorphism; the REE are generally considered to be immobile (Heier, 1973; Tarney and Windley, 1977; Rudnick et al., 1985). Granulites mostly show U depletion, whereas Th, Rb and possibly K, may or may not be depleted (Rudnick et al., 1985). In addition, in situ partial melting will partition all elements between the melt and restite portions of the migmatite, thereby changing the compositions from that of the protolith. However, unless the partial melt is removed from the vicinity, no net compositional change is created by partial melting.

Of the metasediments investigated here, only some of the KSZ samples have been modified by partial melting. Samples KSZ-12 and KSZ-13 represent the melt and restite, respectively, of a partially melted paragneiss outcrop, with biotite remaining as a stable residual phase. When mixed together in 50:50 proportions, the resultant major and trace element composition closely matches that of unmelted paragneiss KSZ-1. Thus, partial melting has not resulted in a net compositional change on outcrop scale.

K/Rb ratios for the KSZ paragneisses are similar to those of unmetamorphosed sedimentary rocks (Table 3), suggesting that fractionation of K from Rb has not occurred during metamorphism. Simlarly, Cs contents

are high and K/Cs ratios are comparable to those of sedimentary rocks, suggesting Cs contents were not affected by metamorphism. The retention of K, Rb and Cs in the metasediments is consistent with the apparent stability of biotite during the metamorphism. In contrast with K, Rb and Cs contents, Th/U ratios are high in all but one sample (KSZ-11 Table 4), suggesting significant U depletion relative to Th. These samples have La/Th values ranging from 2.7 to 6.8. Most Archean sedimentary rocks have La/Th near 3.6 ± 0.4 (McLennan et al., 1980). Thus, little Th depletion can be accomodated in these samples. The one sample with low Th/U (KSZ-11) has a very high La/Th ratio (26, Table 4), suggesting Th depletion. The low Th/U ratio and absolute abundances may reflect retention of these elements in an accessory phase such as zircon.

Three of the Limpopo metasediments and the WGT metapelite have experienced U depletion with little or no change in Th (Th/U ratios >6 and La/Th ratios of 2.3-3.8). Of the remaining Limpopo and WGT samples, two have low Th/U (LP28 and LP14), with relatively high La/Th, suggestive of depletion of Th. The other samples (LP20, MNFN, MNK) have Th/U and La/Th similar to unmetamorphosed sedimentary rocks. Th and U for the two WGT quartz-rich metasediments (MNFN and MNK) are probably controlled by detrital accessory phases (see next section). It is not known whether LP20 has undergone depletion in Th and U or has retained its original LILE concentrations.

The Limpopo sample (LP30) with the enriched HREE is interpreted as follows: from La-Gd, the pattern resembles that of post-Archean sedimentary rocks, with a significant depletion in Eu. From Tb to Yb, the pattern rises steeply, enriched in the heavy REE, typical of garnet (zircon cannot account for this enrichement (see next section)). Although the rock

contains 1-2% garnet, approximately 10% garnet would have to have been added to the rock in order to explain the REE pattern. In addition, the garnet is unaltered and shows no signs of breakdown. Therefore, while the REE require some form of garnet concentration on a hand specimen scale, presumably through metamorphic differentiation, this is not supported by the petrography. The reasons for this paradox are unclear although earlier metamorphic events, not preserved in the present petrography, could account for this discrepancy. Nonetheless, excluding the HREE enrichment, the overall REE pattern of this sample suggests a similarity to that of post-Archean terrigenous sediments, with a signficant negative Eu anomaly.

The Role of Accessory Phases

Accessory phases can affect the concentrations of REE, Th and U in sediments if they are preferentially concentrated by sedimentary processes during deposition. This factor is particularly significant when considering coarser-grained clastic sediments such as sandstones and siltstones. A number of the samples investigated here are quartz-rich, suggestive of being sandstones originally. In addition, zircon, apatite and monazite are present in some of the samples (see Appendix), so a brief discussion of their possible influence on the REE patterns and Th and U concentrations is warranted.

Concentration of detrital zircon could potentially cause HREE enrichment in sediments. The amount of zircon present can be deduced from whole rock Zr concentrations (Table 4). The highest Zr content in our samples is 180 ppm; this corresponds to < 0.04% modal zircon, if all this Zr is in zircon. The HREE content of zircon is extremely variable (Yb = 30 to 4000 ppm (Gromet et al., 1984)). Therefore, detrital zircon could

potentially cause a chondrite-normalized increase of 0.05 to 6 in Yb, dependent upon the Yb content of the zircon. However, all terrigenous sediments will contain some amount of detrital zircon as an integral component (post Archean shales typically have 200 ppm Zr (Taylor and McLennan, 1985)). Therefore, a sediment would have to possess unusually high Zr contents in order for concentration of detrital zircons to influence the HREE content. This is not observed in any of the samples investigated here. It is worth noting that the only sample to show significant HREE enrichment (LP-30) has far too high a Yb value to be contributed by detrital zircon.

Apatite and monazite can also contribute significant amounts of U, Th and REE. Lack of P_2O_5 data prohibits estimates of the concentrations of these minerals in the samples, but thin section examination shows these phases are volumetrically less than 0.1%. The REE are trace components of apatite, but are major constituents of monazite. Consequently, the REE patterns of the two monazite-bearing WGT quartz-rich metasediments may be controlled by this phase. Microprobe analyses of LREE, Th and U contents of monazites in samples MNK and MNFN are given in Table 6. These data show that only 0.01% monazite in MNK and 0.03% monazite in MNFN are required to account for all the LREE and Th in these samples. Similarly, 0.03% monazite in MNFN could account for all the U in this sample, but at least 0.06% monazite would be required in MNK to account for the uranium content. An unrecognized U-rich accessory phase may supply this additional uranium. Thus, the LREE enrichment and high Th and U contents of quartz-rich samples MNK and MNFN, in comparison to sample MN45, are probably due to accumulation of detrital monazite.

It is apparent from the discussion in this section that REE abundances

in sandstones may be seriously affected by the presence of resistant accessory minerals which are concentrated by sedimentary processes. The potential dominance of such secondary effects in siltstones and sandstones indicate that these sedimentary lithologies may be less reliable indexes of broader crustal evolution than shales. For these reasons, we have mostly concentrated on using shales and similar fine-grained sedimentary rocks in our studies of the evolution of the continental crust (e.g. Taylor and McLennan, 1985).

Provenance

In the general absence of alteration of REE patterns due to metamorphism, we interpret the KSZ, Limpopo and WGT REE patterns as primary, reflecting those of the original sedimentary rocks (except for sample LP-30). The chemical characteristics of the KSZ samples are very similar to those of greenstone-belt sediments (McLennan and Taylor, 1984). The steep to moderate LREE-enrichment and low Al_2O_3/Na_2O and K_2O/Na_2O ratios of the KSZ samples point to the mixing of variable amounts of relatively unweathered tonalite-trondjhemite-dacite and mafic rocks to produce the sedimentary protoliths.

In contrast, the Limpopo samples are characterized by chemical variability. Archean sedimentary rocks display much more diversity in REE patterns than their post-Archean counterparts, consistent with less efficient mixing of Archean source lithologies (e.g., McLennan and Taylor, 1984). The Limpopo province represents an extreme example of this diversity, similar in many respects to the REE data from the early Archean high-grade metasediments from west Greenland (McLennan et al. 1984).

