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Archean fluvial deposits: a review

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Abstract

The recognition of Archean fluvial deposits is complicated in many cases by post depositional deformation and metamorphism. Most surviving deposits can be categorized as deposits of alluvial fans with and without debris flows, and sand and gravel-bed braided rivers. Examples of possible intermediate to high sinuosity meandering systems have been tentatively identified in Africa, Australia and India: all lack clear evidence of lateral migration and can be reinterpreted in terms of esturine deposition, shallow tide-influenced marine, and deep-water mass-flow deposits respectively. Mudstone intervals in Archean fluvial strata are rare, and where present, are typically of silt grade. These may represent ponds developed within channel thalwegs, or where more extensive may be of lacustrine rather than floodplain origin. Prior to 3.2 Ga preserved fluvial deposits appear to be largely confined to the flanks of volcanic cones or plateau, perhaps reflecting globally high sea level combined with the small-scale of cratonic nuclei. The onset of modern style plate tectonics in the early Mesoarchean allowed for more extensive generation and preservation of fluvial strata: most of these are first cycle deposits, preserved in rift, strike-slip, and foreland basins, with rare examples accumulating in forearc and syn-tectonic piggy-back basins.

1. Introduction

Documented examples of Archean fluvial deposits are known from Paleoarchean (3.6 – 3.2 Ga) and younger strata, becoming progressively more abundant and extensive towards the Archean/Proterozoic boundary at ~2.5 Ga. The increase in abundance is partly a function of increasing preservation potential, combined with secular changes in global sea level (Flament et al. 2013) and style and intensity of plate tectonic processes (Bradley 2011). Major changes in fluvial style have been strongly influenced by changes in the chemistry of the atmosphere, the Earth's surface temperature, and weathering regime (Long 2011; Bridgland et al. 2014). Eriksson et al. (2013) suggested that during the earliest Archean (Hadean, ca 4.5 – 4.567 Ga) the Earth's surface was largely molten, with initial terrestrial deposits dominated by talus, loess and aeolian strata. When surface temperatures dropped to below 100 °C at approximately 4 Ga (Lunine 2006; Eriksson et al. 2007), temporary accumulations of water would have permitted the development of tallus, fluvial- and debris flow-dominated alluvial fans, lakes and fan-deltas, as can be seen on the surface of Mars (Buhler et al. 2011; Williams et al. 2011) and Venus (Jones and Pickering 2003). All evidence for these earliest terrestrial facies on the Earth was destroyed by a prolonged period of heavy bombardment, ending at about 3.9 Ga (Eriksson et al. 2013).

Surface temperatures on post-Hadean Earth are widely considered to have been higher than during the Phanerozoic, based on oxygen and silicon isotope studies of chert. Sea temperatures were estimated to have fallen from ~75°C at 3.5 Ga to 64 to 41°C by 2.5 Ga (Robert and Chaussidon 2006; Kasting and Howard 2017), although this has been questioned based on climate modeling (Charnay et al. 2017) and dissolution rates of chert and quartz (Sleep and Hessler 2006; Brengman et al. 2016). Intervals when global mean surface temperatures fell below 20°C are indicated by intervals with glacial deposits at ~3.5 Ga (de Wit and Furnes 2016; Charnay et al. 2017), ~2.9 Ga (Young et al. 1998; Williams et al. 2016) and ~ 2.7 Ga (Page 1981; Ojakangas et al. 2014).

Through most of the Archean the atmosphere is generally considered to have been largely anoxic, with high concentrations of CO_2, and fluctuating concentrations of SO_2 and CH_4 (Kasting and Ono 2006; Nutman et al. 2017). Reducing conditions allowed preservation of detrital pyrite and uraninite in placer deposits (Rasmussen and Buick 1999; England et al. 2002 a; Young 2013; Hao et al. 2017; Burron et al. 2018). Despite this, there is mounting evidence for local microbially mediated oxidative weathering, even as early as 3.4 Ga in shallow water facies in the Strelley Pool Formation, in the Pilbara Craton of Western Australia (Wacey et al. 2014). Enhanced levels of CO_2 and SO_2 in the Archean atmosphere may have promoted aggressive weathering of labile material (Donaldson and Kemp 1998; Hessler and Lowe 2006; Hao et al. 2017). Greater production of fines, should have promoted mass-flow and hyper-concentrated flow processes in Archean fan and river systems (Corcoran and Muller 2002, 2004 a; Long 2011), but preservation of these fines in overbank settings would be limited by aeolian activity (Long 1978, 2011).

In Modern fluvial systems sediment character and supply is directly influenced by climate, relief, and the presence of rooted vegetation (Miall 1996; Blum and Törnqvist 2000; Ashmore and Church 2001; Schumm 2005). In the Archean, climate zones may have been strongly influenced by weaker solar radiation, a faster rate of the Earth's rotation, and marked differences in the tilt of the rotational axis (Williams et al. 2016).

This review will consider the known examples of Archean fluvial deposits, in terms of age, lithology, fluvial style, sedimentary architecture and tectonic setting. The main river types identified are summarized in Fig.1. Examples are listed by letter in the following text. Where fan-delta deposits have been identified only topset facies will be discussed.

<Fig 1 here>

2. The geological record

2.1. Eoarchean (3.8-3.6 Ga)

A 10 m thick, 1 km long unit of highly deformed framework- and matrix-supported meta-conglomerate in the >3.7 Ga Isua Greenstone belt was initially thought to be either a gravel-bed river deposit (Fig. 1, type B) or a shallow marine conglomerate (Fedo 1999, 2000; Fedo et al. 2001; de Witt and Furnes 2013), deposited in an intra-oceanic forearc setting (Komiya et al. 1999). Subsequent geochemical analysis of clasts and matrix demonstrated that these are not primary conglomerates, but are pseudo-conglomerates in which boudinized relicts of tonalite are set in a matrix of sheared, metasomatized tonalite (Fedo and Moorbath 2005). This indicates that in highly deformed strata field observations need to be supplemented by geochemical analysis to confirm interpretation. Polymict conglomerates elsewhere in the Isua belt are associated with meta-quartzites and iron formation, and have been interpreted as subaqueous, rather than subaerial, debris flow deposits (Komiya et al. 1999).

Other older migmatic quartzo-feldspathic components of the Ancaster gneiss (4.03-4.0 Ga: Bowring and Williams 1999) and granulites at Mt Sones in Antarctica (3.93

Ga: Black et al. 1986) contain rounded zircons that imply a terrestrial source area (Nutman et al. 2007), but contain no proven remnants of fluvial strata (Fig. 2).

<Fig 2 here>

2.2. Paleoarchean (3.6-3.2 Ga)

2.2.1 Australia

The lower part of the Paleoarchean succession in the Pilbara Craton of Western Australia contains at least four intervals where strata of possible fluvial origin have been identified (Fig. 3). In the lower part of the Warrawoona Group, there is evidence for thick lateritic soil development at the top of 5.9 kms of tholeiitic basalts, belonging to the 3.525 Ga Coontrunah subgroup (Buick et al 1995; Van Kranendonk 2000), implying prolonged subaerial exposure of a volcanic plateau (Van Kranendonk et al. 2006). The unconformity surface can be traced along strike for >75 km, with a relief of at least 10 m (Buick et al. 1995). Locally above this unconformity, clastics, cherts and (silicified) carbonates of the 3.481-3.477 Ga Dresser Formation appear to be confined to the interior of a subaerial volcanic caldera (Van Kranendonk 2010; Hickman and Van Kranendonk 2012). Initially these strata were considered to be of shallow marine origin, with local evidence of tidal activity (Noffke et al. 2013). Gypsiferous strata were interpreted as hot spring deposits formed at the sea floor (Nijman et al. 1998; Buick and Dunlop 1990). Subsequent analysis by Djokic et al. (2017) indicated that these hot spring deposits may have been subaerial. Possible evidence of fluvial activity in this setting was limited to isolated 16-18 cm thick layers of flat pebble conglomerate (geyserite [chert] rudite), and chert pebble conglomerate in units of 20 cm or less thick (Fig. 3, lower right). Given that these coarser clastics were deposited within an active caldera it is probable that the flat pebble conglomerates represent local reworking of chert or silicified carbonate strata (c.f. Buick and Dunlop 1990), by wave action or seismically driven tsunami. The pebble conglomerate, illustrated by Djokic et al. (2017 supplementary material), contains rounded chert clasts that are of local origin, possibly representing reworked botryoidal geyserite, and may also be of shallow marine origin. The conglomerates overlie chert and silicified mudstone beds of subaqueous origin and are only locally overlain by massive volcaniclastic sandstones.

Further up in the Warrawoona Group (Fig. 3) Barley (1993) recorded minor interbeds of framework-supported conglomerate, with angular to well rounded clasts, near the top of the Duffer Formation. These are interbedded with matrix-supported cobble conglomerate of possible debris flow origin. These may represent proximal valley confined straight gravel-bed river and fan deposits (Fig. 1, type A) formed on the flanks of a locally emergent volcanic edifice, similar to recent examples documented by Davies et al. (1978) and Selles et al. (2013).

<Fig 3 here>

At the top of the Warrawoona Group, trough cross-bedded volcaniclastic sandstones in the lower part of the 3.458-3.426 Ga Panorama Formation (Fig. 3, centre right), have been interpreted as fluvial deposits that were developed on the flanks of an emergent volcano (Retallack 2018). Retallack (2018) suggested that the presence of nodules of barite in overlying and interbedded fine grained strata (including chert, banded iron formation and silicified carbonate) was indicative of soil development by acid sulphate weathering. An alternative hypothesis is that the barite may have been introduced by later pervasive hydrothermal alteration (Brown et al. 2011). Elsewhere the formation has been described as predominantly marine (Lowe 1983; DiMarco and Lowe 1989; Buick and Dunlop 1990; Van Kranendonk et al. 2007). Detailed architectural studies are required to confirm or reject the fluvial origin of these strata (c.f. Long 2011).

The Panorama Formation is unconformably overlain by basal clastic strata of the 3.43 Ga Strelley Pool Formation (Fig. 3 upper right) that Van Kranendonk (2006) initially interpreted as fluvial in origin (Fig. 1, B). Subsequent detailed analysis of the basal chert-bearing pebble conglomerate (volcanic-rudite) and matrix-supported conglomeratic sandstone (volcanic-arenite) indicates that these are of coastal marine origin (beach) origin, and are intimately associated with wave-cut platforms, pot-holes, and local cliff faces (Allwood et al. 2006; Van Kranendonk 2010; Wacey 2010; Duda et al. 2016). These clastic strata are overlain by a series of shallow water stromatolitic carbonates and cherts (Wacey et al. 2014; Allwood et al. 2006; Hickman 2008) that locally contain channelized sandstone units. These were interpreted as minor fluvial channels by Van Kranendonk (2011: his Fig. 4), but given the small-scale and location within a marine interval, are more likely to represent nearshore marine gutter casts (c.f. Whitaker 1973; Myrow 1992). These shallow marine strata are overlain by graded volcanic conglomerates (agglomerates) and sandstones. These upper clastic strata (Member 5 of Lowe 1983) were initially interpreted as debris flow deposits (Fig 1, A), of distal alluvial fan origin by Van Kranendonk et al. (2001) and Van Kranendonk (2007 a, 2010). The detailed outcrop map of the Strelley Pool carbonates and cherts at the Trendall locality (Van Kranendonk 2007 a) shows the base of these upper conglomerates to have a local relief in excess of 6 m (Fig. 3, upper right). Boulder to pebble grade conglomerates are laterally discontinuous, and are intimately interbedded with laminated stromatolitic chert, chert breccia, and volcaniclastic sandstone. As the trough cross-stratified sandstones are also inter-collated with chert (possibly replacing original carbonate: Duda et al. 2016; Cammack et al. 2018), these clastic strata may also be of near-shore shallow marine origin, with the conglomerates representing either beach deposits (c.f. Allwood et al. 2006), or subaqueous facies of a prograding fan-delta complex (c.f. Van Kranendonk 2011).

In the Western Gneiss Terrane of the northwestern Yilgarn Craton meta-conglomerates and sandstones at Mount Narryer have been interpreted as braided fluvial by Eriksson et al. (1988) and Williams and Myers (1987). This ~2.5 km thick sequence was deposited between 3.3 and 3.2 Ga, and is characterized by thick packages of (recrystallized) matrix-supported quartz pebble conglomerate of possible debris flow origin (Fig. 1, type A), interbedded with contact framework conglomerates, massive, planar and trough cross-stratified sandstones and pebbly sandstones of braided river origin (Fig. 1, types B, C, L). Mudstone units (3-3.5 m thick), and a 14.5 m thick amphibolite in the 209 m thick section measured by Williams and Myers (1987), may represent lake deposits. Sillimanite-bearing quartzite and paragneiss in the upper part of the sequence (Units D and E in Eriksson et al. 1988) may be of marine origin.

2.2.2. Africa

In the Kaapvaal Craton of southern Africa massive dacitic conglomerates and sandstones in the uppermost part (H6) of the 3.455-3.445 Ga Hooggenoeg Formation (Onverwatch Group, Barberton Greenstone belt; Fig. 4, insert map) were interpreted as products of deposition on alluvial fans and fan-deltas by Lowe and Byerley (1999) and Rouchon et al. (2009). This unit was later re-named the Noisy Complex by De Wit et al. (2011), who interpreted the 100 to 1000 m thick sequence of matrix-supported conglomerates, with both inverse and normal inverse grading, as products of fluvial deposition, associated with intra-oceanic obduction. The conglomerates are directly overlain by turbiditic strata, and are locally interbedded with laminated siltstone and chert. This led Grosch et al. (2011) to reinterpret them as subaqueous marine debris flows, rather than subaerial fan deposits. Later the diamictites were reinterpreted as glacial deposits by de Wit and Furnes (2016), and the overlying sandstones as shallow marine deposits. Minor locally channelized contact framework conglomerates between the diamictites and marine sandstones may be glacio-fluvial, or alternately subglacial outwash fan deposits.

Clastic intervals in the overlying 3.26 to 3.223 Ga (Heubeck et al. 2013; Drabon et al. 2017) Fig Tree Group in the southern part of the Barberton Greenstone belt (Fig 4. insert) have been identified as fan-delta deposits that accumulated in a small foreland basin (Lowe and Nocita 1999). In the 360 m section of the Mapepe Formation measured by Nocita and Lowe (1990), fluvial (delta topset) strata were found at two levels. The lower interval contains a basal unit of massive pebble to cobble conglomerate, overlain by 13 m of crudely stratified pebble conglomerate with lenticular inter-beds, of massive, planar laminated, and planar and trough cross-stratified (pebbly) medium to very coarse sandstone. The upper 62 m of the measured section is dominated by massive to crudely stratified clast-supported conglomerates of pebble to cobble grade. Nocita and Lowe (1990) suggest that these crudely laminate conglomerates may represent channelized sheetflood deposits. Massive conglomerates (67%) tended to be lenticular, occurring in units 0.5 to 1.5 m thick that are characterized by marked lateral variations in thickness and texture, while minor (30%) graded units appear to be channel fills, deposited by normal streamflow processes (type A in Fig. 1). Other conglomeratic strata within the formation are considered to be of submarine fan, subaqueous fan-delta front, and tidal channel origin (Eriksson 1980; Lowe and Byerly 1999; Lowe and Nochita 1999). Thin dacite-rudites at the top of the overlying Schoongezich Formation may also be of braided fluvial (Huebeck et al. 2013) or marine origin.

Further south on the eastern side of the Kaapvaal Craton, the ~3.4 Ga Nondweni Greenstone belt (for location see Fig. 4 insert map), includes a 180 m thick, coarsening upwards sequence of polymict matrix- and contact-framework pebble to cobble conglomerate that lies above a thick sequence of basaltic and komatiitic volcanics of the Witkop Formation. These conglomerates, interpreted as alluvial fan deposits by Wilson and Versfeld (1994), occur in sets 0.15 to 3 m thick, and are locally interbedded with thin sets of cross-stratified volcaniclastic sandstone, and laminated ash-fall beds. They are overlain by 8 m of ash-fall beds, with local stromatolites suggestive of a return to shallow marine conditions.

2.3. Mesoarchean (3.2-2.8 Ga)

The Mesoarchean marked a transition from predominantly plume-dominated lid tectonics to modern style plate tectonics, with a marked growth in the scale and number of cratonic nuclei (Van Kranendonk, 2007 b; Hickman and Van Kranendonk 2012, Satkoski et al. 2017 and references therein). It contains an extensive record of fluvial deposition, predominantly within alluvial fan, braidplain, and fan-delta settings. Fluvial strata have been recorded in Australia, Africa, Canada, India, and Siberia (Fig. 4).

<Fig. 4 here>

2.3.1. Australia

Following, or during the final stage of consolidation of the Pilbara Craton, minor accumulations of possible fluvial strata have been identified in the 3.066 – 3.015 Ga Gorge Creek Group (Bagas et al. 2004; Hickman 2016). Nijman et al. (1999) suggested that lenticular, matrix-supported conglomerates, and associated ferruginous sandstones in his "Clastic Systems II and IV" were deposited as "alluvial fan related" strata (Type A in Fig. 1) that filled small basins formed as compressive growth structures induced by regional thrusting and/or inter-batholith folding. The conglomerates figured by Nijman et al. (1999: their Fig. 4) appear to have formed as thin (< 1m) sheets of predominantly contact framework conglomerate, associated with graded and laminated sandstone. These may represent fan related sheetflood deposits, or could alternatively have been deposited subaqueously in front of a prograding fan-delta, as shallow marine strata (Hickman 2016), or turbidites (c.f. McLennan et al. 1983; Van Kranendonk and Collins 1998).

On the northern margin of the Yilgarn Craton of Western Australia (Narryer Terrane) conglomerates and sandstones of the 3.1-3.05 Ga Jack Hills Group appears to have been deposited when an amalgamated braidplain, fed by coalescing alluvial fans, prograded into an extensional or transtensional basin that may have later transformed into a passive margin (Spaggiari et al. 2007; Eriksson and Wild 2010; Wang and Wilde 2018). Detailed observations by Eriksson and Wild (2010) indicated that, in the sections studied, 36% of the strata could be interpreted as marine or pro-deltaic lacustrine. The remaining strata included two main fluvial facies associations: the first consisting predominantly of sheets of framework-supported pebble conglomerate with laminated medium to very coarse sandstone (variant of type B in Fig. 1), interpreted as deposits of proximal, arid region alluvial fans, dominated by upper flow-regime conditions. The second association was dominated by planar laminated sandstones, representing distal fan related, flash flood deposits (type P in Fig. 1). Minor matrix-supported pebbly sandstones were interpreted as deposits of hyper-concentrated flows. Mudstone was not conspicuous in either facies assemblage.

Post 3.015 Ga plate convergence of the East and Northwest Pilbara Terranes produced a number of impactogen related grabens with local evidence of strike-slip activity (Krapež 1996; Hickman 2016). Within the East Pilbara Craton this resulted in accumulation of a 3 km thick sequence of conglomerates and sandstone of (transverse) alluvial fan and (axial) braided stream origin in the 2.950-2.940 Ga Lalla Rookh Formation (Krapež 1984, 1996; Krapež and Barley 1987; Krapež and Furnell 1987; Van Kranendonk and Collins 1998). Krapež (1996) recognised as many as 16 distinct lithofacies in the Lalla Rookh basin. These included sheets of massive matrix-supported conglomerate (diamictite) generated by debris flows on tallus slopes and on transverse fan systems (type A). Sheets of massive clast-supported boulder to small pebble conglomerate (type B) and conglomeratic sandstone represent axial gravel-bed river deposits (types B, L). The trough cross-stratified sandstones that make up 62% of the basin fill are interpreted as lower flow-regime sandy fluvial deposits (type L). Planar laminated sandstones of upper flow-regime origin make up a minor component of the basin, and are typically found on the fringe of transverse alluvial fan systems (Krapež and Furnell 1987). Bedsets illustrated by Krapež (1996, his Fig 6) indicate the presence of both downstream and upstream accretionary elements, suggesting the presence of low sinuosity channels at least 4 m deep (Fig. 1, M). Interbedded mudstone and wavy and ripple laminated sandstone (4%) were interpreted as overbank deposits or lacustrine turbidites by Krapež (1996), but may represent playa facies. Thick mudstones (4%) probably represent lake deposits.

