A Giant Mesoarchean Crustal Gold-Enrichment Episode: Possible Causes and Consequences for Exploration

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Abstract

Comparison of conglomerate-hosted, Witwatersrand-type gold deposits and/or occurrences worldwide reveals that this deposit type is by no means unique to the Kaapvaal craton but common to most Archean and/or Paleoproterozoic cratons. The age of the variably mineralized fluvial to fluvio-deltaic conglomerates ranges from 3.1 to 1.9 Ga. They were deposited in tectonic settings ranging from continental rifts to passive margins and synorogenic foreland basins, and all of them are paleoplacers. Although several of them show evidence of local mobilization of ore components by postdepositional hydrothermal fluids, purely epigenetic hydrothermal models fail to explain the geometry of the orebodies as well as available lithogeochemical, mineral chemical, and isotope data. Conglomerates older than 2.4 Ga are characterized by an abundance of detrital (and secondary) pyrite, and in most cases also detrital uraninite, whereas most of the younger examples (<2.2 Ga) contain Fe oxides instead. A common denominator of Witwatersrand-type deposits is the stratigraphic position above erosional unconformities adjacent to an Archean to Paleoproterozoic hinterland. The Witwatersrand deposits themselves differ from all other examples of this type by a gold endowment that is two to three orders of magnitude greater, an abundance of gold-rich “carbon” seams that reflect former microbial mats, a scarcity of gold nuggets, and orders of magnitude higher Os contents in the gold.

For the Witwatersrand gold, a genetic model is proposed that involves the following requirements: (1) an anomalous mantle domain as the ultimate source, strongly enriched in siderophile elements, caused by inhomogeneous mixing with cosmic material that was added during intense meteorite bombardment of the Hadean to Paleoarchean Earth, plume-like ascent of relics from inefficient core formation, or plumes from the core-mantle boundary; (2) elevated gold extraction into juvenile crust when mantle temperature reached its maximum in the Mesoarchean; (3) several orders of magnitude higher run-off of gold from the Mesoarchean land surface due to intense weathering under an aggressive, reducing atmosphere and high gold solubility in coeval river water; (4) trapping of gold from river water on the surface of local photosynthesizing microbial (cyanobacterial) mats; and (5) reworking of these mats into erosion channels during flooding events (and by eolian deflation) and redeposition of gold as placer particles. Postdepositional hydrothermal and/or metamorphic overprints explain why much of the gold is now located in texturally late positions but had little significance on the macroscale distribution of the gold. Elsewhere in the world, a less fertile hinterland and/or less reworking of older sediments led to correspondingly lower gold endowment. Most of the Archean sedimentary rocks were affected by crustal reworking in the course of later tectonic overprints. The multitude of fluids and melts involved in these reworking processes gave rise to the great variety of gold deposit types known in post-Archean crustal sections.

The probability of discovering a new supergiant cluster of Witwatersrand-type deposits is considered very low. However, considerable potential exists for finding new smaller economic deposits of this type in Mesoarchean to Paleoproterozoic fluvial to fluvio-deltaic basal conglomerates, deposited especially in foreland basins next to Mesoarchean hinterland and/or auriferous sediment successions that could be reworked.

Introduction

Global gold production up to 2013 reached a total of approximately 164,600 metric tons (t), including an estimated 10,000 t of historic gold production prior to 1851 (Müller and Frimmel, 2011). Over 32,000 t of this, or 32%, have been extracted solely from the Witwatersrand goldfields in South Africa (based on production data from South African Chamber of Mines, with updates from RMG, Version 2014.01.14, © Intierra Raw Materials Group), where the gold occurs in Meso- and Neoarchean conglomerates. Even though the gold production from the Witwatersrand mines—the leading gold producers in the world for almost a century—has fallen constantly since 1970 from a peak of 1,000 t/annum to 170 t in 2012, the conglomerate beds in the Witwatersrand Super-group still host more than 44,000 t of Au, which is about 30% of global Au resources. This is more than any other deposit type can claim. Gold deposits of similar style, aptly referred to as Witwatersrand-type deposits, are known from all continents (except for Antarctica) but are dwarfed by the type ore province in South Africa both in terms of past production and known resources. Does this simply reflect differences in exploration maturity, with the Witwatersrand goldfields being the highly explored examples and other regions still offering great exploration potential? Or is there a specific process (or combination of processes) that led to a uniquely rich gold endowment in the Witwatersrand basin?