Three of the Limpopo samples (LP14, LP20, LP28) have patterns resembling those of low-grade Archean sedimentary rocks, with provenances ranging from basic to felsic, and without Eu anomalies. Such patterns recall the diversity encountered in the Akilia and Malene supracrustal successions (McLennan et al., 1984). In this context, the Limpopo metasediments do not differ from those of many other Archean terrains.

The other three samples (LP4, LP11, LP30), as well as the three Australian samples, in contrast, show REE patterns typical of those of the post-Archean terrigenous sequences (Taylor and McLennan, 1985). They must be derived, to a large degree, from weathering and erosion of K-rich granitic plutons. This is consistent with derivation from a small-scale cratonic source, analagous to the present-day upper crust. The major element data, with high K_2O/Na_2O ratios, is also consistent with such a source. The observed Eu depletion is thus inherited from a K-rich granitic source, which in itself was derived by partial melting within the crust. This granitic source area could not have been exposed to erosion on a large scale, for the sedimentary rocks bearing the signature of Eu depletion are not ubiquitous.

TECTONIC IMPLICATIONS

An important conclusion from the these data is that Archean high-grade terrains cannot be grouped together and considered to be a distinctive tectonic association differing from low-grade terrains. The Kapuskasing, and no doubt other high-grade terrains, simply represent highly metamorphosed equivalents of greenstone belts. REE data from metasedimentary rocks found in such regions can, accordingly, be expected to have the same characteristics as sediments in greenstone belts.

In the Limpopo Province and Western Gneiss Terrain, early Archean metasediments 3.6-3.2 Ga in age preserve evidence of deposition in shallow-shelf environments. It seems reasonable to suppose that they were deposited at the margins of a small craton (Gee et al., 1981; Eriksson and Kidd, 1986). The REE patterns indicate that a variety of sources were available, with input from basic and felsic sources, including K-rich granites with negative Eu anomalies. Isolated sedimentary basins or environments have preserved this evidence on a scale of hundreds of square kilometers. The presence of K-rich granitic rocks contributing Eu-depleted REE to sedimentary basins has the following implication. Such Eu-depleted source rocks must arise by intracrustal melting at depths not exceeding about 40 km (i.e., within the plagioclase stability field). However, available data suggests that such Eu depleted sediments are uncommon in the Archean record, so that cratonic development and inferred intracrustal melting must be very limited, relative to the widespread bimodal igneous activity in Archean terrains. The Na-rich tonalite - trondhjemite - dacite association, at one end of the bimodal suite, are derived through partial melting of mafic-ultramafic compositions from depths where garnet is stable (e.g. Barker et al., 1981).

It is thus necessary to postulate the following sequence of events for the development of the basement and cover rocks in the Limpopo Province and Western Gneiss Terrain:

(i)   Formation of thick crust (>20-30 km).

(ii) Intracrustal melting to produce Eu depleted felsic rocks.

(iii) Erosion and sedimentation producing pelitic rocks, with associated quartzites and carbonates, indicative of shallow shelf environments.

(iv) Subsequent burial and granulite - upper amphibolite facies metamorphism to produce metapelites and associated metasedimentary rocks.

(v) Uplift to give present exposures.

The evidence suggests that localized segments of the Archean crust underwent this sequence of events. It is curious that only sediments in high-grade terrains preserve this record and that the evidence of intracrustal melting is lacking in other Archean terrains. This observation has important tectonic implications. It implies separate development of greenstone belts and of cratonic environments preserved in high grade terrains. The Eu-depleted rocks, sources for the high-grade shelf sediments, were not contributing (<10%) debris to the sedimentary basins preserved in low grade terrains (greenstone belts). This appears to contrast with modern sedimentary environments; for example, modern deep sea turbidite sands from differing active tectonic settings (fore-arc, back-arc, continental arc) very commonly show trace element and isotopic evidence for derivation from both old recycled upper continental crust as well as young volcanic - plutonic sources (McLennan et al., 1985). Rare earth element data for Phanerozoic greywackes show a similar pattern (Bhatia, 1985). Only some sediments deposited in fore-arc basins of wholly oceanic island arcs (e.g. modern Marianas Basin; McLennan et al., 1985, Devonian Baldwin Formation; Nance and Taylor, 1977) are devoid of a signature of intra-crustal melting.

Recycling and the previous extent of high-grade terrains

There are no detailed compilations of the relative proportions of high-grade versus low-grade supracrustal successions, although such

information is critical for models of Archean crustal evolution. Examination of geological maps suggests that high-grade rocks may comprise as much as 50% of all Archean supracrustal exposures. Of these, some proportion is simply a highly metamorphosed equivalent of greenstone belts.

Preferential preservation of cratonic environments, preserved in high-grade terrains is also likely. It is now well established that intracrustal recycling processes are a dominant factor in the preservation and present-day distribution of crustal rocks and that recycling rates vary considerably with tectonic setting (Veizer, 1984; Veizer and Jansen, 1979, 1985). For example, the half-life of sediments preserved in platform settings is about 5 times that of sediments deposited in immature orogenic belts and > 10 times that of sediments deposited in active margin basins and deep sea fans (Veizer and Jansen, 1985). The result of these differing rates of recycling is that active margin and immature orogenic belt sediments have a greater potential for being lost from the geologic record than are sediments deposited on cratons.

Archean sedimentary rocks developed in diverse tectonic settings. High-grade supracrustal rocks from areas such as the Limpopo Province are considered to have been deposited in a stable platform environment or, possibly, a setting analogous to a modern passive margin (Eriksson and Kidd, 1986). The tectonic setting of greenstone belt sediments is a matter of considerable debate, but all workers agree that they were deposited in tectonically active basins; modern analogues are considered to include island arc, back arc and foredeep basins (Dimroth et al, 1982; Tarney et al., 1976; Jackson et al., 1986).

Regardless of the precise depositional models, it is clear that

greenstone belt environments would be recycled at a rate considerably in excess of cratonic environments preserved in high-grade terrains such as the Limpopo Province. It follows that these cratonic environments are preferentially preserved. If relative recycling rates from modern environments are adopted, the cratonic regions may be over-represented by a factor of at least 5. This would indicate that such rocks comprised no more than about 10% of the exposed Archean crust.

A model for Archean crustal growth

There is substantial evidence that much of the growth and stabilization of the continental crust occurred towards the end of the Archean (Taylor and McLennan, 1985). However, a major conclusion of the present work is that throughout the Archean, limited and isolated areas of crust underwent substantial intracrustal melting, formation of upper crustal felsic rocks depleted in Eu, erosion and deposition of sediments depleted in Eu. This process of growth and stabilization is directly comparable to the processes responsible for crustal evolution from the late Archean onwards. It seems that for most of these areas, subsequent burial and metamorphism led to the production of high-grade terrains. The reason why these environments are almost invariably preserved in high-grade terrains remains a fundamental question.

It is unlikely that these regions comprised more than about 10% of the Archean crust. What were the causes of these events? In the Archean, high heat flow led to the formation of many small plates (Bickle, 1978). Continental areas were also small (Taylor and McLennan, 1985). In some cases, tectonic processes, the nature of which are not well understood, led

to thickening (or heating) of the crust with concomitant intracrustal melting. Later, sediments deposited in associated basins were buried to 25-30 km, perhaps by mini-continental collisions with associated overthrusting and double thickening of crust. Because only high-grade terrains appear to preserve sediments derived from K-rich granitic sources, the tectonic processes responsible for the development of this type of Archean crust and the later metamorphism may be related. Such environments provide a fundamentally distinct tectonic environment from the greenstone belts, which may or may not undergo high-grade metamorphism. Probably crustal growth and stabilization occurred in a sequence of small step-like increments throughout the Archean, with a major pulse of continental growth in the late Archean.