Further north on the Pilbara Craton possible fluvial strata have been indentified in the 2.930-2.905 Ga Mosquito Creek Formation (Eriksson et al. 1994), although most of the formation is dominated by turbidites (Bagas 2005; Farrell 2006; Bagas et al. 2001, 2008) Eriksson et al. (1994) identified a 225 m thick interval, along the northern margin of the basin, dominated (64%) by trough cross-stratified medium to coarse grained sandstone (litharenites), with local silt drapes (0.4%), overlain and interbedded with four coarsening upwards cycles characterised by framework-supported pebble and cobble conglomerates (32%) of possible fan-delta origin. Observations by Nijman et al. (2010) indicate that along the northern margin of the basin emergent parts of small, south facing fan-delta systems (his McPhee Creek conglomerate) were dominated by mass-flow processes (type A in Fig. 1). Along the southern margin of the basin, finer grained conglomerates and associated trough and planar laminated sandstones, arranged in 1.5 - 14 m thick fining upwards sequences, in the Dromedary Hills unit (Nijman et al. 2010) are interpreted as northwest facing distal fan deposits, formed in an arid setting (types B, C, and? L), overlain by proximal debris flow dominated fans (type A). Nijman et al. (2010) also recorded a 50 m thick sequence of distal delta plain fluvial facies from the centre of the basin, that reflected deposition in sheetflood dominated rivers on delta topsets (types O and P).

2.3.2. Africa

Southern Africa contains a prolonged record of Mesoarchean fluvial (and aeolian) strata extending from 3.1 to 2.8 Ga (Figs. 4, 5), reflecting stabilization and rifting of the Kaapvaal Craton between 3.2-3.0 Ga, (Dominion, Moodies and Nsuze groups), and subsequent collision with adjacent cratonic blocks, allowing fluvial accumulation (Witwatersrand Supergroup) in the foredeep of an extensive retroarc foreland basin (Catuneanu 2001; Minter 2006; Bumby et al. 2011; Eriksson et al. 2011), as well as in the associated back-bulge where it was superimposed on a rifted continental margin (Swaziland and Pongola Supergroups).

<Fig. 5 here>

Conglomerates at the base of the lower (Md1 or Cluthra Fm.) division of the 3.223-3.216 Ga Moodies Group, in the Swaziland Supergroup of the Barberton Greenstone belt (Fig. 5) are dominated by matrix-supported pebble and cobble conglomerates of probable debris flow origin (Eriksson 1978, 1980). These are locally interbedded with framework-supported (locally bouldery) large to very large pebble conglomerates, and minor interbedded planar laminated and trough cross-stratified sandstone, indicating deposition in proximal gravel-bed rivers with flashy discharge (Fig. 1, type A). Local relief of as much as 10 m at the base of some contact framework conglomerates suggests channel incision. Overlying the basal conglomerates are 5 to 30 m thick storeys of predominantly trough cross-stratified sandstone (Sublith arenite and lith arenite), separated by layers with a few scattered pebbles. The tops of these storeys may have local climbing ripples and planar laminated sandstone, as will as mud and silt laminae (with some mud flakes). These have been interpreted as sandy braided river deposits by Eriksson (1978) and Heubeck et al. (2013), and in places are overlain by up to 60 m of quartz arenite, with inverse graded laminae and large-scale trough cross-stratification that Simpson et al. (2012) considered to be of aeolian origin. Fluvial (fan-delta) strata were also recognized in the upper parts of the Moodies Group (Eriksson 1978; Hose 1990; Eriksson et al, 2006; Heubeck et al. 2013). These include a channelized "impure sandstone" facies that is best developed in the Bavinansorp Formation (Md4), that Eriksson (1978) interpreted as products of deposition by sandy meandering rivers (Fig. 1, type G). This facies is characterized by 2 to 6 m thick lenses (up to 100 m wide) of predominantly trough cross-stratified sandstone, with a thin basal pebble lag, that thins and fines upwards from a single pebble layer, overlain by coarse to medium sandstones. The top 20% of the channel fill units is characterized by finer grained ripple and planar laminated sandstone, capped by laminated siltstone, with thin layers of mud intraclasts (but no dessication cracks). These channels are cut into units of interbedded laminated sandstone and siltstone that Eriksson (1978) interpreted as possible floodplain deposits. The channel-fill sandstones lack lateral accretion surfaces and inclined hetrolitithic stratification, and have strong unimodal paleocurrent distributions, suggesting deposition in a distal sandy braided, rather than meandering facies. Alternatively they may represent deposition in an estuarine setting (Hose 1990; Eriksson et al. 2006), perhaps within an incised channel caused by falling base level (c.f. Karpeta and Els 1999).

Younger strata in the 2.25 km thick Dominion Group (3.086 – 3.074 Ga), were deposited in the interior of the Kaapvaal Craton beneath the Witwatersrand Basin (Fig. 5). These include a basal 15 – 50 m unit (Renosterspruit Formation) dominated by locally cross-stratified arkosic sandstone, with minor conglomerate and pebbly sandstone. This rests on an undulating surface with a well-developed paleosoil (Button and Tyler 1979; Grandstaff et al. 1986) that has been metamorphosed to greenshist facies (Jackson 1991). Watchorn (1980 a, b), and Tankard et al. (1982) considered these sandstones to have been deposited on a low-gradient sandy braidplain, possibly in a rift setting (Jackson 1991; Minter 2006). The basal, 1.2 m thick, laterally discontinuous, planar bedded, small cobble conglomerate may have been deposited in a valley confined shallow gravel-bed river system (Fig .1, B) that rests on a low gradient pediment. A second more continuous 22 cm thick (small pebble) conglomerate, about 10 to 20 m from the base of the formation may represent gravel accumulation on a long-lived pediment surface. This is overlain by a 2.5 cm thick clay-rich unit, with abundant heavy minerals (Rantzsch et al. 2011) that may represent an aeolian deflation lag that accumulated on a long-lived inter-fluvial area. Overlying sandstones are interbedded with andesitic lavas and tuffs. The apparent high clay (sericite) content of these sandstones is probably related to in situ decomposition of feldspars.

On the east side of the Kaapvaal Craton (Fig. 5) sandstones and conglomerates of the Mantonga and Mdelanga formations form the basal 5 to 830 m of the 2.980 - 2.966 Ga Nsuze Group (Pongola Supergroup). These have been interpreted as predominatly sandy braided river deposits (Watchorn and Armstrong 1980; Armstrong et al. 1982; Gronewald 1984; Cole 1994; Wilson et al. 2013; Wilson and Zeh 2018). Detailed descriptions of the Mantonga Formation in its type area by Cole (1994) indicated that a basal (arkosic) paleosoil was overlain by a thick sequence of interbedded matrix-supported conglomerates, diamictites, mudstones and well-bedded quartzose sandstones. Cole (1994) indicated that the well bedded sandstones occured in cycles, characterized by a basal unit of massive or planar bedded pebbly sandstone, overlain by planar laminated sandstones, overlain by medium grained sandstones with small-scale trough cross-stratification, and capped by thin mud drapes. Cole (1994) suggested these represented sandy braided rivers (Fig. 1, types L, M) similar to the South Saskatchewan River model of Cant and Walker (1978) or the Platt River (Smith 1970). An alternate interpretation is that these units are of shallow marine origin, with a basal transgressive pebble lag. The example of cross-stratified sands illustrated by Cole (1994, his Fig 4.7) shows a large-scale composite bedform of possible downstream accretionary origin (DA), overlain by low angle cross-stratification, then medium-scale (<30 cm) trough crossbeds. The rounding of the upper surfaces of the 2nd order foreset boundaries is indicative of reactivation surfaces produced by reversing currents, supporting a marine origin (c.f. Gronewald 1984). Diamictite units with the succession represent mass-flow deposits, derived from contemporaneous volcanics of the Pypklipberg Formation (c.f. Watchorn and Armstrong 1980; Cole 1994).

The 2.97–2.78 Ga Witwatersrand Supergroup was deposited in a foreland basin setting (Eriksson et al. 2011), superimposed on older rift-related strata of the Dominion Group (Fig. 5). The basin is well known for the abundance of placer gold in many of the conglomeratic horizons within the > 7 km of strata preserved within the basin (Tankard et al. 1982; Els 1998 a, b; Frimmel et al. 2005; Minter 2006; Guy et al. 2010, and papers therein). Proximal facies are confined to the western and northern margins of the basin, with fluvial systems prograding towards the centre of the basin, where open marine conditions prevailed. Names of many of the formations change across the basin where they may be restricted to individual goldfields. To avoid confusion, the stratigraphic nomenclature used below, and in figure 5, follows Frimmel et al. (2005), and Guy et al. (2010).

The earliest phase of the Witwatersrand Basin (West Rand Group) is predominantly marine (Hospital Hill subgroup), with a distinct tidal signature (Watchorn 1980 b; Eriksson et al. 1981), passing up-section into sandy fluvial strata with minor conglomerates in the Bonanza Formation, along the southwest margin of the basin (Guy et al. 2010; Fig. 5). Fluvial strata are more extensive in the overlying Government and Jepperstown subgroups (Minter 2006). The Government subgroup contains a series of prograding wedges of fluvial conglomerate and sandstone, interbedded with marine sandstone, mudstone and local iron formation (Guy et al. 2010; Smith et al. 2013). Sandstones and minor conglomerates in the Promise Formation have been interpreted as products of both deep distal gravelly, and sandy braided river systems (type C and M in Fig. 1) by Tainton and Meyer (1990), and distal sandy facies as Platt-type braided rivers (type L in Fig. 1) by Camden-Smith (1980). These pass downstream into distal facies, including rippled and planar bedded sandstones with minor mudstones that have been interpreted in part as floodplain deposits (Camden-Smith 1980), but are more likely coastal marine facies. Thick (3-50 m) diamictite units in this unit (and the overlying Coronation Formation: Cairncross and Brink 1990) were originally interpreted as debris flow deposits by Tainton and Meyer (1990), but have since been reinterpreted as glacial marine by Smith et al. (2013), and are considered equivalent to glacial strata in the Mozaan Group in the Pongola sequence (Beukes and Cairncross 1991; Von Brunn and Gold 1991; Young et al. 1998). Watchorn (1980 b) and Cairncross and Brink (1989) describe possible sandy fluvial strata in the Tusschenin Formation (=Witpoorjie Sandstone Fm.) that in other areas Camden-Smith (1980) had interpreted as coastal and shallow marine.

Watchorn (1981 a, b) and O'Brien (1991) suggested that massive and planar laminated sandy strata in the Elandslaagte Formation, at the top of the Government subgroup, may represent ephemeral stream deposits (Fig. 1, type P), although more proximal conglomeratic and planar and trough crossbedded facies encountered in the subsurface by Cairncross and Brink (1990) indicate deposition in shallow sandy and gravel-bed river systems (types B and L). Silty-mudrocks form 1.1 % of the section intersected in the borehole, occurring in a single 2 m thick unit that could represent a floodplain lake or a channel fill.

The base of the Jepperstown subroup lies disconformably on strata of the Government subgroup, with a relief of 40-45 m (Watchorn and O'Brien 1991). Basal sandstones and conglomerates of the lower Reitkuil Formation (also known as the Koedoslaagte and lower Florida formations: Watchorn 1980 a, 1981 b; Guy et al. 2010) occupy three distinct paleovalleys, up to 2 km wide. Conglomerates are typically thin (0.1-1 m) and occur in 10-15 m wide channels, surrounded by trough cross-stratified sandstones. These valley-fill sandy braided river deposits (type L) are overlain by a laterally extensive massive pebble conglomerate (up to 1.8 m thick) that represents marine reworking of the fluvial braidplain during transgression (Watchorn and O'Brien

1991). The upper part of the Reitkuil Formation (Babrosco and upper Florida formations) consists of a progradational wedge of coarse arkosic sandstones and minor granule and small pebble conglomerates of sandy braided stream origin (type L), that passes up section into tidally reworked sandstones (Watchorn and O'Brien 1991). The uppermost sandstone dominated unit within the Jepperstown Group, the Maraisburg Formation, may be of fluvial, fan-delta (Guy et al. 2010) or marine origin (Watchorn 1980 a, b), but has not been described in detail.

The Central Rand Group (2.89-2.78 Ga: Eriksson et al. 2005) lies unconformably on strata of the West Rand Group, and hosts most of the major gold bearing pyritic quartz pebble conglomerates in the Witwatersrand basin (Minter et al. 1988), separated by intervals of predominantly trough cross-stratified lithic or sublithic sandstone (Tucker et al. 2016). Strata are predominantly fluvial along the northern and western basin margins (Guy et al. 2010), with tidal and storm-influenced clastic deposition prevailing towards the basin centre, in proximity to the Vredefort structure (Minter et al. 1988; Karpeta and Els 1999). The overall basin architecture reflects repeated progradation of fans, fan-deltas or amalgamated braidplains (badja) from fixed riverine entry points in response to base level changes induced by tectonic loading and sea level fluctuation (Minter and Loen 1991; van Eeden 1996; Els 2000; Catuneanu and Biddulp 2001; Minter 2006; Bumby et al. 2011). Els (1998 a) argued that true alluvial fans may not be present, with fan-shaped distributary patterns reflecting selective entry points along the basin margins. In some cases, down-slope divergence of channels and paleocurrents may support the presence of distributary fluvial systems (Davidson et al. 2013) emanating from distinct basin entry points (c.f. Fig 21 in Tucker et al. 2016), although no progressive down-slope decrease in channel dimensions has been documented, except for locally in the B placer in the Welkom Goldfield (Minter 1978). Proximal to distal changes include progressive decrease in maximum grain size (in several units), associated with decreased abundance of gravel towards the basin centre (Els 1998 a; Tucker et al. 2016). Major placer deposits typically develop as thin sheets on erosional discontinuities (sequence boundaries), with thin pebble lags or gravel bar deposits (Gm, Gh), overlain by trough cross-stratified sandstones (St) (Minter et al. 1988; Els 1998, 2000). Upper surfaces of the thinner conglomerate sheets may be leveled off by wind action, or by marine reworking during transgressive phases (Minter 2006). Aeolian strata are typically not preserved, but aeolian activity has been identified in the form of dreikanters and gold toroids (Minter 1999). Although most of the Central Rand Group is poorly exposed and deeply buried, planiform architecture has been inferred from thickness trends of conglomerates, gold concentrations, paleocurrent vectors and facies interpretations (Fig. 6). Discrete 2 to 6 km wide channelized units (Fig 6, A B) indicate valley-incision (c.f. Dalrymple et al., 1994) during intervals of lowered sea level (Els 2000).

<Fig. 6 here>

The basal Johannesberg subgroup, of the West Rand Group, begins with the Blyvooruitzich Formation (Guy et al. 2010) that along the northern and western basin margins begins with a thin small pebble conglomerate (Aida May, Beisa, North reefs), overlain by sublitharenites and minor pebbly sandstones (Bailey 1991). These collectively reflect deposition in Donjeck and Platt type rivers (types B and L in Fig. 1).

The locally overlying Main Formation includes several extensive thin gravel sheets, including the Main, South, Carbon Leader, Commonage, Composite and

Middelvlei reefs (Hargraves 1961; Els 1987, 1991, 1998 a, b, 2000; Tucker et al, 2016). The Main reef, at the base of the formation in the Central Rand Goldfield has been described by Stewart et al. (2004) as an amalgamation of gravel-bed braided rivers (B) that rest on an erosional discontinuity, with distinct braid channels (Fig. 6 A), and is overlain by a thin sheet of marine sandstones, with local bimodal opposed cross-stratification, and thin layers of muddy siltstone of the "green bar" (Stewart et al. 2004; zMh in Fig. 7 B). A similar braided pattern, with channels up to 3 m deep, is evident in the East Rand (Vos 1975). In the Carletonville Goldfield (Tucker et al. 2016) the Main reef incorporates the Carbon Leader reef, with 2 – 5 km wide zones of conglomerate (up to 3 m thick, in ribbons up to 0.5 km wide), separated by broad areas of thin conglomerate or pebbly sandstone (Nami 1983; Buck and Minter 1985), that Tucker et al. (2016) described as "overbank" facies, and Nami (1983) interpreted as sheetflood deposits (Fig. 6 B). Nami (1983) suggested local deposition by Platt type rivers (L) in channels up to 8 m wide and 0.8 m deep. Where the conglomerate is exceptionally thin (< 10 cm) it is often associated with auriferous carbon seams. The origin of the carbon, in this and other reefs, was originally considered to be of algal or bacterial origin that accumulated in abandoned channels or in inter-fluvial areas (Halbauer 1975; Minter 1981). More recent analysis indicates that this may not be the case, and implies that hydrocarbon was introduced during burial and metamorphism of marine mudrocks in the basin centre (England et al. 2002 b; Fuchs et al. 2015, 2017).

<Fig. 7 here>

In the Carletonville (West Rand) Goldfield, the Carbon Leader, and a thin sequence of trough cross-stratified sandstones, are erosionally overlain by a series of broad channels, 2 – 3 km wide and up to 100 m deep (Fig. 6 B), occupied by marine mudstones of the “green bar” member (Els and Karpeta 1997; Karpeta and Els 1999; Els 1998 b, 2000; Stewart et al. 2004; Tucker et al. 2016). These represent incised channel fills, cut during a period of rapid sea level fall, and filled during subsequent marine transgression. Planar cross-stratified sandstones and minor conglomerates at the base of these channels may be fluvial (Karpeta and Els 1999), with overlying strata (including slumped mudstones) representing an estuarine fill, and not meandering river deposits as suggested by Button and Tye (1979). Diamictite within channels in the eastern part of the Central Rand Goldfield (Martin et al. 1989) may be of similar subaqueous gravity flow origin (Fig 6 A), and do not represent glacial deposits as suggested by Wiebols (1955). The Main Formation above the green and black bar mudstones is characterized by a thin basal framework-supported conglomerate, which in the composite reef in the West Rand Goldfield rests directly on remnants of the Main reef and green bar (Fig. 7 B) and is directly overlain by massive fine sandstones of marine origin (Stewart et al. 2004). The flat top of this well sorted small pebble conglomerate (below the massive sandstone in Fig. 7 B) is consistent with marine reworking during transgression. In the Welkom Goldfield (SW end of basin) the upper part of the Main Formation begins with 35–100 m of predominatly trough cross-stratified sandstone (62%) and pebbly sandstone (27%), with minor planar cross-stratified sandstone (2%) and pebbly sandstone (5%), massive sandstone (3%), and thin pebble lags (0.4%), consistent with deposition in sandy braided streams (type L: Els 1991). Minor, thin, silty mudstones (1.1%) may represent fill of abandoned channels rather than overbank deposits. This in turn is overlain by the 1-7 m thick Middelvlei placer (Els 1987, 1991, 1993, 2000; Els and Beukes 1987) that represents a gravelly-sandy braided system. The presence of ~9% of planar laminated sandstone in this placer implies strong seasonal discharge, indicating a depositional style transitional between models B, L and O (Fig. 1.). Sandstones in the Main Formation above the placer appear to have been deposited in sandy braided rivers (type L).

Sandstones and minor conglomerates in the Randfontein Formation (including the Johnston reef) along the northern and western margins of the basin may be fluvial (Guy et al. 2010) passing into marine strata towards the basin centre. Eriksson and Simpson (2012) found clear evidence of tidal deposition in the upper part of the formation. The overlying Luipaardsvlie Formation is predominatly sandstone, with several prominent conglomeratic horizons (Livingstone, Intermediate, Doornfontein and Cobble reefs). The Livingston placer may be a fan (Minter et al. 1986; Minter and Loen 1991; Minter 2006), while others may represent shallow gravel-bed river deposits (type B). Sandstones at the top of the formation along the western margin of the basin (Klerksdorp Goldfield) were interpreted as pro-deltaic (marine) by Catuneanu and Biddulph (2001).