In spite of the enormous economic significance of Witwatersrand-type deposits, their origin has remained one of
the most hotly debated issues in economic geology. Little has changed since Davidson (1965) made a frequently quoted statement to that effect. As late as 2005, the Society of Economic Geologists selected the Witwatersrand gold deposits as the theme for a forum on "Controversies on the Origin of World-Class Gold Deposits" (Muntean et al., 2005). Subsequently the general opinion might have swayed toward a modified paleoplacer model (Frimmel et al., 2005a), although more recently, this was challenged by Horscroft et al. (2011), who advocated a microbially mediated synsedimentary model, and by Phillips and Powell (2011), who reiterated the hypothesis of postdepositional introduction of the gold into the host conglomerates by metamorphic fluids. One of the biggest uncertainties in the various genetic models is the adequacy of source of the estimated total of 97,000 t of gold within the Witwatersrand basin (including past production), be it for detrital gold particles from eroded rocks in the hinterland, or synsedimentary gold growth on microbial mats, or for Au-rich postdepositional fluids percolating through the entire Witwatersrand basin.

Not a single new Witwatersrand (-type) goldfield has been discovered in the past six decades—a failure that has been attributed to possibly flawed exploration models (Phillips, 2013). This highlights that a correct understanding of the genesis of Witwatersrand-type deposits is essential for any reasonable assessment of the potential of a given terrane for the discovery of new deposits of this type. The principle aim of this contribution is, therefore, to address the above questions pertaining to the possible uniqueness of the Witwatersrand goldfields and the potential of finding similar goldfields elsewhere. To that effect, the various Witwatersrand-type deposits in different continents are compared, the multitude of genetic models proposed for this deposit type evaluated, and the likely source of all the gold in these deposits speculated upon. Detailed descriptions of the many geological, mineralogical, geochemical, and geophysical features, especially of the deposits in the Witwatersrand basin, are not repeated here as these have been described in several review papers, such as those by Robb and Robb (1998), Phillips and Law (2000), Frimmel and Minter (2002), and Frimmel et al. (2005a).

**Witwatersrand-Type Gold Deposits and/or Occurrences in the Kaapvaal Craton**

**Witwatersrand Supergroup**

By far the economically most important ore province for Witwatersrand-type gold deposits is the Mesoarchean Witwatersrand basin in the center of the Kaapvaal craton of South Africa (Fig. 1). This structurally preserved basin appears elongated in a northeasterly direction with a maximum diameter...
of more than 400 km. The basin fill consists of up to 7 km of predominantly siliciclastic rocks of the Witwatersrand Supergroup, which rest upon a Paleo- to Mesothermal granitoid-greenstone basement with intercalated bimodal volcanic and coarse-grained siliciclastic, uraninite-rich rocks of the 3074 Ma Dominion Group. The supergroup has long been subdivided into the more shale-dominated shallow marine to distal fluvo-deltaic West Rand Group and the coarser-grained, mainly fluviol to fluviodeltaic Central Rand Group (Fig. 2A). The term “Witwatersrand basin” is in fact misleading because the West Rand and Central Rand groups represent two independent stratigraphic sequences that developed in entirely different basins, separated by a significant hiatus. Age constraints from detrital zircon and xenotime grains (Kositcin and Krapez, 2004), as well as a volcanic unit in the upper part of the West Rand Group (Armstrong et al., 1991), indicate periods of sedimentation between ca. 2985 and 2914 Ma for the West Rand Group and between ca. 2902 and 2840 Ma for the Central Rand Group. A passive margin setting with a paleoslope consistently toward the south and southwest has been suggested for the up to 5,150-m-thick sedimentary and minor volcanic rocks of the West Rand Group (Frimmel and Minter, 2002). Younger cover over most of the Witwatersrand Super group, ranging in age from Neoarchean to Mesozoic, obscures the subcrop extent of the Witwatersrand strata. Equivalents of the West Rand Group exist as the Mozaaan Group in the upper parts of the Pongola Supergroup (Beukes and Cairncross, 1991) some 300 km to the southeast of the known Witwatersrand basin (Fig. 1). Thus, the known remnants of the West Rand Group are only a small portion of what must have been a much larger basin.

Both the tectonic setting and basin architecture of the unconformably overlying Central Rand Group are very different from those of the West Rand Group. Maximum sediment thickness (2,880 m) is reached near the center of the present distribution of Central Rand Group rocks and paleoslope directions consistently change around the margin, pointing toward this center (Fig. 2B). Thus today’s known extent of Central Rand Group strata probably mimics relatively closely the extent of the original Central Rand basin. Much of the sedimentation took place syntectonically with respect to folding, faulting, and uplift along the basin margin, especially toward the west, northwest, and north, resulting in a large number of angular unconformities. Sedimentological as well as geochronological evidence from detrital zircon age spectra (Kositcin and Krapez, 2004) is consistent with a progressively shrinking foreland basin. For a more detailed overview of the geologic setting see Frimmel et al. (2005a). Synsedimentary unroofing of granitoids to the west is evident in the detritus within the upper Central Rand Group, from which a retroarc position of the Central Rand foreland basin has been deduced (Kositcin and Krapez, 2004; Koglin et al., 2010b).