> "Go to the banks of the
> great grey-green, greasy Limpopo River....
> and find out"

[Kipling 1902]

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Appendix

Sample locations and descriptions.

(A) Kapuskasing Structural Zone

KSZ-1 Paragneiss

From paragneiss unit near Ivanhoe Lake cataclastic zone; within orthopyroxene isograd (Percival, 1983). Plagioclase (An_{27}), garnet and orthopyroxene form porphyroblasts in a matrix of deformed quartz, plagioclase, aligned biotite, apatite + zircon. Orthopyroxene is deformed and partially altered to biotite.

KSZ-8 Paragneiss

From paragneiss unit near Ivanhoe Lake cataclastic zone; within cpx isograd (Percival, 1983). It consists of coarse-grained quartz, plagioclase (An_{28}), biotite, garnet, graphite, apatite and zircon. Biotite shows preferred orientation; graphite occurs as large plates aligned parallel to the foliation.

KSZ-11 Paragneiss

From a partially melted paragneiss boulder, on the western end of KSZ within cpx isograd (Percival, 1983). It consists of medium-grained quartz, plagioclase (An_{35}), biotite, garnet, apatite and zircon; biotite crystals are slightly aligned.

KSZ-12 and 13 Melted Paragneiss

From a single outcrop in the western portion of the KSZ within cpx isograd (Percival, 1983). Both samples consist of quartz, plagioclase (An_{26} in 12 and An_{29} in 13), biotite, apatite and zircon. KSZ-12 is coarse-grained with no preferred orientation of biotite. It contains a

large fragmented garnet which is Mn-rich (11 wt. %) and shows textural disequilibrium -- it is partially replaced by biotite and muscovite. KSZ-13 is fine-grained with aligned biotite. Compared with KSZ-12, it has higher proportions of biotite and apatite.

(B) Limpopo Province

LP-4 Metapelite

From Constantia farm, 2 km north of Tshipise. Contains porphyroblastic garnet (~5 mm), with aligned inclusions of quartz, occurring in a medium-grained, foliated matrix of quartz, plagioclase, cordierite, biotite, sillimanite, opaques and zircon. Retrogressive features include cordierite altered to pinite, sericitized plagioclase and minor chlorite replacing biotite.

LP-11 Iron-rich Metapelite

From Dover farm, southeast of Messina. Medium-grained, polygonal garnet, with rounded quartz inclusions is intergrown with quartz, K-feldspar, cordierite and opaques. Minor biotite, zircon and apatite. Feldspar partially sericitized.

LP-14 Ferruginous Metachert

From Randjesfontein farm, southeast of Messina. Layers of magnetite and orthoamphibole are surrounded by quartz, which varies from coarse- to fine-grained. Apatite occurs as an accessory phase. Probably is a chert-rich part of an iron-formation.

LP-20 Metapelite

From Tweedale farm, north of Alldays. Porphyroblastic plagioclase

occurs in foliated matrix of quartz, plagioclase and fibrous biotite. Garnet occurs as minor porphyroblasts with inclusions of biotite and quartz. Zircon and apatite occur as accessory phases. Plagioclase is heavily sericitized in places, and biotite is partially altered to chlorite.

LP-28 Metapelite

From Oom Stammetjie se Kop, east of Swartwater, which is west of Alldays. Coarse-grained garnet and quartz, with symplectically intergrown grunerite and plagioclase. Minor ilmenite and biotite. Plagioclase extensively sericitized.

LP-30 Metapelite

From Graat Reinet farm, north of Swartwater. Large plagioclase porphyroblasts are surrounded by foliated matrix of biotite, quartz, cordierite, plagioclase and opaques. Minor, small garnet porphyroblasts occur. Zircon and apatite are accessory phases. Plagioclase heavily sericitized in places; cordierite partially altered to pinite. Minor chlorite alteration developed around biotite.

(C) Western Gneiss Terrain Samples

MN45 Metapelite

From pelitic layer within Mt. Narryer sequence, northern Mt. Dugel section. Porphyroblastic garnet and orthopyroxene are set in a foliated matrix of cordierite, quartz and biotite. Orthopyroxenes have minor alteration rims of biotite and chlorite. Magnetite and zircon are present as accessory phases.

MNK Quartz-rich Metapelite

From Mt. Narryer sequence. Acicular, aligned. sillimanite is surrounded by irregular quartz and cordierite. Accessory phases, identified by electron microprobe, include zircon, rutile, monazite, xenotime, and a rare Sc silicate, thortveitite ($Sc_2Si_2O_7$). Cordierite variably altered to pinite.

MNFN Metapelite

From Mt. Narryer sequence. Augen of garnet (up to 1.5 mm) are surrounded by a foliated matrix of quartz, acicular sillimanite, cordierite, biotite, magnetite and ilmenite. Zircon and monazite are accessory phases. Cordierite partially altered to pinite.

Table 1   Rare earth element data, in ppm.

	Limpopo Belt						Kapuskasing Structural Zone					Western Gneiss Terrain		
	LP 4	LP 11	LP 30	LP 20	LP 28	LP 14	KSZ-1	KSZ-8	KSZ-11	KSZ-12	KSZ-13	MN 45	MN FN	MN K
La	34.6	29.2	39.8	24.6	14.0	0.68	41.9	16.0	13.0	59.1	19.1	9.05	41	17.3
Ce	65.0	49.7	80.7	50.6	34.1	1.56	73.1	32.5	25.2	113	40.0	20.5	76	-
Pr	6.85	4.89	8.81	4.92	5.16	0.19	9.15	3.85	3.06	11.6	4.37	2.37	7.12	4.36
Nd	26.3	17.4	35.7	20.5	25.9	0.88	29.7	15.8	12.2	40.8	17.2	9.78	22.4	16.6
Sm	5.99	4.16	7.60	3.99	7.38	0.26	5.45	3.57	2.35	6.20	2.76	2.02	2.85	3.15
Eu	1.23	0.77	1.40	1.07	2.75	0.09	1.38	1.26	0.94	1.39	1.07	0.46	0.43	0.39
Gd	5.51	4.26	7.09	3.15	8.29	-	3.25	4.27	1.78	3.38	1.68	2.07	2.00	2.15
Tb	0.92	0.75	1.31	0.42	1.38	-	-	0.80	0.28	0.50	0.29	0.44	0.38	0.40
Dy	5.73	4.86	10.8	2.32	8.68	0.26	2.51	5.19	1.61	2.68	1.72	2.97	2.61	2.67
Ho	1.14	0.90	2.97	0.42	1.80	-	0.57	1.07	0.30	0.48	0.35	0.70	0.58	0.59
Er	3.38	2.66	11.0	1.21	4.94	0.14	1.66	3.02	0.83	1.33	0.92	2.08	1.68	1.84
Yb	3.01	2.24	14.5	1.26	4.43	0.12	1.60	2.94	0.82	1.28	0.89	2.22	1.56	1.71
εREE	160	123	226	115	120	4.2	171	91	63	242	91	55	159	51
Eu/Eu*	0.65	0.56	0.58	0.95	1.07	1.0	1.0	0.99	1.41	0.93	1.52	0.71	0.53	0.46
LaN/YbN	7.8	8.8	1.9	13.2	2.1	3.9	17.7	3.7	10.7	33.0	14.5	2.8	17.8	6.8
Y	29	23	64	15.7	45	1.25	19.5	27	8.4	9.4	12.2	17	14	15