The Krugersdorp Formation non-conformably overlies the Luipaardsvilie Formation and contains a number of placer deposits (Basal, Steyn, Saaiplas, Vaal, Bird, White and Monarch reefs), and is overlain by storm-influenced marine mudstones of the Booysens Formation (Karpeta and Els 1999). The Basal and Steyn reefs in the Welkom Goldfield, at the southern end of the basin, are typically < 1m thick, and have been interpreted as shallow gravel-bed rivers that developed on a low-gradient peneplain, that passed down-slope into a braid delta (Minter 1976, 1978; Minter et al. 1986; Bailey 1991). Although some evidence exists for development of lateral accretion surfaces on channel bends in the Steyn placer (Fig. 7 C, and Fig. 9 in Minter et al. 2006) there is no clear evidence of meandering stream development. Minter (1978) found a marked decrease in both channel depth and gravel abundance down-slope (Figs. 7 C, D). The upper surface of the reefs may have been subject to marine reworking, as they are typically overlain by marine sandstone (Minter 1976, 1978; Bailey 1991). Discontinuous conglomerates in the Leader reef, 10 – 15 m above the basal reef in the Welkom Goldfield (Bailey 1991), locally occupy channels up to 200 m wide and 3 m deep, and are overlain by up to 14 m of sandy braided stream deposits (L) overlain by pro-deltaic or marine deposits (Kingsley 1984).

Further north, at the base of the Krugerdorp Formation in the Klerksdorp Goldfield, the fluvial (B, L) Mizpah and Zaaiplaat conglomerates appear to infill shallow (~ 1 m) low-stand incised channels up to 1 km wide Catuneanu and Biddulph (2001). These authours provide clear evidence of marine reworking of these distal gravel and sand bed rivers to produce an extensive transgressive marine gravel sheet (Vall placers), and an overlying storm-influenced marine sandsheet (Stilfontein quartzites). These are capped by further marine sandstones showing evidence of deposition from long-shore currents (Minter 1972, 1976; Catuneanu and Biddulph 2002) and marine mudrocks in the Booysens Formation (Tweedie 1968).

The Tuffontein Subgroup represents the upper part of the Central Rand Group of the Witwatersrand Supergroup (Fig. 5). The subgroup begins with sandstones of the Dornkorp Formation (Spes Bona Formation of Bailey 1991) that lies directly on the “B” reef. Van den Heever (2008) interpreted the B placer as a composite reef, with a lower interval of channel confined matrix-supported or framework-supported massive or weakly stratified small pebble to cobble conglomerate and interbedded with sheets of trough cross-stratified, massive, and weakly laminated sandstone (Fig. 7 A) that is overlain by a well sorted small pebble conglomerate, possibly reflecting marine reworking. Carbon may be present as a sheet at the base of the channels, or dispersed in other strata. Minter (1978) and Minter et al. (1986) indicated that the conglomerates were confined to possible lowstand incised channels, up to 200 m wide and 2 m deep, that have an interconnected braided pattern and shallow down-slope. This is consistent with deposition in gravel-bed rivers (type B) on a distributary fan system (Davidson et al. 2013). Van den Heever (2008) was able to identify mid-channel and side-bars, as well as confluence bars, and like Minter (1978) found no evidence of deposition from meandering channels. Minter interpreted the overlying 90 m of cross-stratified sandstone and minor conglomerate as fluvial. Elsewhere, in the eastern part of the Central Rand Goldfield the basal conglomerate sheet (Zandfontein reef), and overlying 9 m of interbedded planar and trough cross-stratified fine to medium grained sandstone and mudstone, have been interpreted as marine (Bailey et al. 1989), based in part on the presence of bimodal opposed cross-stratification. Marine influence has also been demonstrated by Els and Meyer (1989) in equivalent strata (Buffelskloof Formation), around the Vredefort structure, where medium to very coarse cross-stratified sandstone, with minor small pebble conglomerate, pass eastwards into marine sandstones with bimodal opposed cross-stratification.

The overlying Kimberley Formation begins in the southern part of the Witwatersrand basin (Welkom Goldfield) with a distinct incised channel fill complex, known locally as the Aandenk Channels (Bailey 1991). These contain a diverse assemblage of sandstones, pebbly sandstones, diamicts, framework conglomerates and (slumped) mudstones that suggest deposition within an estuarine setting, during marine transgression (Fig. 8). The overlying "big pebble" conglomerate may reflect deposition of gravel-bed river deposits initiated by sea level fall, followed by marine peneplanation during subsequent sea level rise. Similar channel fill sequences have been recorded in the East Rand beneath the May (UK9) placers (Tweedie 1986) that Hirdes and Saager (1983) suggested developed on a series of basin margin fans. Earlier Hirdes (1978) had indicated that some distal conglomerates in this area were developed within high sinuosity (1.26-1.6) meandering channels, up to 0.4 m deep, based on local curvilinear patterns of gold distribution. No evidence of lateral accretion was documented in the channels, and associated plots of gravel distribution are more consistent with a braided fluvial origin.

<Fig.8 here>

The ~2 m thick Witpan conglomerate member occurs ~50 m above the "big pebble" marker (Karpeta 1993). It represents deposition of well sorted, clast-supported, large pebble conglomerates in relatively deep gravel-bed rivers (B or C), with terraces, and contains (architectural) evidence of side-bars, longitudinal and diagonal gravel bars, and areas of flow convergence. Karpeta (1993) indicates that the main channels were 1 to 2 m deep and up to 120 m wide, with low sinuosity. Terraces had a relief typically less than 0.2 m, and were up to 500 m wide. The presence of minor (<1.5%) laminated and rippled mudrocks indicates suspension deposits in channels and possibly in floodplain lakes in the terrace settings. Alternatively this may indicate tidal (backwater) effects on the distal braided system. The upper surface of the Witpan conglomerates may have been wave-reworked as it is directly overlain by flat laminated sandstone of probable marine origin. The uppermost part of the Kimberley Formation, above the Uitsig conglomerate, in the southwest part of the basin, is dominated by trough cross-stratified sandstones with unidirectional paleocurrents (Karpeta 1993), and may be of sandy braided stream (L) origin.

The overlying Elsberg (Eldorado) Formation includes both marine and fluvial strata (Karpeta and Els 1999; Guy et al. 2014). In the Welkom Goldfield Minter et al. (1988) suggested that much of the formation near the southwest basin margins is of alluvial fan origin, with evidence of ephemeral conditions indicated by up to 85 cm of thin (1-5 cm) sets of muddy pebbly sandstone (diamictite) containing reworked mud-flakes, separated by thin mudstone partings that may represent suspension deposition in channels following major floods. The presence of faceted pebbles in conglomerates of the Beatrix reef, at the base of the formation, indicates local reworking of exposed surfaces by aeolian activity (Minter 1978, 1999), consistent with deposition in type A rivers with debris flows. Gold concentrations in the Elsberg Composite reefs in the Central Rand Goldfield show evidence of deposition by a radiating series of channels, consistent with fluvial deposition on a small proximal fan or distributary fluvial system (type B; Fig. 6 C; Kingsley 1987; Frimmel et al. 2005; Tucker et al. 2016; Viljonen et al. 2016).

The stratigraphically highest part of the Witwatersrand Supergroup is marked by renewed development of alluvial fans or distributary fluvial systems around the basin margin (Tucker et al. 2016; Viljonen et al. 2016), with conglomerates and pebbly sandstones of the Mondeor Formation possibly representing deposition in shallow gravel-bed rivers (type B).

Along the eastern margin of the Kaapvaal Craton (Fig. 5) strata that is in part time equivalent to the lower part of the Witwatersrand Supergroup accumulated in a back-bulge setting (Bumby et al. 2011) locally influenced by rifting (Burke et al. 1985). In the upper part of the Pietersburg Greenstone belt clastic strata in the 2.90-2.88 Ga Uitkyk Formation (Saager and Muff 1978 a, b; Zeh and Gerdes 2012; Zeh et al. 2013), are of similar age to parts of the Central Rand Group (de Wit et al. 1993). They were interpreted as deposits of alluvial fans and braidplains (B, L) by de Wit et al. (1992), although few details of the stratigraphy or sedimentology of the unit were provided. Saager and Muff (1978 b) indicated that the conglomerates lie above a thick sequence of sandstones, and are marked by rapid lateral facies change. They are overlain by mudstones and siltstones that are cut by local channels filled with cross-bedded coarse sandstone, followed by a succession of poorly sorted chert-bearing sandstone. Jones and de Wit (1986) and de Wit et al. (1992) suggested that the unit accumulated in a linear piggy-back basin.

In the Pongola Supergroup, strata at the base of the Mozaan Group are possible time equivalents of the Orange Grove Formation at the base of the West Rand Group (Fig. 5). Fluvial strata include conglomerates of the Deny-Dalton member of the Agatha Formation and overlying sandstones at the base of the Sinqeni Formation (Watchorn 1980 b; Saager et al. 1981; Dix 1984; Beukes and Cairncross 1991; Nhleko 1998; Hicks 2009; Hicks and Hofmann 2012). Strata include massive framework (36%) and matrix-supported (19%) medium to large pebble conglomerates, pebbly sandstones (36%), and planar and trough cross-stratified sandstones (8%), consistent with deposition in shallow to deep sandy gravel-bed rivers (type C), and where conglomerates are sparse, sandy braided rivers (type L). Minor mudstones may represent abandoned channel fills

(Watchorn 1980 a, b). The fluvial strata pass up-section into shallow marine strata (Hicks 2009; Hicks and Hofmann 2012).

2.3.3. Canada/USA

In North America, Mesoarchean fluvial strata have been recognized in the Slave and Superior structural Provinces (Fig. 4). The oldest known fluvial strata in the Superior Province are in the Sachigo subprovince, where fluvial strata are known from the North Spirit Lake and North Caribou Lakes Greenstone belts. In the North Spirit Lake area, Wood (1980) described massive conglomerates and planar laminated and trough cross-stratified sandstones in the 3.067-2.962 Ga Makatiamik Assemblage of Préfontaine et al. (2008) as representing an alluvial fan complex (type A) overlain by sandy braided river deposits (type L). Further north, in the North Caribou Lakes area, the 2.959-2.869 Ga Eyapamikama Lake, matrix- and clast-supported cobble and pebble conglomerates have also been interpreted as both alluvial fan (at the base) and submarine fan deposits (at the top) (Breaks et al. 2001). The conglomerates are typically massive, although imbricated, planar stratified, graded and inversely graded units are present locally. The presence of scour and fill structures, and local imbrication favours deposition of the basal part of the formation from type A rivers on alluvial fans. Local breccias (volcani-rudites) may represent scree (tallus) deposits. Conglomerates in the upper part of this succession may represent gravity flow deposits on the subaqueous part of a prograding fan-delta

In the Steep Rock Lake area of the Wabigoon subprovince of northern Ontario ~15 m of deformed massive pebble conglomerate (volcani-lith rudite) and minor sandstone and pebbly sandstone of the 2.997- >2.780 Ga Wagita Formation infill a shallow valley cut into the top of a volcanic plateau (Wilkes and Nisbet 1988; Kusky and Hudlestone 1998; Stone 2010; Fralick and Riding 2015). Fralick et al. (2008) and Fralick and Riding (2015) suggested that this ~500 m wide valley may have developed as a feeder for a fan-delta sequence that was subsequently flooded during marine transgression. Overlying sandstones are of pro-deltaic origin (Fralick et al. 2008; McIntyre and Fralick 2017). Fluvial strata of similar age (2.997-2.921 Ga) have also been recognized 400 km to the northwest in the Wallace Lake Greenstone belt, in the lower 200m of the Conley Formation, (Sasseville 2002; Fralick et al. 2008), where thin (30 cm) lenses of pebble conglomerate and trough cross-stratified medium and coarse grained sandstone may represent sandy braided river deposits (type L). Elsewhere the formation is dominated by massive thick-bedded sandstones, and may have been deposited in a subaqueous, pro-deltaic fan environment, rather than on arid region alluvial fans as suggested by Sasseville (2002).

Further to the northwest, ~~in the Sachigo subprovince, fluvial strata are known from the North Spirit Lake and North Caribou Lakes Greenstone belts. In the North Spirit Lake area, Wood (1980) described massive conglomerates and planar laminated and trough cross-stratified sandstones in the 3.067-2.962 Ga Makatiamik Assemblage of Préfontaine et al. (2008) as representing an alluvial fan complex (type A) overlain by sandy braided river deposits (type L). Further north, in the North Caribou Lakes area, the 2.959-2.869 Ga Eyapamikama Lake, matrix- and clast-supported cobble and pebble conglomerates have also been interpreted as both alluvial fan (at the base) and submarine fan deposits (at the top) (Breaks et al. 2001). The conglomerates are typically massive, although imbricated, planar stratified, graded and inversely graded units are present~~

~~locally. The presence of scour and fill structures, and local imbrication favours deposition of the basal part of the formation from type A rivers on alluvial fans. Local breccias (volcani-rudites) may represent scree (tallus) deposits. Conglomerates in the upper part of this succession may represent gravity flow deposits on the subaqueous part of a prograding fan-delta.~~

In the Slave structural Province, possible fluvial strata have been identified in a number of related units in the 2.92-2.81 Ga Bell Lake and Beniah Lake groups of the Central Slave Group (Bleeker et al. 1999; Corcoran 2012). Muller and Corcoran (2001) described sheets of breccia, conglomerate, and sandstone in the Raquette Lake Formation that they interpreted as tallus, fan, and fan-delta deposits that accumulated in a mature backarc basin. They suggested that matrix-supported conglomerates represent deposits of cohesive debris flows in a subaerial setting (type A). Associated massive or weakly stratified framework-supported conglomerates are interpreted as flash flood deposits, while interbedded planar laminated sandstones are interpreted as non-confined sheetflood deposits (type P). Although a fluvial interpretation is possible, there is no clear evidence of exposure or complex architecture, and the high lateral continuity of the conglomeratic beds is more consistent with deposition in a pro-fan-delta setting. Associated massive and planar laminated sandstone facies have been interpreted as storm-influenced shallow marine by Muller and Corcoran (2001), as are other sandstone dominated sequences at the same stratigraphic level in other parts of the province (Roscoe et al. 1989; Rice et al. 1990; Bleeker 2001; Pickett 2002). In the central Slave Province the Beniah Lake Formation, contains several intervals of well sorted slightly radioactive pyritic quartz pebble conglomerate, which Roscoe et al. (1989) interpreted as being deposited by “especially vigorous currents” in a “non-turbiditic” setting, implying a fluvial origin. Subsequent analysis by Rice et al. (1990), confirmed that these were of coastal marine (beach) origin, and contain no fluvial component, and that associated massive, planar laminated, rippled and small-scale cross-stratified sandstones and lesser mudstones are all of shallow marine origin. Muller and Pickett (2005) and Muller et al. (2005) confirmed the marine origin of these and associated strata (Bleeker et al. 1999) in the Bell Lake Group, suggesting deposition in a fluvially-influenced macrotidal, possibly estuarine setting.

2.3.4. India

Possible Mesoarchean fluvial strata have been identified at the base of the 2.96-2.6 Ga Bababudan Group in the Dharwar Craton of western India (Srinivasan and Ojakangas 1986; Raju and Eriksson 2015). Massive, planar bedded and trough cross-stratified (pyritiferous) quartz pebble conglomerates in the basal 1-12 m of the formation are interpreted as the deposits of shallow gravel-bed rivers (type C), possibly influenced by flashy discharge (Srinivasan and Ojakangas 1986). Overlying quartz arenites (with no pebble layers) are interbedded with amygdaloidal mafic flows and lahars, and are characterized by trough cross-stratification in the lower part of the section, with increasing abundance of planar cross-stratified sandstones higher in the section. These have been interpreted by Srinivasan and Ojakangas (1986) as deposits of Platt type rivers (type L) that accumulated on a peneplain in a stable cratonic setting.

The Singbhurn Craton of eastern India hosts fluvial strata at the base of the predominantly marine 3.16-2.8 Ga Mahagiri quartzite, and its equivalents

(Mukhopadhyay et al. 2016; Ghosh et al. 2016; Kumar et al. 2017). The basal conglomerates include both massive matrix conglomerates indicating deposition by debris flows (Mukhopadhyay et al. 2016), and framework-supported conglomerates, indicating normal fluvial flows on an alluvial fan or braid delta. Overlying sandstones are dominated by trough cross-stratified sandstones, implying deposition in sandy braided fluvial systems (type L) in a coastal braidplain.

2.3.5. Eurasia

Mesorchean fluvial strata are largely absent, or under-reported in Eurasia. In the far-east, Kepezhinskas (2016) has recorded 3.1-2.7 Ga gold-bearing conglomerates in the Nemui Formation of Eastern Siberia. Similar deposits may also be present in other parts of the North Asian Craton, but are deeply buried (Starostin et al. 2016 a, b).

In the Baltic shield 2.83-2.72 Ga conglomerate-bearing terrigenous rocks have been recorded in the Kola province (Höltta et al. 2008), and Hattu Schist belt (Slabunov et al. 2006), but no details of depositional style are provided.

2.4. Neoarchean (2.8-2.5 Ga)

The Neoarchean marks an interval when modern style plate tectonics were active, and supercontinents were a conspicuous component of the Earth (Van Kranendonk 2007 b). Clastic fluvial strata have been recorded from many cratonic areas (Fig. 9), and are represented by deposits in Africa, North and South America, and Australia, but have yet to be recorded in Antarctica, or Eurasia. Deposits are dominated by alluvial fan and braided river facies, locally interbedded with aeolian strata, that accumulated in rift, strike-slip, foreland basin, and to a lesser extent, piggy-back basins.

<Fig 9 here>

2.4.1 Australia

Neoarchean fluvial strata have been recognised in both the Pilbara and Yilgarn Cratons of Western Australia, and possibly in the Gawler Craton of South Australia.

The laterally extensive 2.772-2.715 Ga Fortescue Group overlaps at least 90, 000 Km^2 of the Pilbara Craton (including the Lalla Rookh basin). It accumulated in a rift to passive margin setting, and includes both fluvial strata, and complex packages of subaerial basalts at a number of levels (Thorne and Trendall 2001; Bolhar and van Kranendonk 2007; Awramik and Bucheim 2009). Strata below the 2.772-2.763 Ga Mount Roe Basalt, at the base of the Fortescue Group, accumulated in a variety of fluvial and lacustrine environments in small, restricted basins (Blake 1993; Thorne and Trendall 2001). Massive matrix- and clast-supported conglomerates at the base of the formation infill depressions up to 35 m deep. Associated lenses of coarse grained sandstone are characterised by stacked sets of medium-scale cross-stratification. Overlying sandstones are dominated by finer grained sandstone and conglomerate with planar bedding and low-angle cross-stratification. Thorne and Trendall (2001) suggested that these represent valley-fill alluvial and colluvial facies (type A) and associated sheetflood deposits (type P). In the Marble Bar area as much as 0.7-1.8 km of very coarse grained arkosic sandstone with minor conglomerate of probable braided river facies (type L?) are present below the Mont Roe Basalts (Blake 1993). Nearer to the top of the succession, in the Turee Creek area, Blake (1993) and Thorne and Trendall (2001) recognized a subaerial fan-delta facies, consisting of massive and planar stratified conglomerate, pebbly sandstone, and coarse grained sandstone, interlayered with thin basalt flows and reworked tuff.

The Mount Roe Basalts are overlain by as much as 3 km of sandstone, conglomerat and minor mudston, volcanic strata and basalt flows of the ~2.756 Ga Hardey Formation, part of which is of fluvial origin. Thorne and Trendall (2001) interpreted massive clast- and matrix-supported conglomerates at the base of the formation as deposits of gravel-bed rivers with debris flows (type A) that may have accumulated in an alluvial fan setting. Sandy braided stream deposits, characterized by sheets with stacked sets of trough (and minor planar) cross-stratified sandstone, and minor planar laminated sandstone with current lineations, were compared to (type M) fluvial strata in the Devonian Battery Point Formation in eastern Canada, despite a maximum estimated channel depth of only 1.5m. In southern parts of the basin the upper part of the formation reflects deltaic to shallow coastal lacustrine conditions (Thorne and Trendall 2001).