Method: Spark source mass spectrometry (SSMS).   (Taylor and Gorton, 1977)   - :no data

Table 2 Major element data, in weight percent

	Limpopo Belt						Kapuskasing Structural Zone					Western Gneiss Terrain		
	LP 4	LP 11	LP 30	LP 20	LP 28	LP 14	KSZ 1	KSZ 8	KSZ 11	KSZ 12	KSZ 13	MN 45	MN FN	MN K
SiO_2	58.6	57.5	63.0	69.1	56.5	84.5	63.1	62.0	67.3	68.3	59.8	56.0	80.9	84.9
TiO_2	0.72	0.37	0.71	0.42	1.60	0.003	0.60	0.79	0.40	0.45	0.68	0.61	0.13	0.03
Al_2O_3	19.6	15.0	17.0	15.7	14.0	0.06	17.5	18.2	16.5	15.8	17.4	21.6	12.6	11.6
FeO	10.0	19.3	7.3	4.6	14.4	13.8	5.83	5.50	3.30	4.05	6.87	9.10	5.35	1.20
MnO	0.14	0.39	0.18	0.090	0.34	0.069	-	0.15	-	-	-	0.09	0.18	0.15
MgO	6.26	3.74	5.40	2.87	5.74	0.55	3.72	4.05	2.16	2.57	3.94	10.4	0.58	2.10
CaO	1.05	0.68	1.9	1.70	4.4	0.05	3.59	3.60	4.14	2.18	2.78	0.012	0.10	0.017
Na_2O	0.55	0.23	1.16	2.79	1.49	0.05	3.27	3.14	3.50	3.29	3.36	0.41	0.10	0.15
K_2O	2.69	2.00	3.65	2.77	0.94	0.01	2.32	2.51	1.59	2.76	3.17	1.39	0.12	0.02
Σ	99.6	99.2	99.8	100.0	99.4	99.1	99.9	99.9	98.9	99.4	98.0	99.6	100.1	100.2

Methods: Electron microprobe on fusion glasses for Kaspuskasing samples; combination electron microprobe, ICP and AA for Limpopo and Western Gneiss samples, - :no data.

Table 3 Trace element data for large ion lithophile elements, in ppm.

	Limpopo Belt						Kapuskasing Structural Zone					Western Gneiss Terrain		
	LP 4	LP 11	LP 30	LP 20	LP 28	LP 14	KSZ 1	KSZ 8	KSZ 11	KSZ 12	KSZ 13	MN 45	MN FN	MN K
Cs	3.1	0.48	3.9	11.7	0.14	-	4.9	2.0	0.8	1.3	3.4	0.4	0.14	0.19
Rb	-	-	-	-	-	-	65	54	42	89	95	-	-	-
Ba	390	620	290	600	170	4.0	1055	806	445	793	1270	460	31	130
Sr	22	32	37	154	73	1.5	487	449	384	452	463	2.7	2.8	3.8
Pb	27	11	14	17	3.2	0.34	9.3	6.0	8.0	12.7	7.6	1.8	8.6	2.5
K/Rb	-	-	-	-	-	-	296	385	314	279	245	-	-	-

Methods: SSMS except Sr in Limpopo and Western Gneiss samples by ICP and Rb and Sr in Kapuskasing samples by XRF, - :no data.

Table 4 Trace element data for high valency cations, in ppm.

	Limpopo Belt						Kapuskasing Structural Zone					Western Gneiss Terrain		
	LP 4	LP 11	LP 30	LP 20	LP 28	LP 14	KSZ-1	KSZ-8	KSZ-11	KSZ-12	KSZ-13	MN 45	MN FN	MN K
U	1.37	1.45	2.32	1.62	1.91	0.4	0.73	0.36	0.31	0.83	0.32	0.49	4.67	1.90
Th	9.15	12.5	14.8	7.69	1.39	0.10	6.11	2.69	0.49	22.1	5.58	3.23	16.4	9.19
Th/U	6.7	8.6	6.4	4.7	0.73	2.4	8.4	7.5	1.6	27	17	6.6	3.5	4.8
Hf	2.38	4.55	4.95	4.0	3.22	-	3.60	4.12	4.84	3.81	3.34	4.20	2.34	3.03
Zr	87	180	183	-	90	1.3	129	148	158	120	121	107	62	85
Zr/Hf	36	39	37	-	30	-	36	36	33	32	36	25	26	28
Sn	2.0	1.6	9.5	3.9	8.0	0.3	2.7	2.2	2.0	3.1	3.2	1.7	2.1	2.4
Mo	2.0	0.74	1.7	0.5	1.3	1.0	1.7	1.6	1.2	1.5	2.3	0.7	2.5	1.5
Nb	11	9.0	39	11	12	0.58	7.5	7.9	6.6	8.3	12.4	7.5	6.2	8.0
La/Th	3.78	2.34	2.69	15.2	10.1	6.80	6.86	5.95	26.5	2.67	3.42	2.80	2.50	1.89
Th/Sc	0.29	0.96	0.64	0.64	0.036	0.5	-	-	-	-	-	0.09	3.0	1.2

Methods: Spark source mass spectrometry, except Zr and Y by XRF in KSZ-8 and KSZ-11.   - :no data

Table 5 Trace element data for ferromagnesian trace elements, in ppm.

	Limpopo Belt						Kapuskasing Structural Zone					Western Gneiss Terrain		
	LP 4	LP 11	LP 30	LP 20	LP 28	LP 14	KSZ 1	KSZ 8	KSZ 11	KSZ 12	KSZ 13	MN 45	MN FN	MN K
Cr	517	255	260	230	120	2.2	172	–	–	142	269	570	68	26
V	225	85	148	98	250	4.6	130	–	–	85	145	260	27	12
Cr/V	2.3	3.0	1.8	2.3	0.48	0.48	1.3	–	–	1.7	1.9	2.2	2.5	2.2
Sc	32	13	23	12	39	0.2	–	–	–	–	–	35	5.5	7.8
Ni	220	43	110	120	29	5.5	69	–	–	73	79	213	31	12
Co	50	18	26	19	35	0.6	–	–	–	–	–	60	12	12
Ni/Co	4.4	2.4	4.2	6.3	0.8	9.2	–	–	–	–	–	–	–	3.6
Cu	100	13	44	7.0	1.4	8.0	47	–	–	40	21	–	11	–
Mn	1080	3020	139	700	2630	535	–	–	–	–	–	710	1360	1160
Zn	–	–	–	–	–	–	88	–	–	86	122	–	–	–
Li	6	–	42	35	10	–	–	–	–	–	–	–	–	–

Methods: ICP for Limpopo samples and Western Gneiss; XRF for Kapuskasing samples, – :no data

TABLE 6. LREE, Th and U contents of Monazites, wt. percent.

	MNK		MNFN	
	Range	Average n = 5	Range	Average n = 8
La	12.3-13.2	12.7	13.2-16.4	14.0
Ce	25.3-26.7	26.1	24.2-31.2	26.0
Nd	7.97-8.48	8.23	7.07-9.08	7.70
Sm	1.51-1.67	1.60	1.65-2.16	1.76
Th	5.62-6.85	6.32	1.78-8.00	5.13
U	0.14-0.37	0.29	1.39-1.96	1.62

FIGURE LEGENDS

Fig.1.   Generalized geological maps and sample localities for the Limpopo
          Province (top left) (after Watkeys, 1981), the Kapuskasing
          structural zone (bottom left) (after Percival, 1983) and the Mount
          Narryer region of the Western Gneiss Terrain (right)(after
          Williams et al., 1983).

Fig.2.   Chondrite-normalized REE plots for metasedimentary rocks from the
          Kapuskasing structural zone. Note that the patterns have variable
          La/Yb ratios and lack negative Eu-anomalies. Sample KSZ11, with
          the positive Eu-anomaly, may have had a small partial melt
          fraction removed. These patterns are similar to those found in
          sedimentary rocks deposited in most Archean greenstone belts.

Fig.3.   Chondrite-normalized REE plots of melt and restite phases of a
          single outcrop of paragneiss from the Kapuskasing Structural Zone.
          Also shown is a REE pattern derived from 1/1 mixing of the restite
          and melt phases. This pattern is similar to those shown in Fig.2
          and indicates there probably has not been large scale removal of
          any melt phase on an outcrop scale.

Fig.4.   Chondrite-normalized REE plots of metasedimentary rocks from the
          Limpopo Province.   a).   Samples showing significant negative
          Eu-anomalies. These REE patterns are typical of post-Archean
          sedimentary rocks and differ from typical Archean sedimentary
          rocks found in greenstone belts. Such patterns indicate that in
          some regions, Archean crust underwent intra-crustal melting to
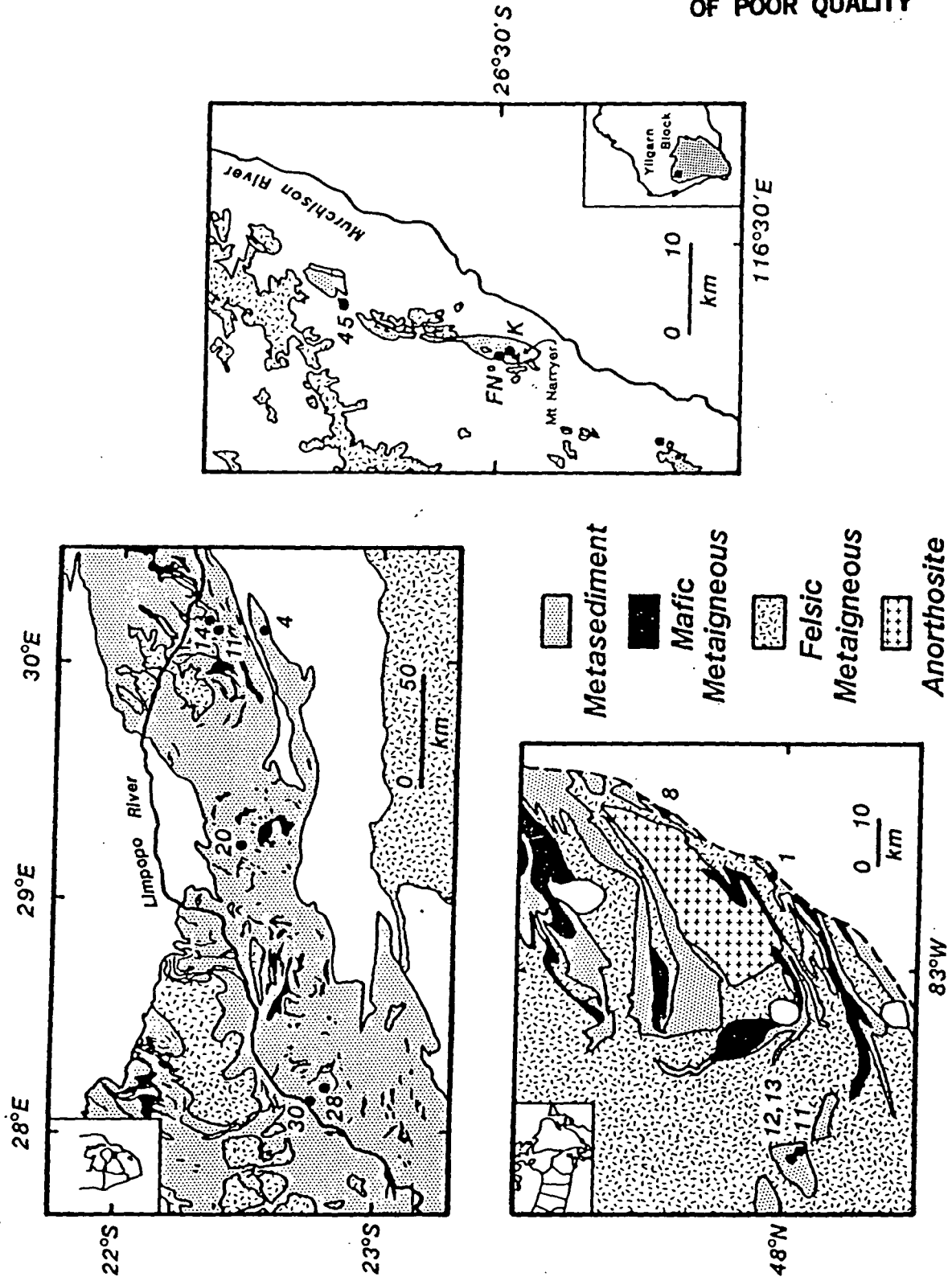   produce K-rich granitic rocks, which were then incorporated into

terrigenous sediments. b). Samples showing variable La/Yb ratios and lacking negative Eu-anomalies. Sample LP-14 is a ferruginous metachert and is probably from an iron-formation. These data indicate that even in the high-grade terrains which have sedimentary rocks possessing post-Archean type REE patterns, other sedimentary rocks with REE patterns typical of Archean greenstone belts are also common. Overall, these data suggest that the K-rich cratonic sources, with negative Eu-anomalies, were a local rather than regional feature.

Fig.5.   Chondrite-normalized REE plot of a Limpopo metapelite showing a negative Eu-anomaly and severe HREE-enrichment. The HREE-enrichment is best explained by addition of garnet during metamorphic differentiation, however the small amount of garnet present in the sample is insufficient to explain the magnitude of the enrichment.   A complex metamorphic history may be indicated. In any case, we interpret the negative Eu-anomaly as being a primary feature and accordingly include this sample with the others showing negative Eu-anomalies (see Fig.4a).