In the upper half of the Fortescue Group, Sakurari et al. (2005) found pebble and cobble conglomerate interbedded with coarse to very coarse grained sandstone in erosional depressions at the base of the 2.72 Ga Tumbiana Formation. The conglomerates are typically framework-supported, and include planar bedded and planar and trough cross-stratified varieties, in units up to 1 m thick. Interbedded sandstones occur in lenses up to 0.5 m thick, characterized by planar and trough cross-stratification. Sets of mudstone and muddy sandstone in units up to 2.3 m thick are present in the upper part of the succession, interbedded with very fine to medium grained sandstone, in units up to

0.1 m thick. Sakurari et al. (2005) interpreted the conglomerates as deposits of gravel-bed rivers on an alluvial fan, with local evidence of hyper-concentrated flows indicated by local inverse and inverse to normal grading. As no evidence could be found to support a wedge shaped geometry, these conglomerates could alternatively represent gravel-bed river systems that fill incised river valleys (type B). Mudstones in the upper part of the package may represent distal floodplain (playa) facies, or could reflect rising water levels within the basin, and may mark a transition to predominantly marine (Sakurari et al. 2005) or lacustrine (Bolhar and van Kranendonk 2007; Awramik and Bucheim 2009) facies in the upper parts of the formation.

In the Yilgarn Craton Giles and Hallberg (1982) and Hallberg (1986) have suggested that poorly sorted lithic wackes within the Welcome Well Volcanic complex (part of the 2.692-2.655 Ga Black Flag Group) represent alluvial mass-flow deposits that accumulated on the flanks of a large stratvolcano. As these are locally overlain by turbidites, a subaqueous fan (pro-fan-delta) is considered more likely (Brown et al. 2001).

Fluvial to deep-water strata are also preserved in a number of linear late strike-slip basins that formed between 2.658 and 2.648 Ga within the eastern Goldfields (Krapež et al. 2008; Brown et al. 2001; Krapež and Barley 2008; Krapež and Pickard 2010; Wyche 2012). Fluvial strata are well developed in the Mergougil basin, where Krapež et al. (2008) recorded > 1.6 km of sheet-like conglomerates and sandstones of axial braidplain origin. The conglomerates include both clast-supported and matrix-supported pebble and cobble lith rudites, as well as single pebble lags. Associated sheets of predominantly medium to coarse grained sandstone are either massive or planar bedded. Krapež et al. (2008) also records channel fill sequences of planar-laminated sandstone, often with a single pebble lag at the base, and sheets of trough and ripple cross-stratified sandstone. Krapež et al. (2008) interpreted the conglomerates as within channel and between channel deposits of gravel-bed rivers (type B?) with associated planar-laminated sandstones reflecting upper flow-regime deposits that accumulated on bar tops and in small braid channels. Thicker sheets of planar-laminated sandstone were interpreted as deposits of sheetfloods (type O or ? P). Local sheets of trough and ripple cross-stratified sandstone interbedded with the planar laminated sandstones were interpreted as products of deposition from dunes under lower flow-regime conditions. Similar facies were recorded in the Navajo Sandstone below turbidites of the <2.679 Ga Kurrawang Formation, and in strata in the Jones Creek Formation, where conglomerates are predominantly massive, clast-supported, cobble and boulder lith rudites (type B), with lesser abundance of massive and planar laminated sandstone of flash flood origin (type O). The Scotty Creek Formation is dominated by sheets of clast-supported cobble and pebble conglomerate of shallow gravel-bed river origin (type B). Minor sheets of massive, planar bedded and locally trough cross-stratified channel-fill sandstones (up to 3 m thick), represent deposits of shallow sheetflood dominated sandy braided rivers (type O). Minor laminated siltstones overlie some of the sandstone sets and may represent overbank deposits (Krapež et al. 2008). The Yandal basin contains a similar suite of fluvial facies, but also include some larger-scale planar cross-stratified sandstone, which may represent deposition in axial Platt type river systems (type L).

The Gawler Craton of South Australia contains some highly metamorphosed (~2.52-2.48 Ga) sandstones of uncertain origin, associated with rifting in a backarc setting (Reid and Hand 2012). Zang (2007) tentatively interpreted "layered to cross-layered" metasandstone, at one locality in the Mulgathing complex, as fluvial or estuarine in origin, but did not provide definitive evidence. Given that these strata are highly sheared amphibolites, associated with komatiitic rocks, a fluvial interpretation must be considered as highly suspect.

2.4.2. Africa

Fluvial strata are present at the base of the 2.714-2.665 Ga Ventersdorp Supergroup in southern Africa (Eriksson et al. 2002; Schneiderhan et al. 2011), where they underlie flood basalts of the Klipriversberg Group, and overlap strata of the Witwatersrand basin. Krapež (1985) noted that these strata of the Ventersdorp Contact Placer (Venterspost Fm) were concentrated in incised valleys developed on a bedrock pediment. Krapež (1985) recognized massive matrix-supported conglomerate, with local inverse to normal coarse tail grading, which he considered to be of debris flow origin. Associated massive, planar bedded and cross-stratified contact framework conglomerates with well developed clast imbrication were interpreted as longitudinal bars developed in a gravel-bed river system under lower flow-regime conditions. Associated thin pebble lags and planar laminated medium grained sandstones represent erosional lags and strata developed under upper flow-regime conditions. A thin mudstone lamina, overlying basal conglomerates reflects deposition of muds in falling-stage or slack-water conditions. The overall interpretation is consistent with deposition from type A rivers in a confined (terraced) bedrock setting, with debris flows produced locally by slumping of material accumulated by earlier floods on the valley walls (McWah 1994). Rust (1994) suggested that the regional depositional slope beneath the conglomerates was in the order of 2-5 m/km, and that floodplain widths of individual river systems, made up of numerous braid channels, may have been as much as 2 km wide, with peak flow depths of 5-10m, and flow velocities of 2-3 m/s. Other minor fluvial units may also be present between flood basalts in the overlying, ~ 2km thick, Klipriversberg Group (Eriksson et al. 2002).

Anhaeusser et al. (2010) have suggested that minor layers of laminated granular sandstone, interbedded with conglomerate in the upper part of the fill of the 2.79-2.71 Ga impact structure at Setlagole (NW Kapvaal Craton), may be fluvial. As the bulk of the formation is boulder grade breccia of mass-flow origin, with deposition presumably triggered by impact tsunami or resurge in a marine setting, a (subaqueous) resurge origin appears more likely.

Strata in the volcaniclastic 2.714-2.709 Ga Kameeldoorns Formation, at the base of the overlying Platberg Group, may include basin margin alluvial fans, which prograded into deeper water mudrocks away from bedrock horsts (Buck 1980; Crow and Condie 1988; Karpeta 1993; Eriksson et al. 2002). In the Hartbeesfontein basin the Kameeldoorns Formation is characterized by massive, very poorly sorted, matrix-supported conglomerate (diamictite), in beds up to 8m thick, containing clasts up to 2 m in diameter that Myers et al. (1990), and Karpeta (1993), suggested were of mass-flow origin. As no interbeds of cross-stratified sandstone are present it is probable that only the subaqueous portion of these basin margin fan-deltas is preserved. In the vicinity of the Welkom Goldfield Buck (1980) recognized similar stata in the Video member of his lower Klippan Formation, which pass laterally into lacustrine facies. He interpreted these as deposits of tallus slopes and debris flow dominated alluvial fans (type A), and interpreted sandstones in the overlying Dirksberg member as braidplain deposits that infilled the local grabens. He noted that the latter occurred as fining upward cycles, capped by mudstones with local curled laminae and dessication cracks. These may represent ephemeral fluvial deposits (type O) or may represent playa margin sediments. Conglomerates and sandstones at the base of the stratigraphically equivalent Mohle Formation of the Hartswater Group, in a graben west of the main Ventersdorp basin (de Kock et al. 2012) may have a similar origin, as may basal strata of the basal Platberg Group, intersected in boreholes in the Virginia basin, near the south end of the Welkom Goldfield (Meintjes et al. 1989).

Clastic strata in the Bothaville Formation, which lies stratigraphically above the Platberg Group, at the base of the 2.712-2.64 Ga Pniel Group (Schneiderhan et al. 2011), have been interpreted as gravel-bed river deposits (type B) by both Buck (1980) and Karpeta (1993). In the Welkom basins the formation consisting of poorly sorted contact framework conglomerates, interbedded with lenses of cross-stratified and planar bedded sandstone, consistent with deposition in type B rivers. Associated mudstone (drapes and beds) are planar laminated, with some dessication cracks and mud curls (Buck 1980). In the Hartbeesfontein basin (Karpeta 1993) equivalent strata are only 20 m thick, and are characterized by trough and planar cross-stratified, and planar laminated sandstones (with current lineations), that are associated with minor discontinuous clast-supported conglomerates in lenses up to 1.5 m thick and 20 m across. Karpeta (1993) suggested that the conglomerates and cross-stratified sandstones represent deposits of gravel-bed rivers (type A), while the laminated sandstones represent deposition by sheetfloods in shallow parts of the river system (or alternatively sheetflood dominated distal river systems of type O).

Strata of the Sodium Group, on the southwestern corner of the Kapapvaal Craton, have been correlated with the Platberg Group in the Virginia basin (Meintjes et al. 1989; Van der Westhuizen 1991; Altermann and Lenhardt 2012). Poorly sorted clast- and matrix-supported conglomerates at the base of the Ongers River Formation may represent debris flow and streamflow deposits (type A) on basin margin fans, as suggested by Grobler et al. (1989). These are locally interbedded with, and overlain by, lenticular bed of planar cross-stratified, coarse grained arkose, with lesser graded sandstones and minor tuff layers, that Grobler et al. (1989) interpreted as sandy braided river deposits. Alterman and Lenhardt (2012) suggested that some of the sandstones represent flash flood deposits, and that associated fine sediments accumulated in distal lakes and as pro-deltaic turbidites, a setting similar to that proposed for the stratigraphically equivalent Kameeldoorns Formation in the main parts of the Ventersdorp Supergroup (see above). Grobler et al (1989) also suggested that minor intervals of cross-stratified sandstones, interlayered with mafic and intermediate lava flows (and minor straomatolitic carbonates) in the Omdraavlei Formation, near the top of the Sodium Group, could represent piedmont fan systems, but gave insufficient information to determine fluvial style. Alterman and Lenhardt (2012) identified two facies in this formation that might have been deposited in fluvial systems: the first (their facies LFT4) was characterized by massive to poorly stratified sandy gravels of possible mass-flow origin, and the second (their facies LFT5) consisting of massive, laminated, cross-stratified and ripple cross-stratified tuffaceous fine to medium grained sandstones, with minor mudstone intervals up to a few decimetres thick. They interpreted these deposits to be the product of deposition in a shallow lacustrine or fluvial setting, with local presence of rain imprints and dessication polygons providing evidence of at least local exposure. This is more likely to indicate deposition in a playa lake setting, with local graded sandstones produced as turbidites in a pro-deltaic setting.

Outside of the main outcrop area of the Ventersdorp Supergroup, possible fluvial strata are preserved below the Transvaal Supergroup in a number of small pull-apart basins, that have collectively been referred to as "protobasinal" (Tankard et al. 1982; Eriksson and Reczko 1995; Eriksson et al. 2002). These include strata in the Buffelsfontein (Buffalo Springs) Group and Wolkberg Group. Barton et al. (1995) suggested that strata in the 2.743-2.664 Ga Buffelsfontein Group were contemporary with strata of the Kameeldoorns and Bothaville Formations in the Venterdorp Supergroup (see above). The (basal) Hampton Formation contains a range of clastic material that Tyler (1978 a, b, 1979 a) interpreted in terms of braided stream, meandering stream and near shore (shallow marine) facies. The "arkose and shale" member, at the base of the formation in the northeastern part of the basin, is characterized by interbedded packages of granule conglomerate and coarse to fine grained sandstone, separated by intervals of laminated mudstone. Flute casts are present at the base of some units where they overlie mudstones. Basal sections of the sandy units are characterized by trough cross-stratifcation, and are typically succeeded up section by planar bedded sandstone with only minor trough cross-stratified intervals. Mudstone intraclasts are common on scour surfaces within and at the base of the sandstones. Mudstone intervals make up 25% of the member, in units up to 5 m thick. They include graded sand to silt beds and sand-filled dessication cracks up to 50 cm deep. Tyler (1978 a, b, 1979 a) argued that this combination of features is best interpreted as reflecting deposition in a sandy meandering river system (type G), although deposition in a distal flash flood dominated sandy system (type O) adjacent to a playa lake would explain the abundance of mudstone, the depth of dessication features, and the absence of lateral accretion surfaces. Overlying the basal member is a succession of quartz arenites with low-angle cross-stratification, with heavy mineral laminae, and bimodal opposed cross-stratified sandstone, indicating deposition in a near shore to shallow marine setting (Tyler 1979 a). Higher in the formation, sandstones (and minor conglomerates) in the "middle arenite member" are characterized by small-scale cyclical units with minor low-angle trough cross-stratification, overlain by predominantly flat lamination (Tyler 1979 a). This is consistent with deposition in ephemeral systems (type O and possibly N). Local lobes of clast-supported cobble and pebble conglomerate, with minor interbeds of pebbly sandstone in the "conglomerate member" were interpreted by Tyler (1979 a, b) as braided stream deposits (type B) that because of their lobate geometry might be interpreted as basin margin fans. Tylers's "upper sandstone member: is dominated by flat laminated sandstones with minor siltstone drapes with mud cracks, consistent with deposition in a distal sheetflood dominated fluvial system (type O).

In the Wolkberg basin, strata of possible fluvial origin are present in the Sekororo and (stratigraphically higher) Schelem formations (Button 1972, 1973; Bosch et al. 1993; Tankard et al. 1982). Bosch et al. (1993) describe the (basal) Serororo Formation as a valley-fill succession that begins with an upward fining succession of matrix-supported cobble and pebble conglomerates, overlain by 1-28 m thick cycles of massive and planar bedded framework-supported conglomerate, and pebbly planar and trough cross-stratified sandstones, with minor laminated mudstone and siltstone units (1 mm to 2 m thick).

These are overlain by a coarsening upwards succession of 1-40 m thick units of planar bedded, and planar and trough cross-stratified, sandstone, with minor laminated mudstone. Massive, framework-supported conglomerate and pebbly sandstone become more prominent up section, with local development of clast imbrication in the conglomerates. Bosch et al. (1993) interpreted the basal sequences as alluvial fan deposits, influenced by debris flows (type A). Sandstones could represent a combination of shallow perennial braided rivers (type L) and more distal sheetflood dominated systems (Tyle O). Conglomerates at the top of the formation appear to have been deposited in shallow gravel-bed rivers (type B).

The Schelem Formation contains 1-15 m thick massive contact framework conglomerates at the base, overlain by fining upwards, cyclic alternations of normally graded and planar and trough cross-stratified conglomerate, planar bedded and planar and trough cross-stratified sandstones and laminated to massive mudstone. Interbedded sandstones and mudstones become more abundant towards the top of the formation. These strata are interpreted by Bosch et al. (1993) as including deposits of fan, fan-deltas and braidplains in a semi-arid environment, consistent with deposition in shallow gravel-bed rivers (type B), and more distal sheetflood dominated systems (Tyle O). Elsewhere in the basin there is evidence of deposition from local debris dominated fans (type A), and from subaqueous flows in a pro-deltaic setting, with mudstones preserved in floodplain and pro-delta lakes, and possibly as overbank deposits on floodplains.

The 2.642-2.55 Ga Black Reef Formation marks the base of the predominantly marine Transvaal Supergroup, that represents deposits of an intracratonic sag basin, and lies unconformably above strata of the Witwatersrand and Ventersdorp Supergroups

(Catuneanu and Eriksson 1999). The basal part of the formation consists predominantly of trough cross-stratified pebbly and coarse grained sandstones, with minor contact framework medium pebble to cobble conglomerate and conglomeratic sandstone, that are largely confined to valleys that cut the underlying bedrock (Eriksson 1972; Tyler 1979 b, c; Clendinin et al. 1991; Van den Berg 1994; Els et al. 1995; Coetzee 1996; Catuneanu and Eriksson 1999). These represent bedrock-incised gravelly-sandy braided rivers (types B and L). Tyler (1979 c) recorded local matrix-supported conglomerates that may represent mass-flows, derived from the valley walls, likewise Coetzee (1996) found minor accumulations of massive monomict breccia, which he interpreted as rock-avalanche deposits (tallus) derived from local cliffs. Much of the upper part of the formation is characterized by a discontinuous sheet of finer grained sandstones, with planar lamination, ripple, trough and planar cross-stratification, and mudstone. This is typically interpreted as deposits of coastal to shallow marine origin (Tyler 1979 c). At other locations the upper part of the formation is dominated by fine to medium grained sandstones, with adhesion warts, dessication cracks, and raindrop impressions. Associated massive and laminated siltstone and mudstone contain halite clasts, indicating deposition in a supratidal or sabkha setting within 10 to 30º of the equator (Eriksson et al. 2005). Beach, aeolian and shallow marine strata have also been recognized in equivalent strata of the Vryberg Formation at the base of the Transvaal Supergroup in the Griqualand West area (Schröder et al. 2006; Alterman et al. 2014).

Neoarchean fluvial strata have been recognized in several formations in the Zimbabwe Craton. Hunter et al. (1998) identified fluvial strata in the lower and middle parts of the ~2.7 Ga Manjeri Formation, in the Belingwe Greenstone belt in the southern part of the craton. The basal Spring Valley member of the formation is dominated by tide and wave influenced shallow marine sandstone. In the eastern part of the basin channelized conglomerates at the base of the member are identified as gravel-bed river deposits (type B). These are overlain by shallow marine strata, including both sandstones and stromatolitic carbonates. The middle Rubweruchena member is characterized by a wedge of poorly sorted conglomerate, or arkosic sandstone and conglomerate, that is not present in the western part of the belt. These are interpreted by Hunter et al. (1998) as deposits of small, basin margin alluvial fans and fan-deltas (type B?) that pass laterally and vertically into turbiditic facies. In the northern part of the craton Hofmann et al. (2002) identified fluvial strata in the Pote Formation, at the base of the ~2.68 Shamvaian Group. The basal 170 m of the formation is dominated by sheets of massive and crudely laminated framework-supported conglomerate, with minor interbeds of planar laminated, and trough cross-stratified pebbly coarse sandstone that Hofmann et al. (2002) identify as products of deposition by flash flood dominated rivers on unconfined segments of an alluvial fan (type A). Overlying this are ~250 m of planar laminated and trough cross-stratified pebbly sandstones, reflecting deposition in flash flood dominated sandy fluvial systems (type O). Massive channel fill conglomerates above the pebbly sandstones are interpreted as reflecting deposition in a gravel-bed river system (type B?) with highly variable discharge. In the central part of the craton Hofmann et al. (2004) indicated that fluvial strata may be present towards the top of both the upper and lower "greywacke" formations in the upper part of the ~2.68 Ga Shamvaian Group. Hofmann et al. (2004) suggested that although most of the "lower greywacke formation" was of marine tidal origin, that some cross-stratified sandstones, with mudstone drapes, near the top of the formation, might indicate deposition in shallow ephemeral channels (type O), or alternatively the mudstone may represent overbank deposits. The “upper greywacke formation” is dominated by shallow marine storm-influenced deposits. Fluvial strata were recognized at the top of the formation, overlying shoreface sandstones with local evidence of hummocky cross-stratification. They include thin layers of massive or normally graded clast-supported pebble conglomerate, and trough cross-stratified and planar laminated pebbly sandstones, consistent with deposition in a sandy braided system with variable discharge (type L).

Further north, on the Tanzania Craton, Sanislav et al. (2014) have suggested that metasediments in the Kavirondian Group represents a molasse facies deposited between 2.66 and 2.64 Ga (Henckle et al. 2016). Ngecu (1992) and Ngecu and Gaciri (1995) suggested that wedge-shaped bodies of framework- and matrix-conglomerate, and associated massive and laminated sandstone, in the Shivakal Formation, at the base of the group, were deposited on alluvial fans and fan-deltas. Although the inferred presence of imbrication in some beds might be supportive of fluvial transport, the formation is overlain by, and is laterally equivalent to, a thick sequence of turbidites, and so may be better interpreted as reflecting deposition from submarine fan systems. Crude planar bedding and pervasive clast alignment parallel to bedding (Ngecu 1992, p.79) is suggestive of either hyper-concentrated debris flows in a subaqueous setting, or later tectonic stretching.