Fig.6.   Chondrite-normalized REE plot of metasedimentary rocks from the Mount Narryer region of the Western Gneiss Terrain. All samples show LREE enrichment, flat HREE patterns and negative Eu-anomalies.   The variable enrichment of LREE may be related to concentrations of the heavy mineral monazite in the quartz-rich samples.   Such REE patterns are consistent with a K-rich cratonic provenance.
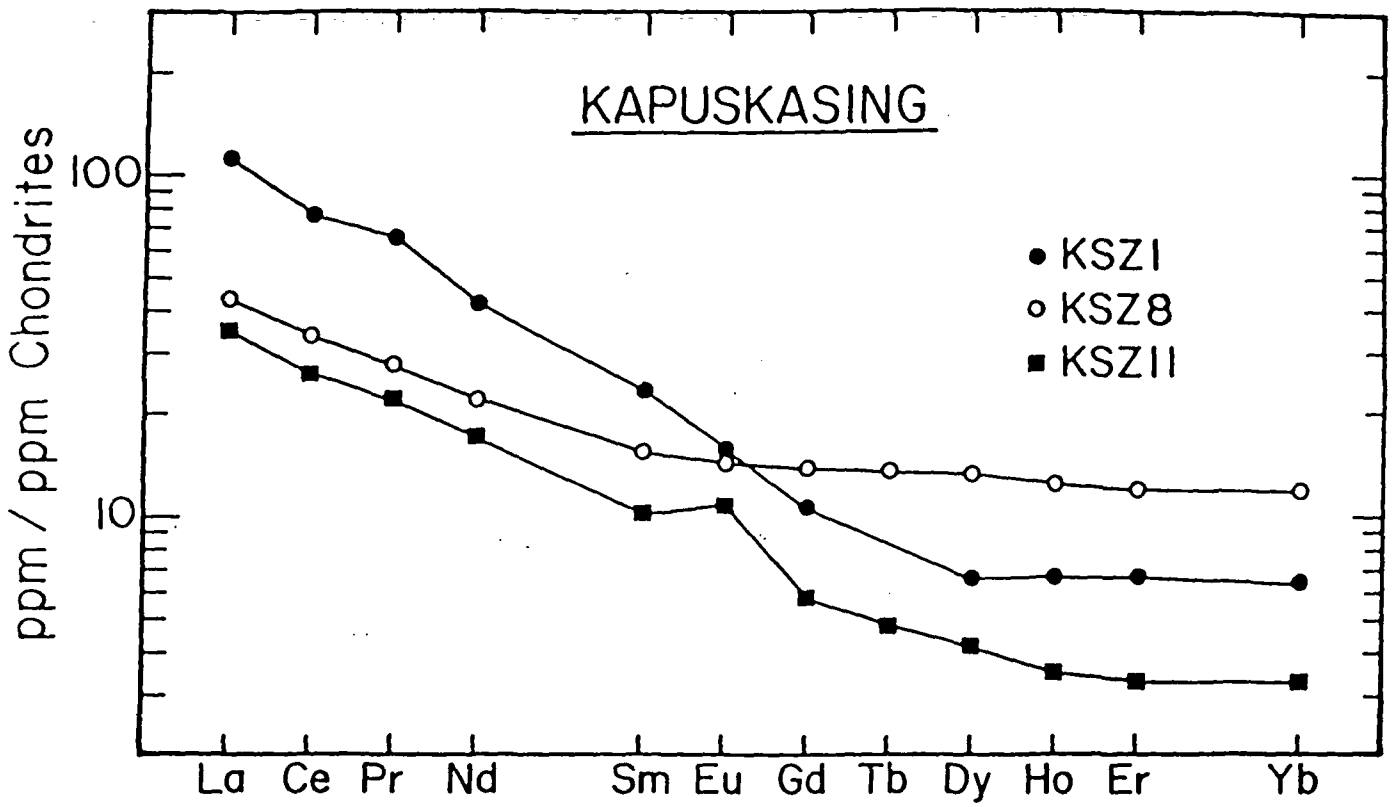
Fig.7. Plot of Al_2O_3 versus Na_2O for metasedimentary rocks from Archean high grade terrains. Samples with open symbols have negative Eu anomalies ($Eu/Eu^* < 0.9$) and samples with filled symbols do not ($Eu/Eu^* > 0.9$). Also shown are lines of constant Al_2O_3/Na_2O ratio, corresponding to average Archean and post-Archean shales (Taylor and McLennan, 1985) and Pleistocene Amazon Cone muds, derived from highly weathered sources (Kronberg et al., 1986). The heavy arrow indicates the general trend of increasing effects from weathering. Samples from the Kapuskasing region (lacking negative Eu-anomalies) show little or no signs of weathering, whereas WGT samples (with negative Eu-anomalies) are derived from severely weathered sources. The Limpopo samples show variable influences of weathering, which correlate with negative Eu anomalies.

Taylor et al. Fig. 1

Metasediment
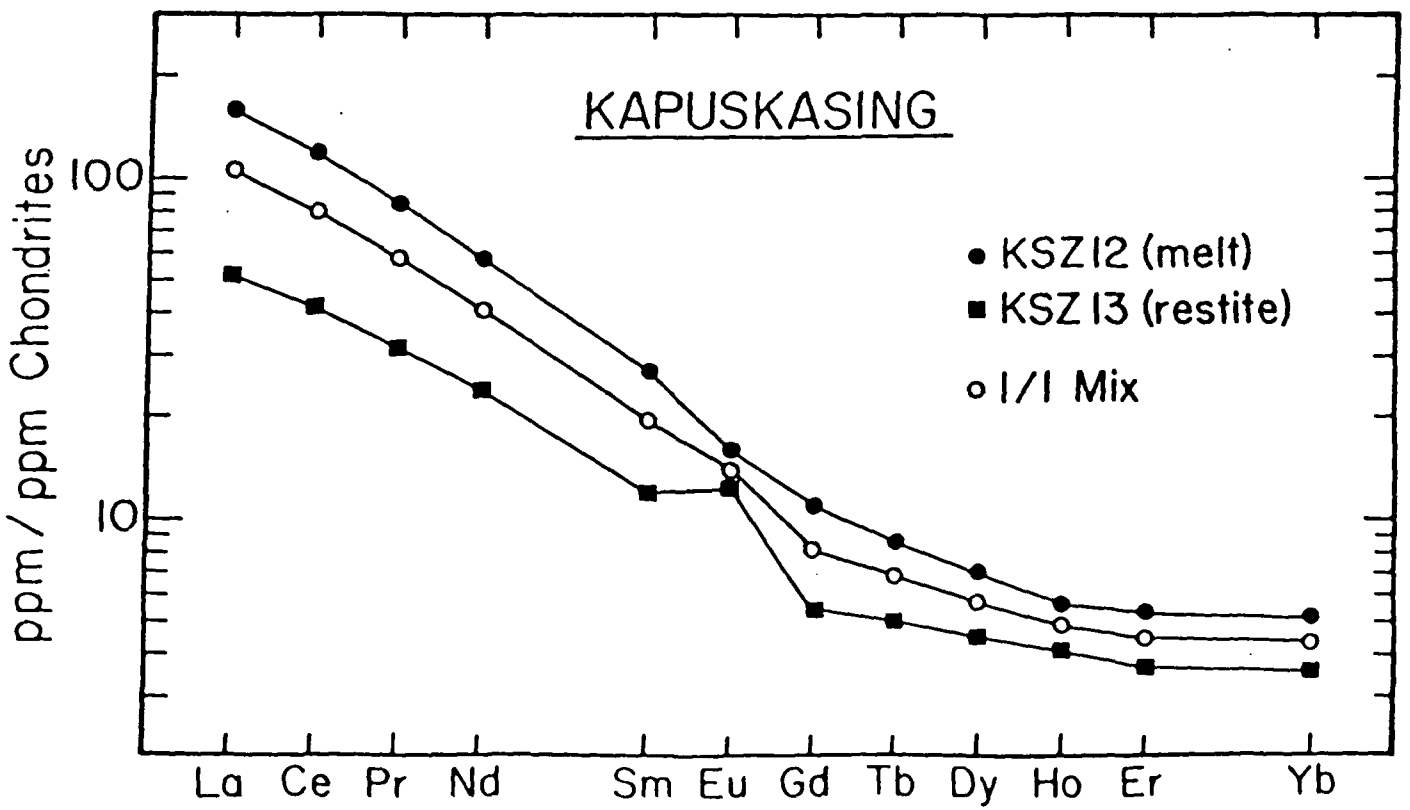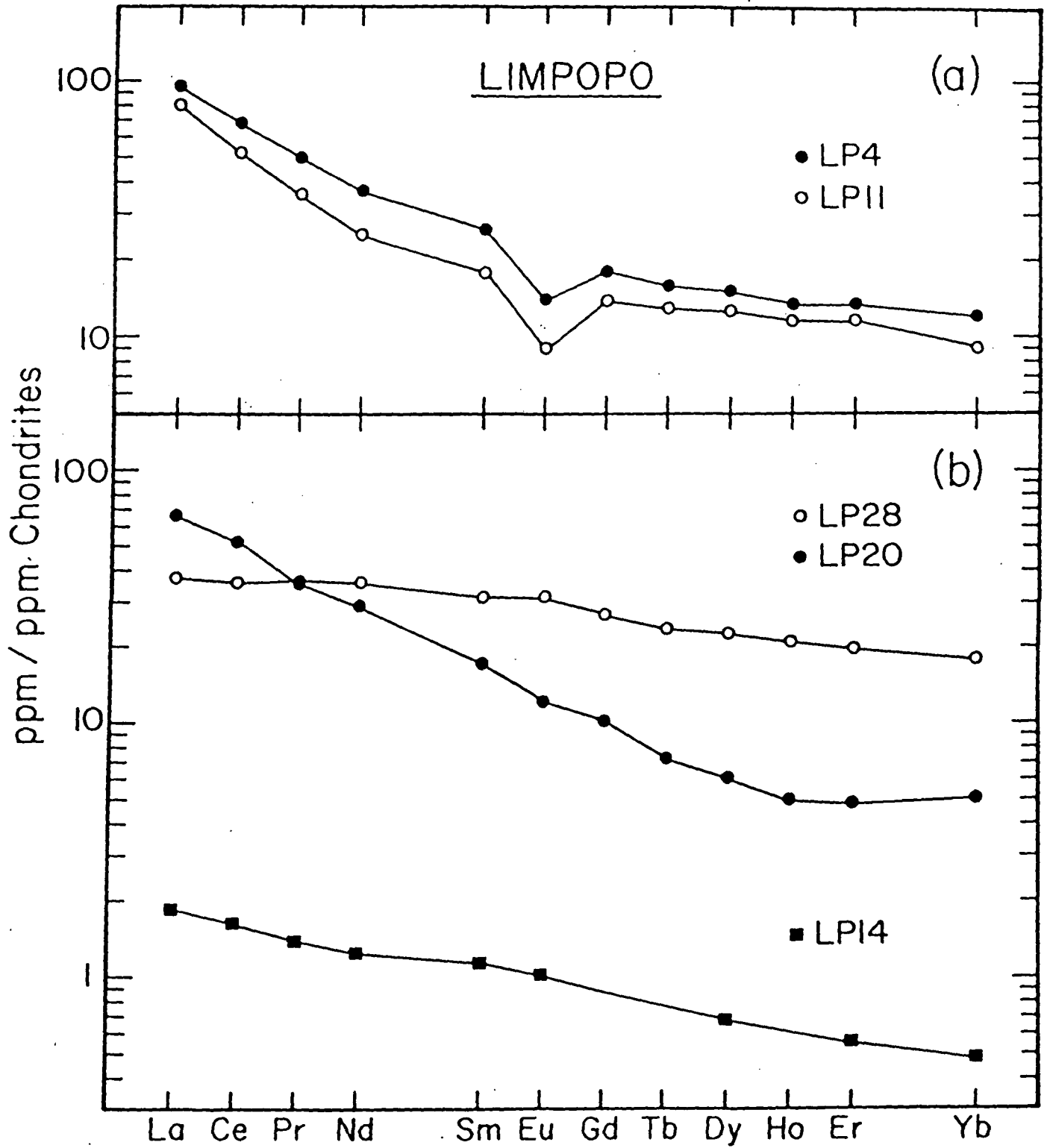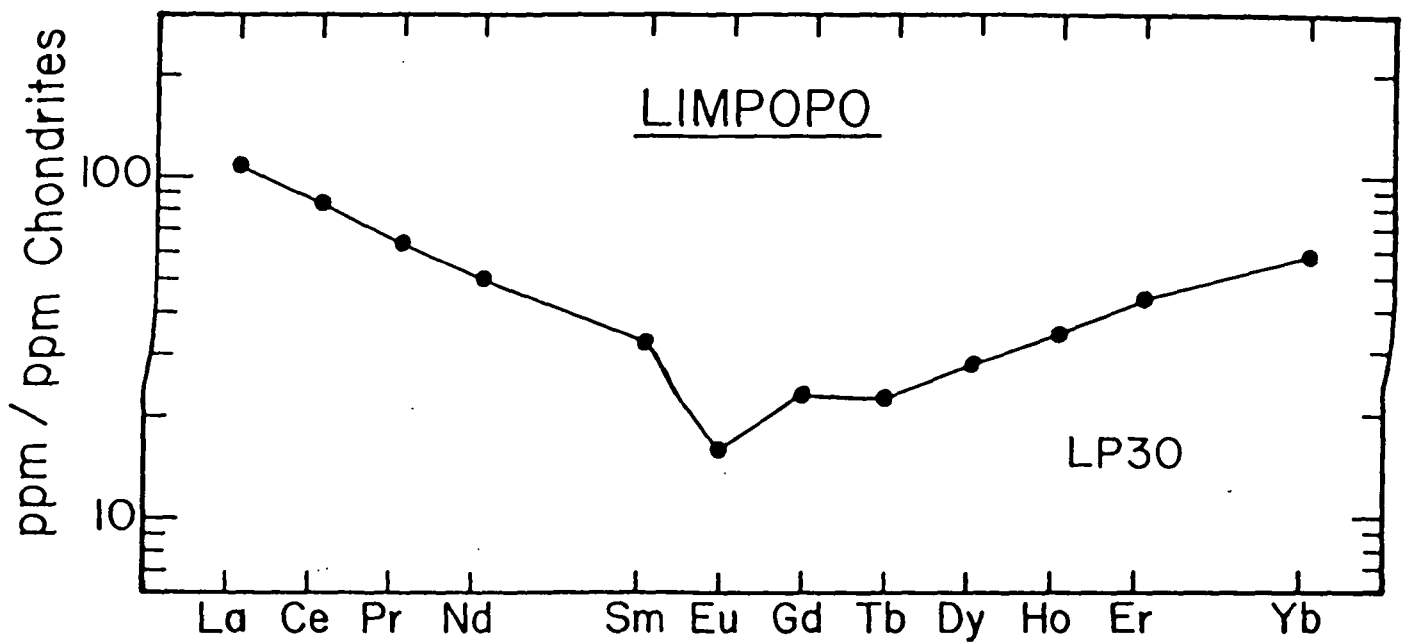
Mafic
Metaigneous

Felsic
Metaigneous

Anorthosite

KAPUSKASING

- KSZI
- KSZ8
- KSZII

Taylor et al. Fig. 2

KAPUSKASING

- KSZ12 (melt)
- KSZ13 (restite)
- 1/1 Mix

LIMPOPO

(a)

- LP4
○ LPII

(b)