2.4.3. Canada/USA

In North America Neoarchean fluvial strata have been reported from the Slave, Superior, Rae and Wyoming Provinces (Fig. 9).

In the Slave Province, Corcoran (2012) recorded numerous ~2.6 Ga conglomerate-bearing successions of possible fluvial origin in strata unconformably overlying the 2.72-2.661 Ga Yellowknife Supergroup. Most of these appear to be associated with trans-tensional basins that developed late in the history of the craton. The most intensively documented succession is the Beaulieu River (or Rapids) Formation (Roscoe et al. 1989; Rice et al. 1990; Corcoran 1996; Corcoran et al. 1999; Corcoran and Muller 2002). Detailed observations by Corcoran (1996) and Corcoran et al. (1999) indicated that matrix-supported conglomerates and associated sandstones at the base of the succession were deposited by debris flows, traction currents, and sheetfloods, on proximal to distal parts of alluvial fans or fan-deltas (type A rivers). Overlying sandstones were deposited in a sandy braidplain (type L) with minor overbank deposits, with siltstone-dominated intervals representing shallow lacustrine deposits. Overlying contact framework conglomerates represent a second phase of basin margin deposition from fluvial fans (type B), overlain by a further sequence of sandy braided river deposits (type L). Stratigraphically equivalent strata in the Jackson Lake (Henderson 1975; Muller et al 2002) and Keskarrah formations (Corcoran et al. 1998; Henderson 1998; Corcoran and Muller 2002) may include some fan or braid delta deposits, but are predominantly of tide influenced, shallow marine origin.

In the Superior Province (Fig. 10) fluvial strata have been associated with deposition in forearc, interarc, foreland, and piggyback basins, developed during consolidation of the craton, and with later elongate basins associated disruption of the craton by strike-slip activity, especially along terrane boundaries (Devaney and Williams 1989; Muller and Corcoran 1998; Corcoran and Muller 2004 b, 2007).

<Fig 10 here>

Conglomerates are a conspicuous component of many greenstone belts in the Superior Province: most are of submarine fan, or pro-fan-delta origin, although some are of fluvial, or undetermined origin. For example the Visean Greenstone belt in northern Ungava contains a thin succession of highly metamorphosed clast-supported conglomerate, dated at <2.708 Ga (Percival et al. 1993). These overlie a regolith developed on an irregular surface (implying a fluvial origin), yet interbedded sandstones are typically massive, favoring a subaqueous origin. A second example of uncertain origin occurs in the 2.712-2.699 Ga Island Lake Group (Lin et al. 2013), in the Sachigo sub province (Fig. 10). Clast-supported conglomerates, interbedded with graded and cross-stratified sandstone, might be of fluvial origin, but as the underlying unit of graded-bedded sandstones and mudstones is clearly of turbidic origin, a fluvial interpretation of these, and overlying massive and cross-stratified sandstone is suspect. Medium to thick-bedded conglomerates in the <2.706 Ga Oxford Lake Group (Syme et al. 1998; Lin et al. 2006) may also be of subaqueous rather than subaerial origin.

Fluvial and marine strata have been identified in the Cross Lake Greenstone belt in the Sachigo subprovince (Fig. 10). Strata of the 2.744-2.728 Ga Gunpoint Group include a generally fining upwards succession, including clast-supported conglomerate, sandstone and mudstone, interbedded with felsic volcanic strata (Corkery et al. 1992). At the base of the succession thick-bedded contact framework conglomerates have been interpreted as alluvial fan deposits (type B): these are overlain by trough cross-stratified sandstones interbedded with contact framework conglomerates, representing deposition from gravelly sandy braided rivers (type C or L). Fluvial strata are capped by a 1-3 m thick rhyodacite flow, then a 100 m thick unit of thick-bedded breccia that could represent tallus, debris flow dominated fans, or subaqueous fan deposits. The upper part of the Group consists of wavy and flaser bedded sandstones and siltstones of marine origin (Corkery et al 1992). The overlying 2.713-2.695 Ga Cross Lake Group begins with a succession of thick bedded clast-supported conglomerates (type B?) overlain by thick bedded, cross-stratified clast-supported conglomerates, cross-stratified matrix-supported conglomerates and trough and planar cross-stratified sandstones (type L), with interbedded pyroclastic flows (Corkery et al. 1992). Overlying siltstone, graded siltstone and mudstone may be of lacustrine or shallow marine origin. Along the southern margin of the Sachigo belt, fluvial strata have been recognized in younger strata within the Favourable Lake and North Spirit Lake Greenstone belts (Corfu et al. 1988). Highly tectonised conglomerates near the base of the 2.75-2.73 Ga Makataiamik Group (Wood 1980; Corfu et al. 1998) are interbedded with planar laminated sandstones, passing up section into pebble and cobble conglomerates and with interbeds of trough cross-stratified sandstone. The conglomerates may reflect deposition on alluvial fans, as suggested by Wood (1980), or may reflect progradation of fan-deltas, with the lower half representing pro-delta slope deposits, and the upper half representing gravel-bed rivers (type A) deposited on the delta topset. The conglomerates are overlain by granular medium to fine trough cross-bedded sandstones of more distal sandy braided origin (type L). Overlying strata, including local carbonate and mudrocks are of marine origin (Wood 1980). Alluvial (and submarine) fan deposits have also been recorded in the northern part of the belt in the 2.725 Ma North Trout Lake assemblage (Thurston et al. 1991), and (possibly) associated with volcanic strata in the ~ 2.7 Ga Bijou Point complex (Corfu et al. 1988).

In the Uchi belt Devaney (1999 a, 2001 a) interpreted highly deformed pebble to boulder grade conglomerate, interbedded with massive and planar-laminated sandstone, with minor mudstone intraclasts in the Woman Bay assemblage (Confederation Lake Group), were of fluvial origin. He suggested that they may have been deposited as proximal fluvial facies (type B), on basin margin fans in an intra-arc (hinterland) or backarc basin, associated with an Andean type collision at ~2.74 Ga. A similar setting has been suggested for the San Antonio Formation in the Rice Lake Greenstone belt, where some highly deformed contact-framework conglomerates, and cross-stratified sandstones have been interpreted as alluvial fan (type A or B) and sandy braided river deposits (type L), as well as subaqueous pro-deltaic deposits (Weber 1971; Anderson 2004, 2013). Local clast-supported boulder conglomerates along the southern margin of the basin may represent (subaqueous or subaerial) debris-flow deposits or tallus (Andersen 2013). Equivalent strata in the <2.7 Ga Flanders Lake Formation, and in the <2.72 Ga Hole River basin, may include some fluvial strata (type A), but are predominantly marine (Posehn 1976; Gilbert 2005; Bailes and Percival 2005; Percival et al. 2006). Two thin fluvial intervals have been reported in the 2.713-2.702 Eagle Island Group. This unit, previously considered as being entirely of deep-water submarine fan origin, was reinterpreted by Fralick and Pufahl (2006) as a repetitive shoaling sequence, in which deep-water mudrocks and graded sandstones give way to trough cross-stratified sandstones of shallow marine origin, overlain by a conglomerate sandstone sequence of beach origin, interbedded with fluvial influenced mouth-bar deposits, and a 7 m thick sequence of gravel-bed braided river deposits (type B) reflecting progradation of a wave influenced delta.

The Uchi subprovince also contains a number of elongate, strike-slip (pull-apart) basins, typically refered to as “Timiskaming type” basins (Muller and Corcoran 1998), that developed between 2.715 and 2.720 Ga (Devaney 1999 a, b), after the craton stabilized. Clast-supported conglomerates, with minor interbeds of cross-bedded, ripple laminated, and planar laminated sandstone at Sundown Lake, include both fluvial (fan) and resedimented (pro-fan-delta) facies (Devaney 1999 a, 2001; Percival et al. 1999). Similar (bouldery) cobble conglomerates at Birch Lake have been interpreted as proximal gravel-bed river (fan) deposits (type B), while local pebble conglomerates may represent distal fluvial, lacustrine or even shallow marine facies (Devaney 1997). Similar strata have been recorded at Slattery Lake (Devaney 1999 b) and Bee Lake (Percival et al. 1999).

In the Wabigoon subprovince (Fig. 10), fluvial strata occur in both arc-related and later strike-slip related basins. Minor fluvial reworking has been identified on the flanks of locally emergent volcanoes in the 2.733 Ga Neepawa Group, near Sioux Lookout (Babin et al. 1996; Devaney 2000) intermixed with pyroclastic facies. Fluvial strata include rare clast-supported polymict conglomerate (with rounded clasts), and planar laminated sandstone with minor pebble bands, and rare cross-stratification. A high gradient, valley confined straight gravel-bed river (or fan) setting is indicated by the presence of local (internal) stratification in the conglomerate matrix, analogous to sieve textures developed by infiltration of fines between clasts on alluvial fans (Long 1974).

The 2.703-2.696 Ga Stormy Lake Basin is dominated by arc-related volcanic rocks, but does include extensive massive cobble and boulder grade contact-framework conglomerates of probable fluvial (type B?) origin (Muller and Corcoran 1998; Dorstal et al. 2004). Thin interbeds of planar to low-angle cross-stratified, and trough cross-stratified sandstone may indicate upper and lower flow-regime falling stage deposition within fluvial channels. Mudstone laminae could have accumulated in abandoned channels, or in a floodplain, or lacustrine setting (c.f. Fig. 9 in Muller and Corcoran 1998).

The 2.699-2.694 Ga Crowduck Lake Group, includes coarse clastic strata that have been tentatively interpreted as fan-related, fluvial strata that accumulated in an interarc rift by Ayer and Davis (1997), although the description in Davies (1965) suggested that most of the group may have been deposited in a subaqueous, submarine fan setting. Corcoran and Muller (2007) interpreted local breccia at the base of the succession as possible tallus deposits, with associated siltstone indicating deposition in a shallow lake. They also suggested that a conglomerate, that fills a 10 m deep erosional feature higher in the group, may be of fluvial origin (type B?).

Framework-supported conglomerate, and cross-stratified sandstone, in the lower part of the 2.698-2.695 Ga Amet Bay Formation, near Sioux Lookout (Pettijohn 1934; Turner and Walker 1973; Devaney 2000), appear to have accumulated in a linear piggyback basin, that was overthrust by older rocks. Devaney (2000) interpreted the conglomeratic strata as proximal gravel-bed river deposits (types B and C), with thicker sandy intervals representing more distal fluvial facies (type L). Rare sigmoidal crossbeds may indicate intervals of higher energy ephemeral flows (type N). Equivalent strata in the

Savant-Sturgeon Greenstone belt (Sandbourn-Barrie et al. 2006; Fralick and Pufahl 2006) may include some similar fluvial strata.

Devaney (2000) suggested that clast-supported (basalt-clast) conglomerates in the (undated) Patra Formation accumulated in a 20 km long syn-orogenic pull-apart basin that predates the later "Timiskaming type" basins in the belt. Devaney (2000) suggests coarser facies were deposited in fan-related proximal gravel-bed rivers (type B, C), that prograded into marine or lacustrine strata. Sandstone intervals, characterized by low-angle cross-stratification, sigmoidal bedding, overturned bedding, and dessication cracks were considered to reflect deposition from high-energy ephemeral sandy braided rivers (type N).

In the northern metasedimentary belt of the Beardmore-Geraldtown Greenstone belt, Devaney and Fralick (1985) and Devaney and Williams (1989) recognized a thick succession of crudely bedded cobble conglomerates interbedded with minor lenses of massive, planar laminated and rarely planar cross-stratified pebbly medium to fine grained sandstone in the <2.692 Ga Namewaminikan Group. They interpreted these as Scott (type B) and Donjeck (type C) river deposits that accumulated on alluvial fans in a forearc basin (Fralick and Pufhal 2006). Distal facies are exposed to the south in the central metasedimentary belt, where they wedge out into pro-deltaic and turbidite facies (Devaney and Fralick 1985; Devaney and Williams 1989; Barrett and Fralick 1989). Thick cross-stratified sandstone units in this area were interpreted as more distal sandy fluvial facies (type L).

Fluvial conglomerates have also been recognized in a restricted (Timiskaming type) basin at the contact of the Wabigoon and Quetico subprovinces. The 2.693-2.686

Ga Seine Group includes a large-scale fining-upward succession, with (deformed) conglomerate-dominated facies at the base, and sandstone dominated facies at the top (Wood 1980). Czeck and Fralick (2002) interpreted pebble and cobble conglomerates near the base of the succession as Scott type river deposits (type B), with interbeds of medium grained trough cross-stratified sandstones representing infill of chute channels. Up section the proportion of sandstone increases, with stacked sequences of trough, and minor planar cross-stratification reflecting deposition in South Saskatchewan type rivers (type M).

In the Wawa subprovince, fluvial strata have been identified in a number of late, strike-slip related “Timiskaming type” basins. Jirsa (2000) identified possible deposits of debris flow dominated (type A) and fluvial dominated (type B) fan systems in the Midway Sequence at the western end of the subprovince. These mixed conglomeratic and volcanic rocks overlie a deep-water mud-rich turbidite succession, and could represent subaqueous resedimented conglomeratic facies, rater than fluvial strata. Evidence for a fluvial origin appears to be based on the presence of subaerial volcanic flows and associated pyroclastics in the “volcanic facies” near the base of the succession.

Within the ~2.690 Ga Shebandowan assemblage volcanic rocks interfinger with conglomerate, sandstone and minor mudrock that Shegelski (1980) considers to be analogous to “Timiskaming type” basins in the Abitibi. Monomict conglomerates, associated with extrusive shoshonites may have accumulated in both subaerial and submarine settings, and are locally associated with tidal flat, and storm influenced shallow marine strata (Fralick, pers. com. 2018). Polymict conglomerates (~2.688 Ga) are typically massive and poorly sorted, and appear to have been deposited as debris flows (Shegelski 1980). Discontinuous layers of massive, laminated and cross-stratified sandstone occur as lenses between some of the conglomerates, as do laminated mudstone units, with ripple drift cross-lamination and rare dessication structures. In the eastern part of the belt sandstones are characterized by planar lamination, large-scale cross-stratification (with mud-flakes distributed on foresets), and ripple cross lamination. This is suggestive of deposition on debris-dominated fans (type A) that pass downstream into sheetflood dominated sandy fluvial systems (type O).

At the eastern end of the subprovince, conglomeratic fan-delta, pro-delta and turbidite deposits have been documented in the Doré Group (Rice and Donaldson 1991 a, b; Corfu and Sage 1992). Rice and Donaldson (1991 a) interpreted most strata within the Doré Group in terms of submarine fan and shallow shelf facies, They considered that some highly deformed conglomerates in the northern part of the Michipicoten belt may be fluvial, but as these lack primary sedimentary structures, this could not be confirmed. In the southern belt they tentatively identified a thin sequence (~10 m) of sandstone with large-scale trough cross-stratification, as being fluvial, but conceded that they could also represent a subaqueous fan facies. Wendland et al. (2012) identified a 454 m thick succession of diamond bearing conglomerates (Leadbetter conglomerate), which they interpreted in terms of deposition on an arid region fan (type A). Matrix-supported cobble to boulder conglomerates, which characterize much of the lower part of the succession, occur in beds 0.8-4 m thick, interbedded with clast-supported conglomerates and trough cross-stratified sandstones. In places the clast-supported conglomerates are overlain by laminated sandstone. The upper part of the succession consists of interbedded clast-supported, massive, planar bedded and (minor) trough cross-stratified conglomerate, interbedded with trough, and (minor ripple), cross-stratified, and planar laminated sandstone, consistent with deposition in Scott type (type B) river systems that were strongly influenced by torrential floods (Wendland et al. 2012). Overlying sandstones indicate rapid marine transition and a return to deeper water conditions.

Muller and Donaldson (1992) recognized (minor) fluvial facies in 2.715-2.705 Ga strata in the northern part of the Abitibi Belt, and in younger 2.685-2.675 Ga "Timiskaming type" basins in the south. Strata in the older, northern belt include the Stella and Hauy Formations (Opamiska Group), which accumulated in an elongate interarc basin. The early phase of deposition (Stella Formation) involved a system of coastal fans, characterized by proximal units of massive and planar cross-stratified contact framework cobble and boulder conglomerates, interbedded with minor (<10%) granular to coarse grained sandstones that Muller and Dimroth (1987) interpret as the product of Scott type rivers (type B), that prograded into, and are interbedded with, shallow marine and deep-water facies. Sand-rich units, with 10-50% massive matrix and clast-supported conglomerate, and trough and planar cross-stratified sandstones (with minor fine grained sandstone and siltstone) were interpreted as Donjeck type river deposits (type C). Fan deposits in the overlying (syn-volcanic) Hauy Formation are more sand-rich, and were also interpreted as deposits of Donjeck type rivers (type C).

In the southern part of the Abitibi belt numerous elongate 2.685-2.675 Ga "Timiskaming type" basins were developed adjacent to major orogen-parallel (strike-slip) faults. These include the Penhorwood, Matheson, La Bruére and Duparquet basins, adjacent to the Destor-Porcupine fault zone (Rocheleau 1980; Dimroth et al. 1982; Muller et al. 1991, 1994; Muller and Corcoran 1998; Berger 2002; Bleeker et al. 2015), and the Borden Lake, Shining Tree, Kirkland, Grenada and Rouyn-Noranda basins, adjacent to the Cadillac-Larder Lake fault zone (Hyde 1980; Muller and Donaldson 1992; Muller et al. 1994; Corcoran and Muller 2007; Lin et al. 2006, 2013; Diop 2011). Most of these basins contain a mixture of fluvial and shallow to deep marine or lacustrine strata, as well as extrusive igneous and volcaniclastic facies. Fluvial strata are typically confined to fans or fan-deltas which prograded from the basin margins. They include both type A and B gravel-bed rivers, with associated sandstones deposited in more distal shallow perennial (type L) or sheetflood dominated systems (type O) (c.f. Muller et al. 1991, 1994; Muller and Corcoran 1998; Diop 2011).

Several small, fault bound "Timiskaming type" basins have also been recognized in the La Grande River subprovince of northern Quebec (Fig. 10). The fill of the Brune, Magin, and Keyano basins include conglomerates of fluvial origin that have been both dynamically deformed and metamorphosed to amphibolite grade (Duparc 2014; Duparc et al. 2016). Despite this, Duparc (2014) was able to identify deposits of Scott (B) and Donjeck (type C) rivers. Further west, Roscoe and Donaldson (1988) identified a 1.7 km thick succession of sandstones at Lac Sakami that are capped by an interval of uraniferous and auriferous quartz pebble conglomerate, overlain in turn by a succession of calc-silicate rocks, iron formation and basalt. In places the (deformed) conglomerates are seen to infill channels, and are associated with trough cross-stratified sandstones, implying deposition in shallow gravel-bed rivers (type B) in a distal (axial) setting. Closer to James Bay, 2.716-2.733 Ga diamond bearing conglomerates have been discovered at Ekomiak lake, but have yet to be described in detail (Cloutier 2008).

In the southern part of Wyoming province, the 2.67-2.69 Ga Deep Gulch conglomerate (Souders and Frost 2006), at the base of the Phantom Lake metamorphic suite in the Sierra Madre Mountains, has been interpreted as a fan-related braided fluvial system by Kratochvil (1981), Houston and Karlstrom (1987), and Huston et al. (1992). Kratochvil (1981) describes the basal 5m of the formation as a massive, poorly sorted, arkosic, small pebble conglomerate. Above this the formation is dominated by massive and trough cross-stratified sandstones, with thin lenticular pebble lags up to 4 cm thick, that near the base of the formation can be traced for 1-40 m along strike. A 20 m thick conglomeratic sequence in the upper half of the formation (unit 3 of Kratochvil 1981) contains uraniferous, pyritiferous small pebble conglomerates in sets 2 to 70 cm thick, interbedded with trough cross-stratified medium to very coarse sandstones. Although Kratochvil (1981) and others, have suggested deposition on a wet-alluvial fan, deposition in shallow gravel-bed (type B) and shallow gravelly-sandy braided rivers (type L) appears more likely. Stratigraphically equivalent metaconglomeratic strata of the Stud Creek and Rock Mountain formations, in the Medicine Bow Mountains in the southeastern part of the Wyoming Craton, may be of similar braided stream and alluvial fan origin (Houston et al. 1992) but have not been described in sufficient detail to evaluate fluvial affinity.

2.4.5. India

While sandstones, interbedded with highly deformed metaconglomerates, at the base of the Vinivillas Formation, on the Dharwar Craton of southern India, were considered to be glaciofluvial by Pichamunthu (1935). Subsequent studies by Ojakangas et al. (2014) have demonstrated that the strata that form the ~2.7 Ga Talya and Kaldurga conglomerates are entirely of glaciomarine origin. Van Loon and De (2015, 2017) and Van Loon et al. (2012) have identified a series of conglomerates in the Singburn Craton of northwest India that they identified in terms of both meandering and unique (Van Loon and De 2017) river models. The Rajkharsawan conglomerate occurs as isolated lenses of pebble conglomerate near (or below) the base of the 2.6-2.1 Ga Dahnjori Formation, that have been extensively stretched parallel to strike. After correction for a 5:1 stretching ratio, the documented channels would have been between 13.5 and 55.1 m wide, and 2.1 to 7.9 m thick. These are encapsulated in schist, with no primary sedimentary structures. The overall geometry of the system is more consistent with deposition on a small submarine fan, than a distal fluvial plain with meandering gravel-bed rivers. This is supported by the presence of matrix-supported conglomerates of presumed debris flow origin in the Bara Kumabra, Chhota Kumabera, Dhitar Dari and Bisrampur conglomerates documented by Van Loon and De (2015).

2.4.6. Eurasia

Neoarchean fluvial strata appear to be absent in the Siberian Craton: this may reflect poor exposure, or dominance of marine clastic strata. In the Baltic Shield Slabunov et al. (2006) and Hölttä et al. (2008) reported ~2.68 Ga conglomerates in terrigenous strata in the Kolmozerop-Voron'ya Terrane of the Kola Province of northern Baltica (Fig. 9), but provide no detailed descriptions. Saverikko (1998) interpreted aphobolite grade sandstones and conglomerates in northern Finland, in the Oraniemi suite of the 2.79-2.70 Ga Lapponian Supergroup, as predominantly sandy braided river deposits (type L). At the base of this succession Saverikko (1998) reported minor gravelly braided river deposits (type B) that accumulated on a fluvial fan, adjacent to a rift (aulacogen) margin. Much of this succession is dominated by coarse-grained trough cross-stratified sandstone, in sets 5-25 cm thick, separated by micaceous partings. The sandstone succession is overlain by a thick succession of interbedded mudstone and sandstone of deep-water origin. Although a fluvial interpretation is possible for the underlying clastics, detailed architectural and paleocurrent observations are needed to confirm that the Oraniemi suite is not at least in part or whole of shallow marine origin.

Epiclastic and syn-volcanic cobble conglomerates were identified by Peltonen et al. (1988) in the 2.526 Ga Rookkiiaapa Formation in northern Finland. They interpreted them as Scott type (type B) braided stream deposits that accumulated on the lower flanks of an andesitic volcano. The fluvial conglomerates include sheets of both framework, and (minor) clast-supported cobble conglomerate, with rounded clasts. Interbedded with these are tuff layers, thin beds of planar and ripple laminated volcaniclastic sandstone, and minor volcanic breccias that may have been emplaced as lahars or tallus. Similar lahars have been recorded in a late "Timiskaming type" basin in central Finland, in the <2.75 Ga Juuikkajärvi conglomerate that Lehntonen (2016) describes as fluvial or shallow marine.

2.4.7. South America

In the Rio das Velhas Greenstone of the São Francisco Craton 2.75-2.67 Ga volcaniclastic, momomict and polymict breccias in the Palmital Formation have been tentatively interpreted as including both alluvial fan and submarine fan deposits, that accumulated in both a forearc and extensional backarc setting (Baltazar and Zucchetti 2007; Moreira et al. 2016; Farina et al. 2016). Association with (minor) planar and trough cross-stratified, and local sigmoidal crossbedded, volcaniclastic sandstones suggests emergent parts of the fan may have been characterized by ephemeral flash foods (type O and P rivers). Possible fluvial strata of similar age has also been recorded in the upper part of the ~2.75 Ga Agus Clarus Formation in the southeastern part of the Amazonian Craton, where they overlie platformal mudstone, and shallow marine to inter-tidal sandstone (Nogueira et al. 1995).

The Casa Forte Formation overlies the Rio das Velhas Greenstone belt, and is dominated by clast-supported conglomerates, interbedded with coarse and medium grained sandstone, including planar bedded and trough trough cross-bedded varieties, deposited in a foreland basin setting. Baltazar and Zuccchetti (2007) reported channel scour and fill structures up to 0.5 m thick. They interpreted these strata as products of deposition from gravel-bed rivers (type B) in a mid-fan setting. Associated 0.5 to 1.5 m thick fining upwards cycles, consisting of planar bedded and laminated granule bearing sandstone, and small to medium-scale planar and trough cross-stratification, caped by thin layers of silt or clay, were interpreted as products of unconfined distal braided rivers. The abundance of planar lamination, and the mudstone caps suggest these represent ephemeral sandy systems (type O or P).

Stratigraphically equivalent strata in the <2.68-2.62 Ga Moeda Formation consists of three cycles of clastic strata, beginning with auriferous conglomerate, passing up section into quartz arenite (Koglin et al. 2014). Minter et al. (1990) suggested that a thin diamictite (< 1m) at the base of the succession may be of debris flow origin. Cobble and large pebble conglomerates above the diamictite appear to have been deposited in bedrock confined shallow gravel-bed rivers (type B). These pass up section into trough cross-stratified quartz arenites (type L rivers), and then into massive fine grained sandstones of shallow marine origin. These are succeeded by a thick succession of trough cross-stratified sandstones, with minor pebble lags, consistent with deposition in shallow sand-bed braided rivers (type L).

Although clastic strata in the Jacobina Group contain only 3.5-3.2 Ga Paleoarchean zircon populations (Teles et al. 2015), most authors consider this sequence to be of Paleoproterozoic age, associated with foreland basin development during the Tranzamazonian orogeny at ~2.0 Ga (Ledru et al. 1997; Teixeira et al. 2001; Farina et al. 2016). In contrast Teles et al. (2015) considers the pyritic conglomerates at he base of the group (Serra do Córrego Formation) to indicate that these are pre-oxyatmoversion, and so could be Neoarchean or early Paleoproterozoic. Scarpelli (1991) and Teles et al. (2015) considered the contact-framework conglomerates and associated planar and trough cross-stratified sandstones to be deposits of alluvial fans (type B) and shallow sandy braided rivers (type L). Scarpelli (1991) argued that, as the main conglomerate channels are at right angles to strike, the extent of the conglomerate lenses is indicated by strike length, giving potential channel dimensions for the basal conglomerate as 300 m wide and 4 m deep, suggesting that at least locally, deep gravel-bed rivers were involved (type C). Given the high lateral continuity of conglomerates in this belt (26 km at the base, 35 km at the top) this formation may represent a prograding distributary fluvial system (c.f. Davidson et al. 2013).

3. Discussion

In Archean strata there are clear problems in identifying depositional style, especially where primary structures are not preserved as a result of shearing or metamorphism. Even where "clasts" are recognized in highly strained conglomerates, in many cases there is no clear evidence whether the deposits are of fluvial, beach or deep-sea origin, hence interpretation must be based on stratigraphic and geotectonic context, perhaps augmented by geochemical analysis. The increased preservation potential of fluvial deposits after 3.2 Ga may reflect a profound change in plate tectonic style on the Archean Earth. Maruyama et al. (2018) suggest that plate tectonics may have been initiated in the middle Hadean (4.37-4.2 Ga) as a result of heavy bombardment by meteorites. No records survive from this period, and during the succeeding Eoarchean and Paleoarchean plate tectonic processes were probably active on a more limited scale than in modern systems and were initially dominated by thick volcanic plateau developed above mantle plumes (Van Kranendonk 2007 b; Hickman and Van Kranendonk 2012). Dhuime et al. (2015, 2017) suggest that 65% of the Earth's continental crust was generated prior to 3 Ga. They suggest that the onset of subduction driven plate tectonics at this time resulted in a decreased rate of generation of new continental crust, with increased crustal thickening triggering greater rates of emergence, erosion and weathering. The Pilbara Craton contains a near continuous record of Paleoarchean crustal development, beginning with development of a series of volcanic plateau, and collapse basins, above mantle plumes between 3.53 and 3.23 Ga (Nijman et al. 2017), followed by rifting and reassembly by modern style plate tectonic processes (Hickman and Van Kranendonk 2012). A similar association is seen in the eastern Kaapvaall Craton, where strata of the Noisy Complex (Furnes et al. 2013), Mapepe Formation (Nocita and Lowe 1990), and Witkop Formation (Wilson and Versfeld 1994) overlie thick sequences of komatiite and theolitic basalt.

Although fluvial systems would be expected to form as a common feature on the flanks of emergent volcanoes, these have only a very limited preservation potential. In addition, diagenetic and metamorphic processes make identification of reworked pyroclastic and epicalstic material difficult to distinguish from the parent volcanic strata, and tallus. The dominance of volcanic plateau, and paucity of extensive areas of felsic continental crust (Bradley 2011), combined with globally high sea levels (Eriksson et al. 2013; Flament et al. 2013), may also have contributed to the scarcity of fluvial strata in the Paleoarchean. Metasediments at Mount Narryer in the Yilgarn Craton may reflect local development of extensional basins at 3.3-3.2 Ga, or may alternatively reflect local preservation within a foreland basin setting (Eriksson et al. 1988).

Modern style plate tectonic processes can be traced back to the beginning of the Mesoarchean (Van Kranendonk 2007 b; Hickman and Van Kranendonk 2012, and references therein), resulting in the development of cratonic nuclei. Condie (2018) indicates a transition between 3 and 2 Ga from lid tectonics to plate tectonics. By the Neoarchean many of the major cratonic areas had stabilized. Corcoran (2007) suggests that Neoarchean fluvial deposits accumulated both in forearc and compressional foreland basins during the terminal phase of orogenic collisions, and in later strike-slip basins (especially along terrane boundaries).

4. Conclusions

Based on the Miall (1999) classification of river types (Fig. 1), shallow gravel-bed rivers (type B) form a third (33%) of all documented Archean fluvial strata (Fig. 11, top). Gravel-bed rivers with debris flows (type A), and shallow perennial sand-bed rivers are next (18% each), with sheetflood dominated sandy systems forming 12% of known deposits. Deep gravel-bed rivers make up 8% of the record, with other varieties making up the last 10% (Fig. 11). No meandering systems could be confirmed. Secular trends (Fig. 11, bottom) indicate a slight reduction in type A systems with time, and a corresponding increase in flashy ephemeral systems (N, O, P) in younger strata.

<Fig. 11 here>

Although it is evident that fluvial systems must have been active from very early in Earth's history, the planiform architecture of these systems is not always clear from the geological record, even if the deposits can be matched with existing models of vertical successions (c.f. Miall 1996). Planiform architecture has been demonstrated successfully using plans of conglomerate thickness and/or gold grade in strata of economic interest (Fig. 6). In many studies of Archean fluvial deposits interpretation is based entirely on the presence (or absence) of matrix- and framework-conglomerate, or presence of cross-stratified or planar laminated sandstone. Very little has been done to document the three-dimensional architecture of, or distribution of, architectural elements in Archean fluvial strata. Exceptions include the studies by Karpeta (1993) and Krapež (1985). Karpeta (1993) documented channel fill, lateral accretion, and downstream accretion elements in the Witpan conglomerates of the Kimberley Formation. Krapež (1985) documented the geometry of two channel fill sequences in the Venterdorp Contact Reef. There is considerable room for future research into the nature of Archean fluvial systems. Furture studies need to incorporate an architectureal element approach, with detailed measurements of depositional surface directions and dip (Miall 1999; Long 2011) in order to resolve the problem of three-dimensional geometry of the alluvial bodies. This may be facilitated using LiDAR, or structure-from-motion techniques, where sufficient lateral exposure is evident (c.f. Carrivic et al. 2016).

While several authors have interpreted Archean strata in terms of deposition from meandering systems, none have provided clear proof of sustained inclined heterolithic stratification, or systematic upsection changes in paleoflow directions. Where (local) mudstone lamina are preserved on sandstone foresets, this may be a reflection of backwater effects in tide-influenced situations (as in some distal Witwatersrand conglomerates), rather than providing evidence of meandering. Mudstone intervals in general are rare in the Archean fluvial deposits recorded in this study. McMahon and Davies (2018) record an average of 1 % (maximum = 14%) for a more limited selection of Archean fluvial deposits. One percent may be a high estimate, as it includes thicker mudstone intervals that might better be interpreted as lake or pond deposits. Overbank deposits are uncommon, but have been recorded in several Archean fluvial systems. Detailed observations of pedogenic textures are needed to confirm overbank, rather than in-channel deposition of silt and mud grade material, yet these may be very difficult to detect in highly metamorphosed strata. The apparent absence of fluvial strata on several cratons is enigmatic (Figs. 2, 4, 9). This may be a function of high sea levels, flooding craton interiors (c.f. Eriksson et al. 2006), or prior erosion of terrestrial strata. Alternatively, fluvial deposits may have simply been obscured by extensive overburden, or metamorphic processes.

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List of Figures

Figure 1. Principal river types represented in the Archean rock record, based on models proposed by Miall (1996). Examples of Miall's river types D (Gravel-bed wandering), E (Gravel-bed meandering), H (Ephemeral sandy meandering), I (Fine grained meandering), J (Anabranching-anastomeosed), K (Low sinuosity braided with alternate bars), have yet to be recognized in the Archean record (based on Long 2011).

Figure 2. Distribution of known (stars) and suspect (triangles) fluvial strata prior to 3.2 Ga. Current distribution of Archean cratons in red, after Tang et al. (2013).

Figure 3. Left: Paleoarchean stratigraphy of the Pilbara Craton (Hickman and Van Kranendonk 2012), with location of possible fluvial intervals, including the Dresser

Formation (Djokic et al. 2017), Duffer Formation (Barley 1993), Panorama Formation (Retallack 2018) and Strelley Pool Formation (Van Kranendonk 2007 a, 2011).

Figure 4. Distribution of known (stars) and suspect (triangles) fluvial strata of Mesoarchean age. Distribution of Archean cratons in red after Tang et al. (2013).

Figure 5. Mesoarchean stratigraphy of the Kaapvaal Craton. Stratigraphy of the Dominion Group from Rantzsch et al. (2011); Witwatersrand Supergroup from Frimmel et al. (2005), and Guy et al. (2014) (Note: many formation names change across the basin); Swaziland Supergroup (Barberton Greenstone belt) from Hubeck and Lowe (1994), and Hubeck et al. (2013); Nsuze Group of the Pongola Supergroup from Cole (1994); Mozzan Group from Young et al. (1998). Ages from sources in the text. Insert map from Eglington and Armstrong (2004): Basement granite and gneisses in pink; Greenstone belts in green; Major sedimentary basins in yellow; Bushveld complex in purple; Cover rocks in grey.

Figure 6. Planiform architecture of auriferous conglomerate horizons modeled on different stratal characteristics. A: Reconstruction of the Main Reef (base of Main Fm.) in the Central Rand Goldfield, based on channel depth (thickness), measured in bore-holes (Stewart et al. 2004). B: Reconstruction of the Main Reef, further to the west in the Careltonville/West Rand Goldfield) based on inferred depositional facies (Tucker et al. 2016). C: Reconstruction of the Elsburg Composite Reef (Elsburg Fm) in the Klerksdorp Goldfied, based on relative gold content (Tucker et al. 2016, after Vilonen).

Figure 7. Representative sections through Witwatersrand Reefs. A: W-E section through the B placer, Kimberley Fm., in the Welkom Goldfield (Van Den Heveen 2008). B: Composite Reef, Main Fm., in West Rand Goldfield (Stewart et al. 2004). Lower two matrix-supported conglomerates (massive large and small pebble sandy gravels) may be lateral equivalents to the Main reef further to the west (c.f. Fig. 6 A), The overlying sandstone unit with bimodal-opposed cross-stratification is interpreted as a marine facies of the Main Fm. Sandstone. These are overlain on topographic highs by thin lenses of marine slity mudstone (zMm) of the "Black Bar"(equivalent to the "green bar in the Carletonville Goldfield). Overlying small pebble conglomerate is considered equivalent to the Main Leader Reef, and may reflect marine reworking of fluvial strata during marine transgression. The thin lenses of granule to small pebble conglomerate in the overlying sandstone may be equivalent to the South Reef. C: Transverse sections through proximal and (D) distal channels of the Steyen placer system in the Krugersdorp Fm (after Minter 1978).

Figure 8. Reconstruction of an Aandenk channel at the base of the Kimberley Formation in the Welkom Goldfield (after Bailey 1991).

Figure 9. Global distribution of known Neoarchean fluvial Strata.

Figure 10. Neoarchean fluvial strata in the Superior Province, Canada (Base map after Beakhouse et al. 1995). Yellow stars represent "Timiskaming type" basins.

Figure 11. Relative abundance of Archean river types (Letter codes as Fig. 1). Top, as number of each type identified; bottom as percentage of each fluvial style, by era.