○ LP28
● LP20

■ LP14

La Ce Pr Nd   Sm Eu Gd Tb Dy Ho Er   Yb

ppm / ppm Chondrites

Taylor et al. Fig. 4

LIMPOPO

LP30

Taylor et al. Fig. 5

WESTERN GNEISS TERRAIN
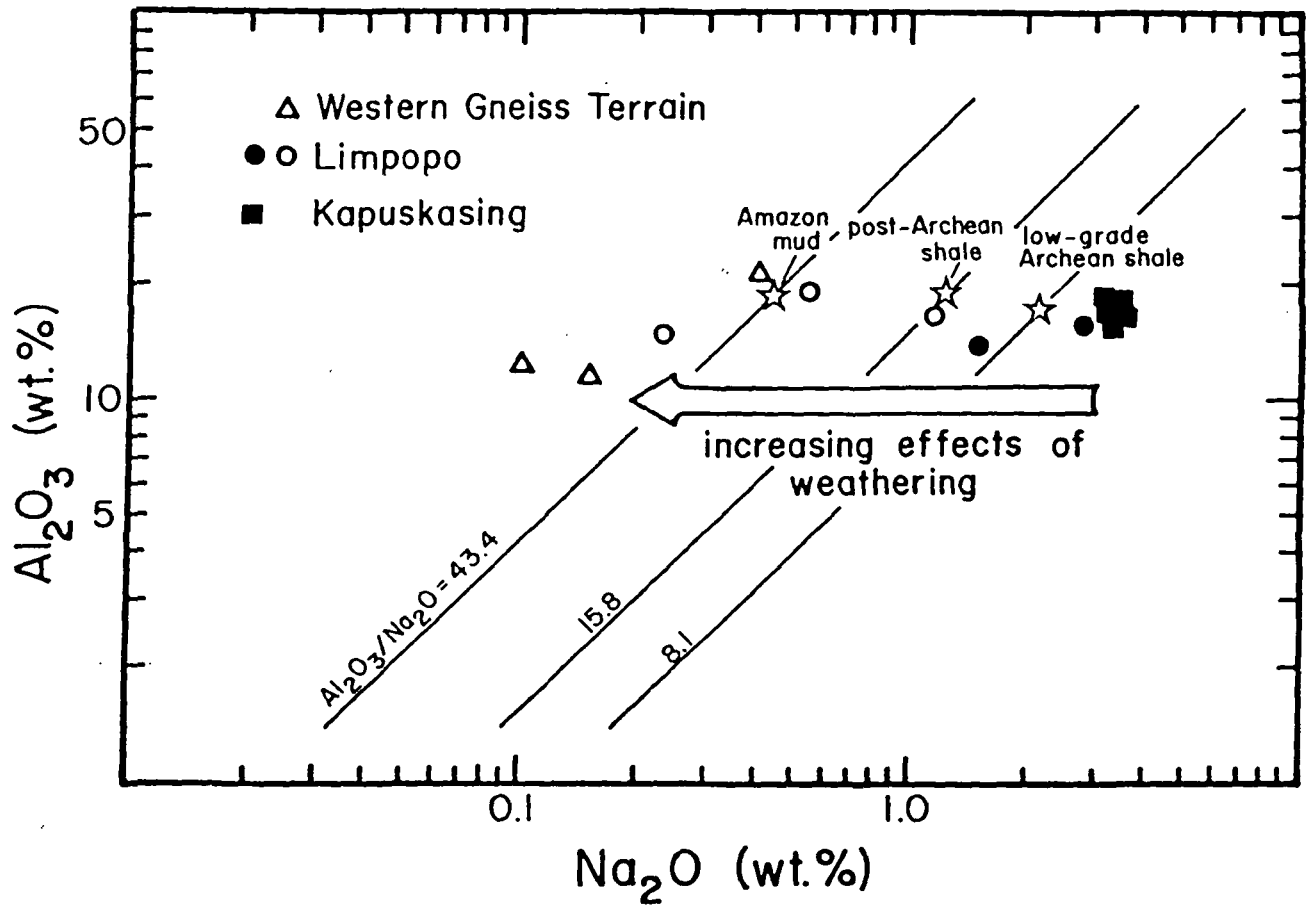
- MNFN
- MNK
- MN45
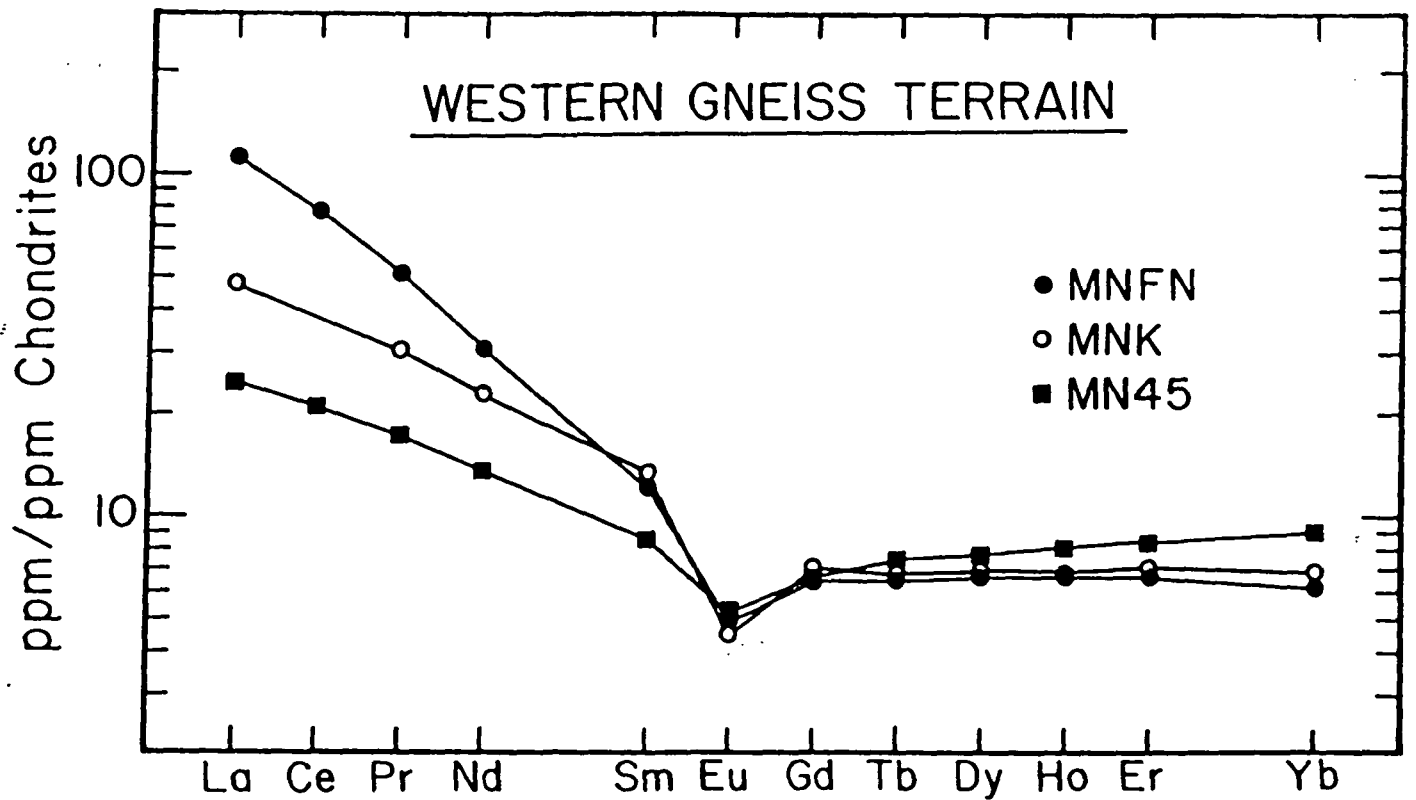
Taylor et al. Fig. 6

Taylor et al.   Fig. 7