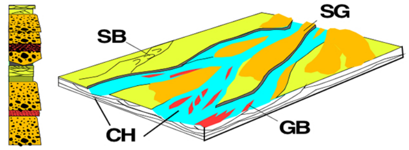
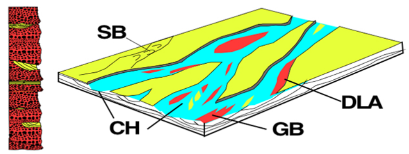
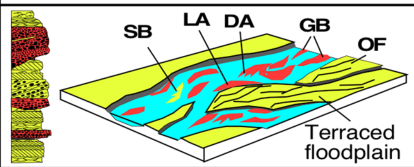
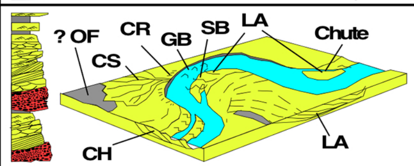
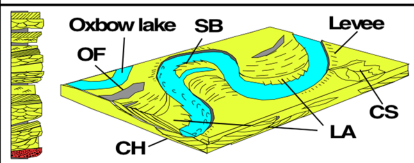
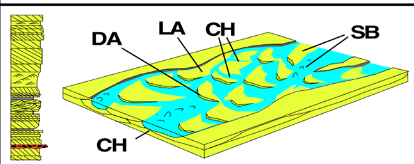
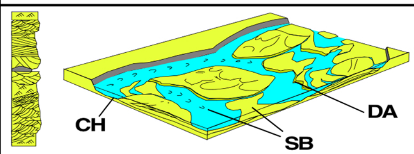
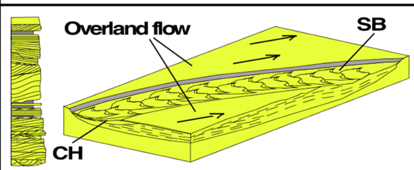
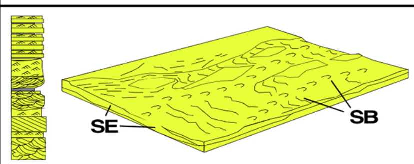
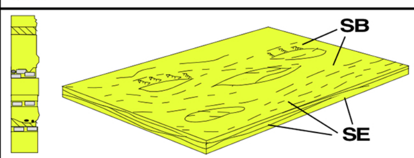
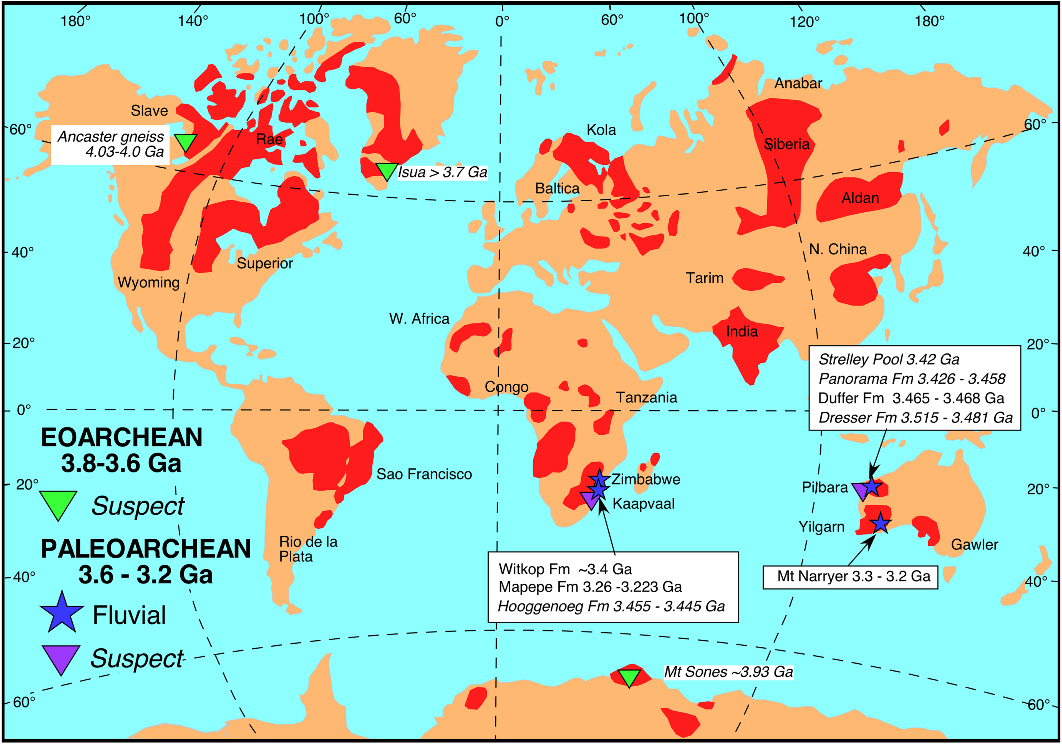
RIVER TYPES (Miall 1996)		Archean Examples
A Gravel-bed rivers (+ fans) with sediment gravity flows		**Paleoarchean**: Duffer, Mapepe, Witkop Fms; Mt Narryer. **Mesoarchean:** George Creek, Moodies Gps; Lalla Rookh, Mosquito Ck., Wagita Fms; basal Mahagiri quartzite. **Neoarchean:** Venterdorp contact reef; Mount Roe, Ongers R., Omdraavlei, Sekororo, Omdraavlei, Bothaville, Hardey, Mohle, Klippan, Pote, Schelem, San Antonio, Black Reef Fms +8 more
B Shallow gravel-bed braided rivers		**Paleoarchean**: Mt Narryer sequence; ? Witkop Fm. **Mesoarchean:** Jack Hills Dominion, Moodies Gps; Lalla Rookh, Mosquito Ck., Elandslaagte, Blyvooruitzch,Main, Monde, Uityk. **Neoarchean:** Mergougil Cg.; Jones Creek, Yandal, Boothaville Hampton, Sekororo, Tumbiana, Schelem, San Antonio, Scotty Ck., Black Reef, Manjeri, Beaulieu R. Fms., + 29 more
C Deep (>3m) gravel-bed rivers		**Paleoarchean**: Mt Narryer sequence. **Mesoarchean:** Mosquito Creek, Promise, Witpan mbr, Kimberley, Deny-Dalton mbr, Agatha, Sinqeni Fms. L Bababudan (cong) **Neoarchean:** Amet Bay, Patra, Stella, Hauy, Serra do Córrego Fms.; Gunpoint, Namewaminikan Gps.
F Gravelly-sandy meandering rivers		No proven examples
G Sandy meandering rivers		No proven examples
L Shallow perennial braided		**Paleoarchean**: Mt Narryer sequence. **Mesoarchean:** Lalla Rookh, Mosquito Ck., Mantonga, Conley, Elandslaagte, Bonanza, Reitkuil, Blyvooruitzch. Main, Sinqeni Fms; Uitsig mbr., Kimberley Fm; Bababudan Gp; Mahagiri Qtz. **Neoarchean:** Mt Roe, Yandal, Sekororo, Black Reef, Ongers R. Beaulieu R., Moeda, Amet Bay, Stud Ck., Deep Gulch Fms; Shamvaian, Makataiamik, Gunpoint, Cross Lake Gps.+ 2 more
M Deep (>3m) perennial braided rivers		**Mesoarchean:** Lalla Rookh, Mantonga, Promise Fms. **Neoarchean:** Hardey Fm?, Scotty Creek Fm., Seine Gp.
N High-energy ephemeral sandy braided rivers		**Neoarchean:** ? Middle arenite mbr, Hampton Fm; Amet Bay, Patra Fms.
O Sheet-flood dominated distal braided rivers		**Mesoarchean**: Mosquito Creek Fm. **Neoarchean:** Palmital Fm; Casa Forte Fm; **Neoarchean:** Merougil Cong., Navajo sst, Jones Creek, Yandal, Schelem, ? Boothaville, Sekororo, Scotty Ck., Mohle, Klippan, Pote. Ongers River, Casa Forte, Palmital Fms.; Timmins, Duparquet basins; Shamvaian Gp. Shebandowan assemblage.
P Flashy sheet-flood dominated rivers		**Mesoarchean**: Mosquito Creek, Elandslaagte Fms.; Jack Hills Gp. **Neoarchean:** Palmital, Casa Forte, Mt Roe, Ongers River Fms. ? Merougil Cong.;

Figure 1



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Slave
Ancaster gneiss
4.03-4.0 Ga
Rae
Isua > 3.7 Ga
Kola
Anabar
Siberia
Baltica
Aldan
Superior
Wyoming
N. China
Tarim
W. Africa
India
Congo
Tanzania
Strelley Pool 3.42 Ga
Panorama Fm 3.426 - 3.458
Duffer Fm 3.465 - 3.468 Ga
Dresser Fm 3.515 - 3.481 Ga
EOARCHEAN
3.8-3.6 Ga
Suspect
Sao Francisco
Zimbabwe
Kaapvaal
Pilbara
Yilgarn
Gawler
PALEOARCHEAN
3.6 - 3.2 Ga
Fluvial
Suspect
Rio de la
Plata
Witkop Fm ~3.4 Ga
Mapepe Fm 3.26 -3.223 Ga
Hooggenoeg Fm 3.455 - 3.445 Ga
Mt Narryer 3.3 - 3.2 Ga
Mt Sones ~3.93 Ga

Figure 2

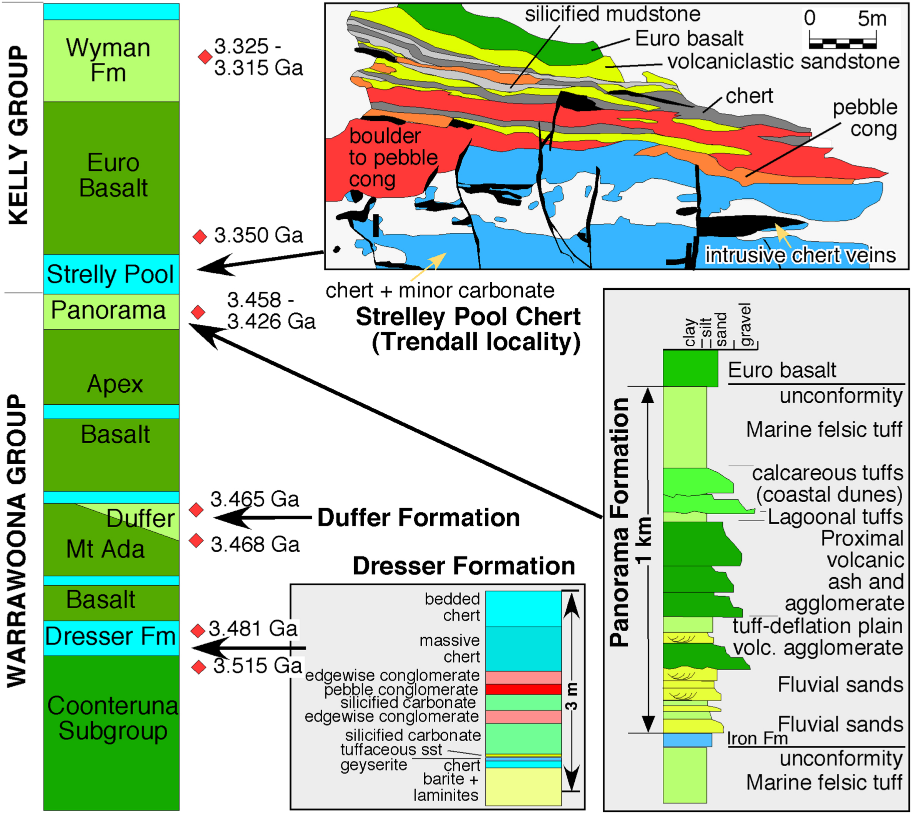
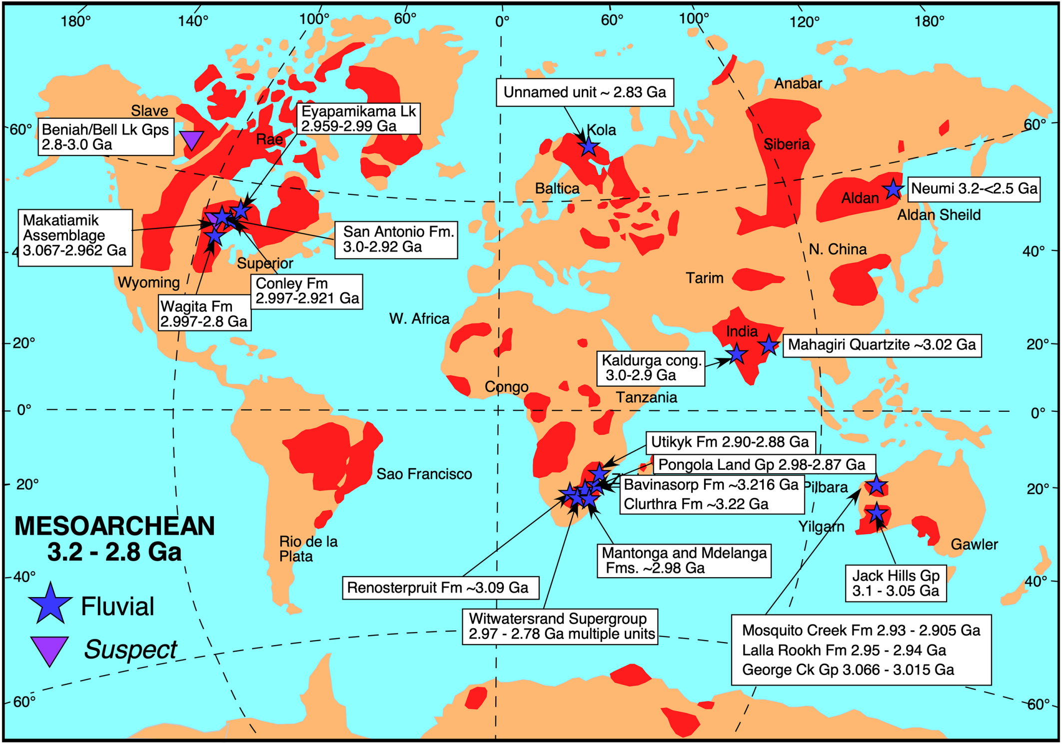

KELLY GROUP
WARRAWOONA GROUP
Wyman Fm
Euro Basalt
Strelly Pool
Panorama
Apex
Basalt
Duffer
Mt Ada
Basalt
Dresser Fm
Coonteruna Subgroup
3.325 - 3.315 Ga
3.350 Ga
3.458 - 3.426 Ga
3.465 Ga
3.468 Ga
3.481 Ga
3.515 Ga
silicified mudstone
Euro basalt
volcaniclastic sandstone
0 5m
chert
pebble cong
boulder to pebble cong
intrusive chert veins
chert + minor carbonate
Strelley Pool Chert (Trendall locality)
Duffer Formation
Dresser Formation
bedded chert
massive chert
edgewise conglomerate
pebble conglomerate
silicified carbonate
edgewise conglomerate
silicified carbonate
tuffaceous sst
geyserite
chert
barite + laminites
3 m
Panorama Formation
1 km
clay
silt
sand
gravel
Euro basalt
unconformity
Marine felsic tuff
calcareous tuffs (coastal dunes)
Lagoonal tuffs
Proximal volcanic ash and agglomerate
tuff-deflation plain
volc. agglomerate
Fluvial sands
Fluvial sands
Iron Fm
unconformity
Marine felsic tuff

Figure 3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60°
Slave
Beniah/Bell Lk Gps
2.8-3.0 Ga
Rae
Eyapamikama Lk
2.959-2.99 Ga
Makatiamik
Assemblage
3.067-2.962 Ga
San Antonio Fm.
3.0-2.92 Ga
Superior
Wyoming
Conley Fm
2.997-2.921 Ga
Wagita Fm
2.997-2.8 Ga
Unnamed unit ~ 2.83 Ga
Kola
Baltica
Anabar
Siberia
Aldan
Neumi 3.2-<2.5 Ga
Aldan Sheild
N. China
Tarim
India
Mahagiri Quartzite ~3.02 Ga
Kaldurga cong.
3.0-2.9 Ga
W. Africa
Congo
Tanzania
Sao Francisco
Rio de la
Plata
Utikyk Fm 2.90-2.88 Ga
Pongola Land Gp 2.98-2.87 Ga
Bavinasorp Fm ~3.216 Ga
Clurthra Fm ~3.22 Ga
Pilbara
Yilgarn
Gawler
Mantonga and Mdelanga
Fms. ~2.98 Ga
Jack Hills Gp
3.1 - 3.05 Ga
Renosterpruit Fm ~3.09 Ga
Witwatersrand Supergroup
2.97 - 2.78 Ga multiple units
Mosquito Creek Fm 2.93 - 2.905 Ga
Lalla Rookh Fm 2.95 - 2.94 Ga
George Ck Gp 3.066 - 3.015 Ga
MESOARCHEAN
3.2 - 2.8 Ga
Fluvial
Suspect

Figure 4

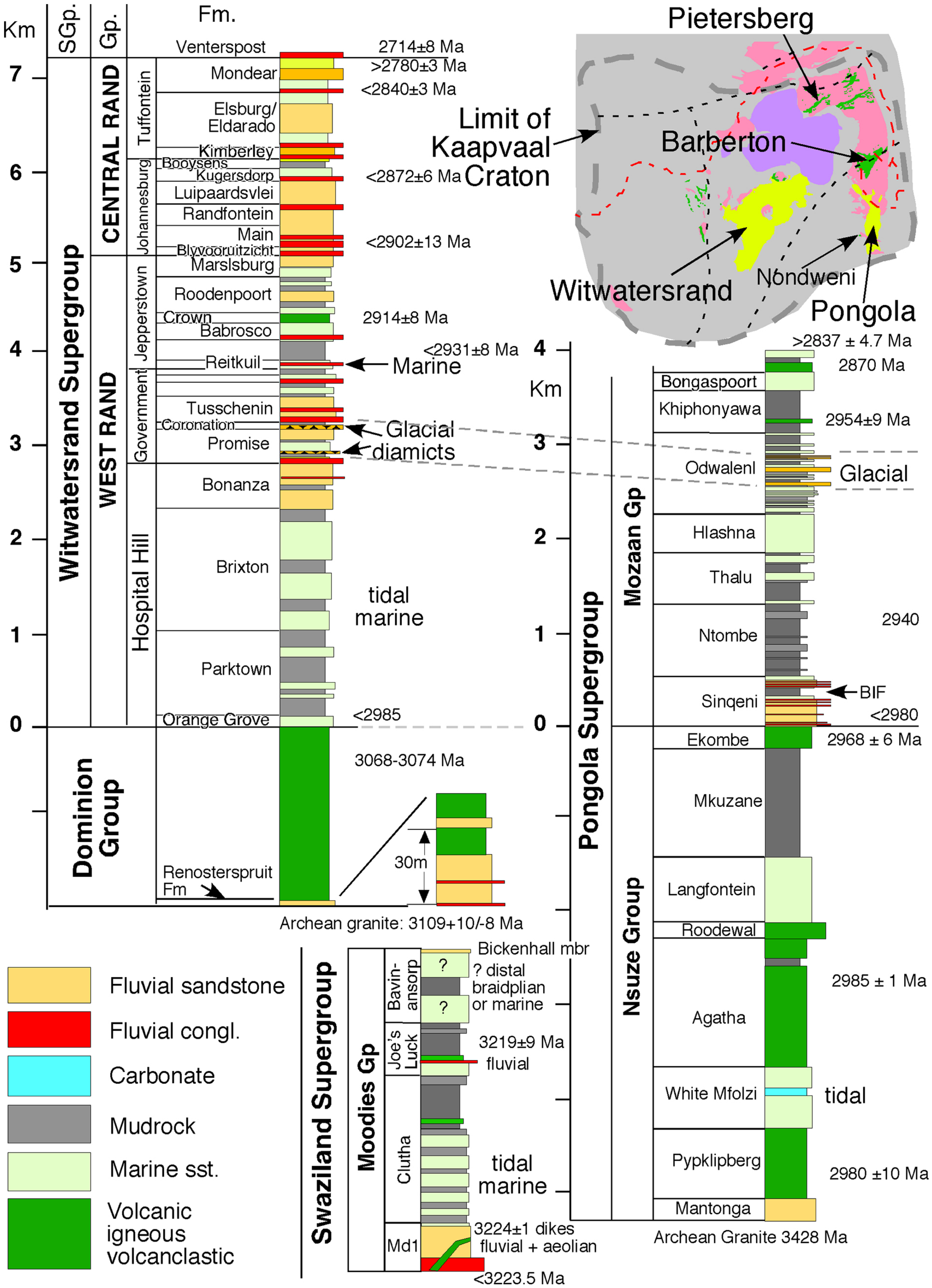

Km
SGp.
Gp.
Fm.
Venterspost
2714±8 Ma
Mondear
>2780±3 Ma
<2840±3 Ma
Elsburg/ Eldarado
Kimberley
Booysens
Kugersdorp
<2872±6 Ma
Luipaardsvlei
Randfontein
Main
Blyvooruitzicht
<2902±13 Ma
Marslsburg
Roodenpoort
Crown
2914±8 Ma
Babrosco
<2931±8 Ma
Reitkuil
Marine
Tusschenin
Coronation
Glacial diamicts
Promise
Bonanza
Brixton
tidal marine
Parktown
Orange Grove
<2985
3068-3074 Ma
30m
Renosterspruit Fm
Archean granite: 3109+10/-8 Ma
Tuffontein
Johannesburg
Jepperstown
Government
Hospital Hill
CENTRAL RAND
WEST RAND
Witwatersrand Supergroup
Dominion Group
Pietersberg
Limit of Kaapvaal Craton
Barberton
Witwatersrand
Nondweni
Pongola
>2837 ± 4.7 Ma
2870 Ma
Bongaspoort
Khiphonyawa
2954±9 Ma
Odwalenl
Glacial
Hlashna
Thalu
2940
Ntombe
Sinqeni
BIF
<2980
Ekombe
2968 ± 6 Ma
Mkuzane
Langfontein
Roodewal
Agatha
2985 ± 1 Ma
White Mfolzi
tidal
Pypklipberg
2980 ±10 Ma
Mantonga
Archean Granite 3428 Ma
Mozaan Gp
Nsuze Group
Pongola Supergroup
Fluvial sandstone
Fluvial congl.
Carbonate
Mudrock
Marine sst.
Volcanic igneous volcaniclastic
Swaziland Supergroup
Moodies Gp
Bavin-ansorp
Joe's Luck
Clutha
Md1
Bickenhall mbr
? distal braidplian or marine
3219±9 Ma
fluvial
tidal marine
3224±1 dikes
fluvial + aeolian
<3223.5 Ma

Figure 5

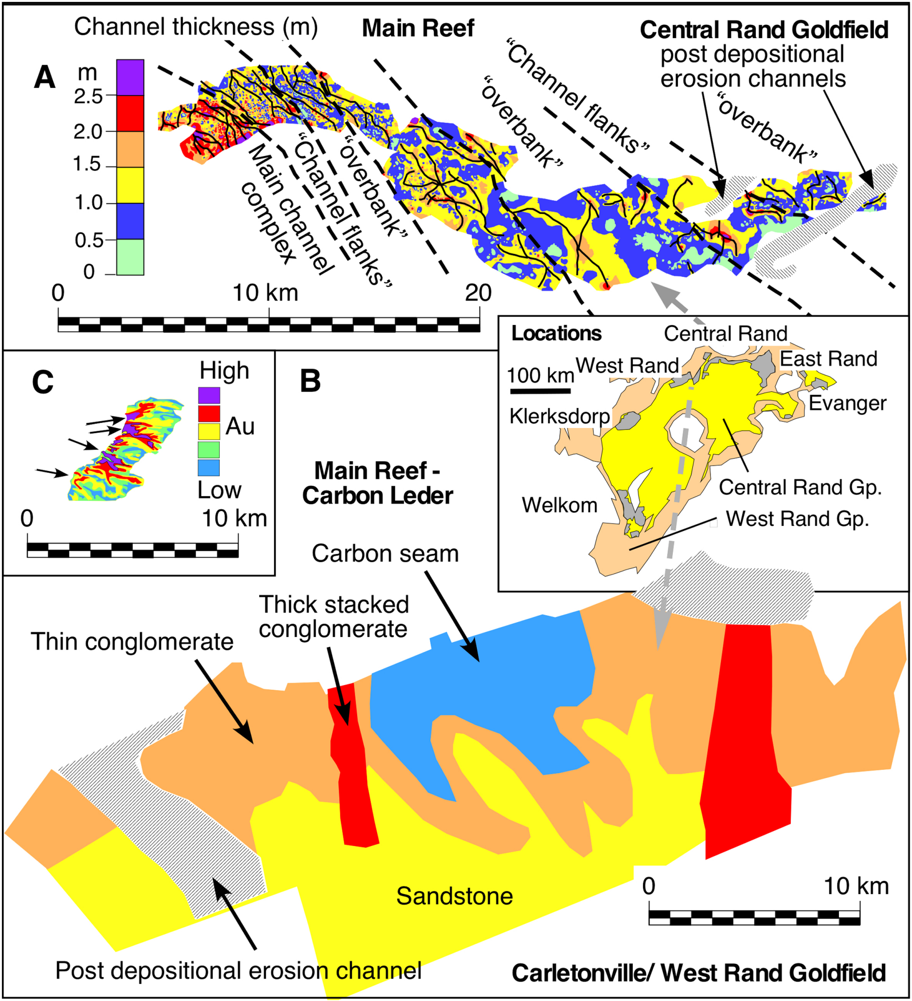
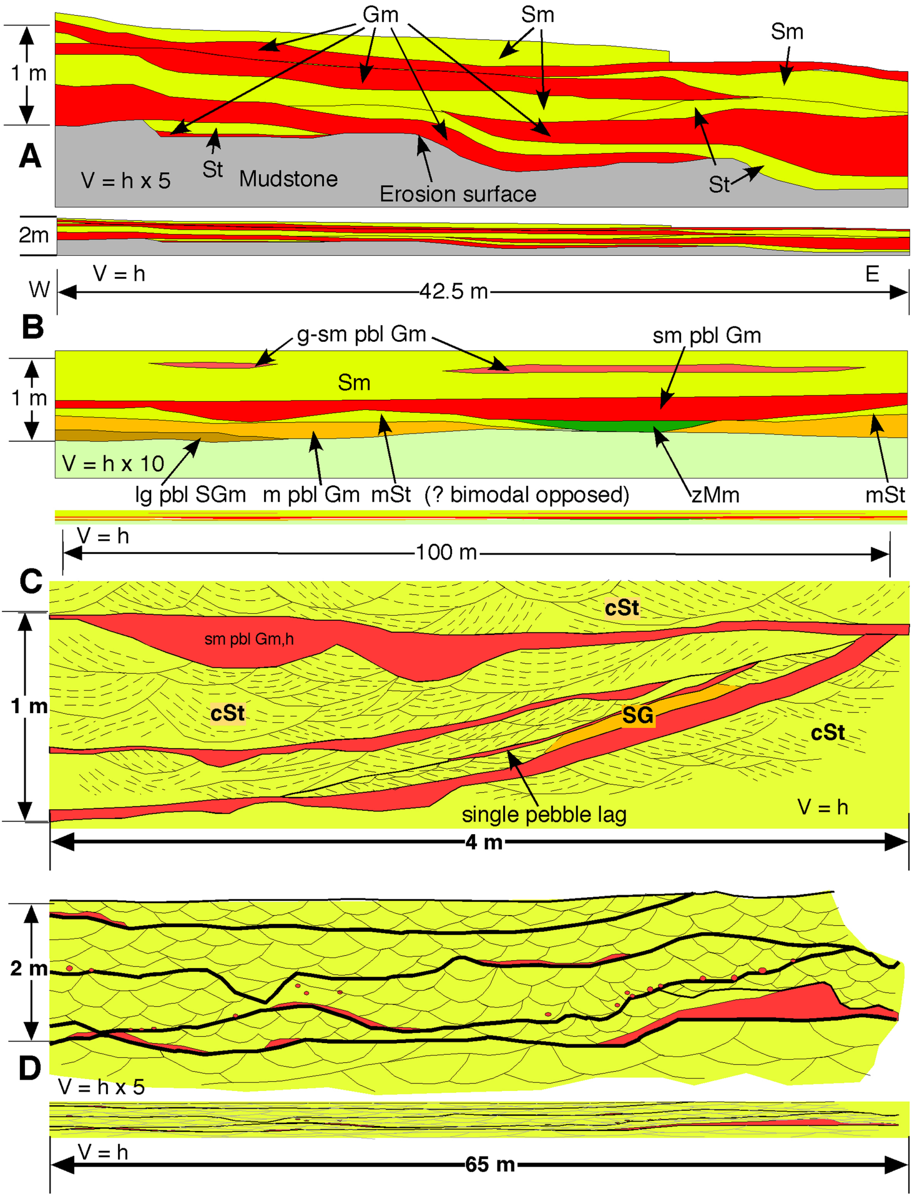
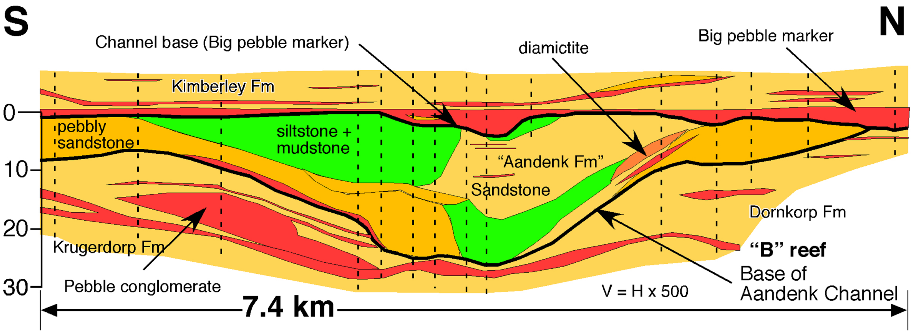

A
Channel thickness (m)
m
2.5
2.0
1.5
1.0
0.5
0
Main Reef
Central Rand Goldfield
post depositional
erosion channels
"Channel flanks"
"overbank"
"overbank"
"Channel flanks"
"overbank"
Main channel
complex
0
10 km
20
Locations
100 km
Central Rand
West Rand
East Rand
Evanger
Klerksdorp
Central Rand Gp.
Welkom
West Rand Gp.
C
High
Au
Low
0
10 km
B
Main Reef -
Carbon Leder
Carbon seam
Thick stacked
conglomerate
Thin conglomerate
Sandstone
0
10 km
Post depositional erosion channel
Carletonville/ West Rand Goldfield

Figure 6



A
Gm
Sm
Sm
1 m
St
Mudstone
V = h x 5
Erosion surface
St
2m
V = h
W
E
42.5 m
B
g-sm pbl Gm
sm pbl Gm
Sm
1 m
V = h x 10
lg pbl SGm
m pbl Gm
mSt
(? bimodal opposed)
zMm
mSt
V = h
100 m
C
cSt
sm pbl Gm,h
1 m
cSt
SG
cSt
single pebble lag
V = h
4 m
2 m
D
V = h x 5
V = h
65 m

Figure 7



S
N
Channel base (Big pebble marker)
diamictite
Big pebble marker
Kimberley Fm
0
10
20
30
pebbly sandstone
siltstone + mudstone
"Aandenk Fm"
Sandstone
Dornkorp Fm
Krugerdorp Fm
"B" reef
Pebble conglomerate
V = H x 500
Base of Aandenk Channel
7.4 km

Figure 8

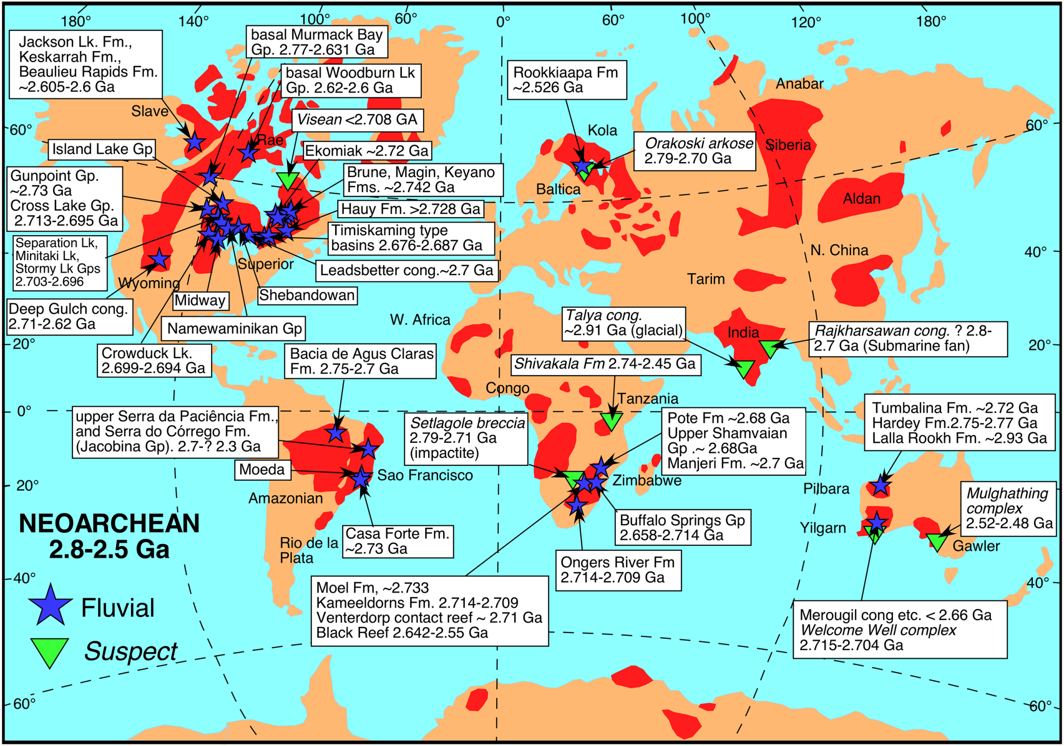
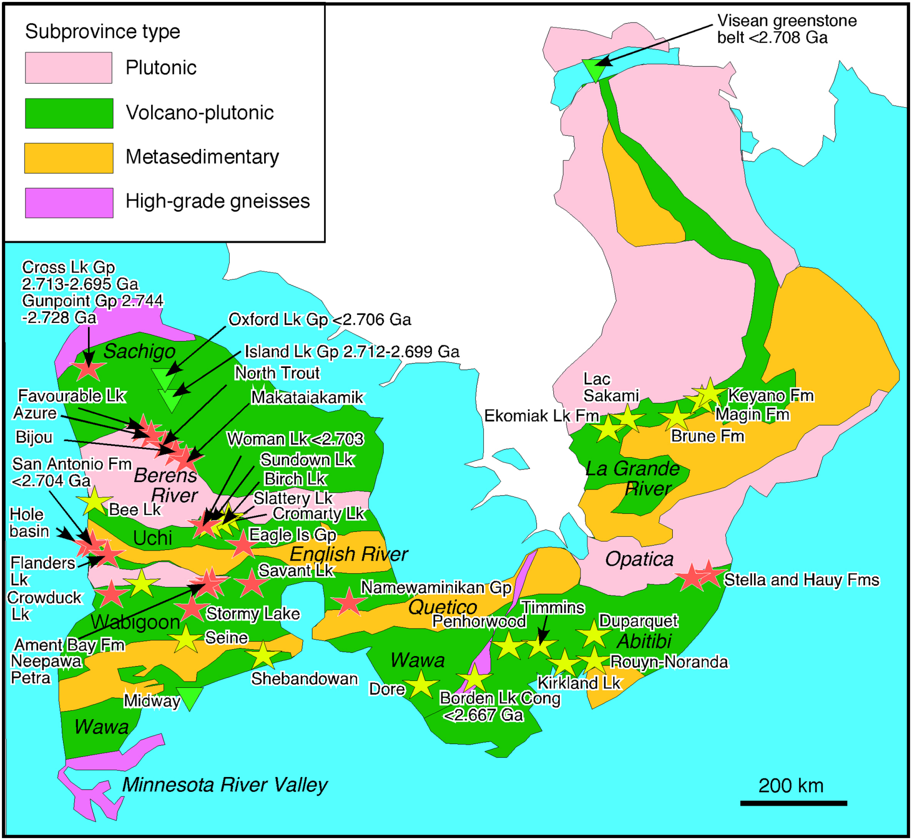
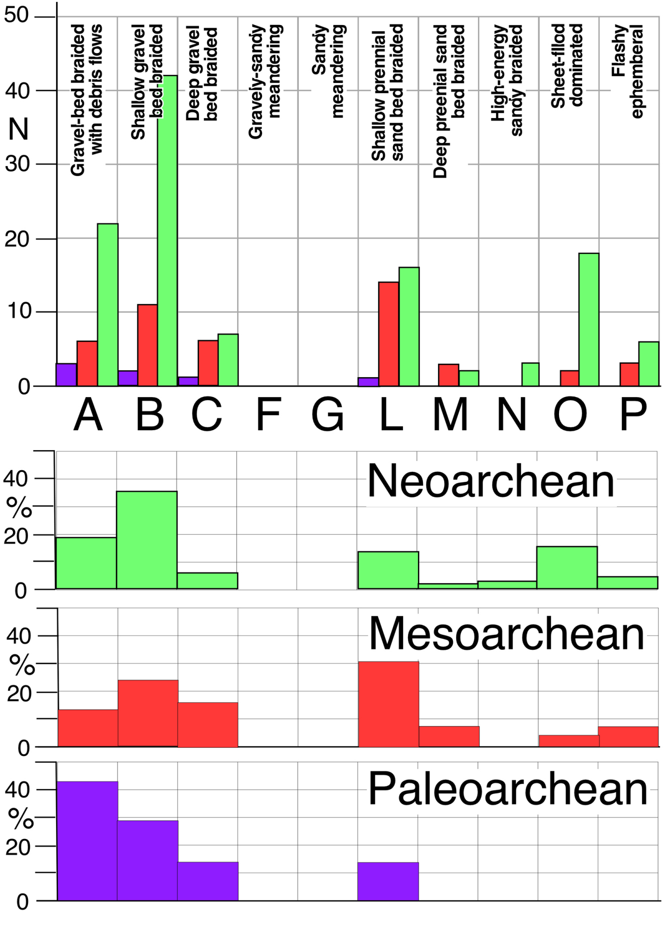

180°
140°
100°
60°
0°
60°
100°
120°
180°
60°
40°
20°
0°
20°
40°
60°
Jackson Lk. Fm., Keskarrah Fm., Beaulieu Rapids Fm. ~2.605-2.6 Ga
basal Murmack Bay Gp. 2.77-2.631 Ga
basal Woodburn Lk Gp. 2.62-2.6 Ga
Visean <2.708 GA
Ekomiak ~2.72 Ga
Brune, Magin, Keyano Fms. ~2.742 Ga
Hauy Fm. >2.728 Ga
Timiskaming type basins 2.676-2.687 Ga
Leadsbetter cong.~2.7 Ga
Shebandowan
Namewaminikan Gp
Midway
Crowduck Lk. 2.699-2.694 Ga
Deep Gulch cong. 2.71-2.62 Ga
Separation Lk, Minitaki Lk, Stormy Lk Gps 2.703-2.696
Gunpoint Gp. ~2.73 Ga Cross Lake Gp. 2.713-2.695 Ga
Island Lake Gp
Slave
Rae
Superior
Wyoming
Rookkiaapa Fm ~2.526 Ga
Kola
Orakoski arkose 2.79-2.70 Ga
Baltica
Anabar
Siberia
Aldan
N. China
Tarim
India
W. Africa
Talya cong. ~2.91 Ga (glacial)
Rajkharsawan cong. ? 2.8-2.7 Ga (Submarine fan)
Shivakala Fm 2.74-2.45 Ga
Congo
Tanzania
Bacia de Agus Claras Fm. 2.75-2.7 Ga
upper Serra da Paciência Fm., and Serra do Córrego Fm. (Jacobina Gp). 2.7-? 2.3 Ga
Moeda
Sao Francisco
Amazonian
Rio de la Plata
Casa Forte Fm. ~2.73 Ga
Setlagole breccia 2.79-2.71 Ga (impactite)
Pote Fm ~2.68 Ga Upper Shamvaian Gp .~ 2.68Ga Manjeri Fm. ~2.7 Ga
Zimbabwe
Buffalo Springs Gp 2.658-2.714 Ga
Ongers River Fm 2.714-2.709 Ga
Moel Fm, ~2.733 Kameeldorns Fm. 2.714-2.709 Venterdorp contact reef ~ 2.71 Ga Black Reef 2.642-2.55 Ga
Tumbalina Fm. ~2.72 Ga Hardey Fm.2.75-2.77 Ga Lalla Rookh Fm. ~2.93 Ga
Pilbara
Yilgarn
Mulghathing complex 2.52-2.48 Ga
Gawler
Merougil cong etc. < 2.66 Ga Welcome Well complex 2.715-2.704 Ga
NEOARCHEAN 2.8-2.5 Ga
Fluvial
Suspect

Figure 9



Subprovince type
Plutonic
Volcano-plutonic
Metasedimentary
High-grade gneisses
Visean greenstone belt <2.708 Ga
Cross Lk Gp 2.713-2.695 Ga Gunpoint Gp 2.744 -2.728 Ga
Sachigo
Oxford Lk Gp <2.706 Ga
Island Lk Gp 2.712-2.699 Ga
North Trout
Makataiakamik
Favourable Lk
Azure
Bijou
Woman Lk <2.703
Sundown Lk
Birch Lk
Slattery Lk
Cromarty Lk
San Antonio Fm <2.704 Ga
Berens River
Bee Lk
Hole basin
Uchi
Eagle Is Gp
English River
Flanders Lk
Savant Lk
Crowduck Lk
Namewaminikan Gp
Quetico
Stormy Lake
Wabigoon
Seine
Ament Bay Fm
Neepawa
Petra
Shebandowan
Midway
Wawa
Minnesota River Valley
Lac Sakami
Ekomiak Lk Fm
Keyano Fm
Magin Fm
Brune Fm
La Grande River
Opatica
Stella and Hauy Fms
Timmins
Penhorwood
Duparquet
Abitibi
Rouyn-Noranda
Kirkland Lk
Wawa
Dore
Borden Lk Cong <2.667 Ga
200 km

Figure 10



50
40
N
30
20
10
0
Gravel-bed braided with debris flows
Shallow gravel bed-braided
Deep gravel bed braided
Gravely-sandy meandering
Sandy meandering
Shallow prennial sand bed braided
Deep preenial sand bed braided
High-energy sandy braided
Sheet-fllod dominated
Flashy ephemberal
A B C F G L M N O P
Neoarchean
Mesoarchean
Paleoarchean
40
%
20
0

Figure 11