Burial and basin dewatering beneath several kilometers of Neoarchean and Paleoproterozoic strata, syn and postdepositional crustal thickening along the western and northern basin margins, enhanced geothermal gradients during emplacement of the world’s largest layered intrusive complex, the 2054 Ma Bushveld Complex, and circulation of meteoric waters in the aftermath of the world’s largest known meteorite impact, the 2023 Ma Vredefort event, in the middle of the Witwatersrand basin all led to low-grade metamorphic overprint and several events of postdepositional hydrothermal alteration at various times during the first 800 m.y. after sedimentation. Maximum metamorphic temperatures did not exceed 350°C ± 50°C (Frimmel, 1994; Phillips and Law, 1994) except in the immediate vicinity of the Vredefort dome, where lower parts of the basin fill that had experienced medium- to high-grade thermal metamorphism during the Bushveld event were subsequently upturned during the Vredefort impact event (Gibson and Wallmach, 1995; Frimmel, 1997).

Gold concentrations at economic or subeconomic levels occur almost exclusively in coarse-grained siliciclastic rocks that reflect a variety of fluvial lithofacies. These comprise clast-supported oligomictic conglomerate, loosely packed conglomerate, pebbly arenite, and pebble lag surfaces associated with trough cross-bedded quartz arenite. Rarely, gold is associated with debris-flow lithofacies (Minter et al., 1988). The upper and lower contacts of the gold-bearing, predominantly conglomeratic beds (referred to as “reefs”) are sharp. Typically, Au contents across these sharp contacts by up to three orders of magnitude, with gold grades reaching as high as 25 g/t in the richest reefs but not exceeding 20 ppb in the footwall and hanging-wall rocks. Dispersion of gold into the immediate vicinity below and above the conglomeratic host is the exception and confined to distances of 1 to 2 m away from the reef, where gold contents of as much as 1 ppm are observed (Phillips, 1987).

A strong sedimentological control (Fig. 3A, B), that is the almost exclusive association of gold with coarse-grained, largely conglomeratic, siliciclastic rocks that are enriched in heavy minerals, has been used successfully for more than a century as a guide to the day-to-day exploitation of orebodies. Notably, crosscutting features, such as faults and veins, play no or only a very subordinate role as hosts of gold. The few exceptions are auriferous quartz veins or faults that are located within, or very close to, the mineralized conglomerate beds. On a basin-wide scale, not a single major crosscutting orbody has been discovered to date.

Gold has been mined from at least 30 reefs in both the West Rand Group and the Central Rand Group. Most of the typically lens- or sheet-like orebodies, which define fluvial bar and channel bedforms or mark eolian deflation surfaces, are located in the Central Rand Group which accounts for about 95% of total Witwatersrand production (Robb and Robb, 1998; Fig. 2A). All of the main goldfields are located along the margin of the former Central Rand basin at the mouths of complex river systems (Minter and Loen, 1991). Maximum gold grades are located in the channel facies but also on eolian deflation surfaces (Frimmel et al., 2005a).

Carbonaceous matter is abundant in the Witwatersrand metasedimentary rocks and occurs as both kerogen and solid (pyro-)bitumen (Mossman et al., 2008). Several workers have emphasized a strong spatial and, by inference, genetic relationship between gold and this carbonaceous matter. Nagy (1993) estimated that 40% of all Witwatersrand gold is closely associated with carbonaceous matter. Solid bitumen is sited within minute fractures in quartz pebbles (Parnell, 1999), as several millimeter-sized glassy globules within the conglomerates, or in conglomerate-hosted veins (Gartz and Frimmel, 1999), whereas kerogen forms stratiform millimeter-
**Fig. 2.** A. Generalized stratigraphic column for the Witwatersrand Supergroup: also shown are the positions of the main auriferous conglomerate beds (reefs) and their relative significance as gold producers (inset). B. Simplified surface and subsurface geologic map of the Witwatersrand basin, also showing the location of the goldfields and paleocurrent directions in the Central Rand Group; modified after Frimmel et al. (2005a).
centimeter-thick layers, referred to in the literature as "carbon seams." The biogenic nature of the carbonaceous matter is beyond doubt (Spangenberg and Frimmel, 2001), but different opinions have been voiced on its indigeneity, that is whether the "carbon" seams represent solidified bitumen derived from migrating oils (Gray et al., 1998; Drennan and Robb, 2006) or in situ former microbial mats (Hallbauer, 1975; Mossman et al., 2008). The latter authors provided an extensive list of observations that strongly support an indigeneous origin of the carbon seams, such as their truncation by paleoerosion channels, resedimented clasts of carbon seams, carbon-draped foreset beds in a (well-mineralized) planar crossbedded pebbly sandbar, presence of heavy mineral grains interstitially to columns of carbon, and a stromatolite-like columnar structure of the carbon seams. Figure 4A illustrates a thin film of kerogen draping dension cracks on the bottom contact of the Vaal Reef. Kerogen lining such delicate sedimentary structures is difficult to reconcile with introduction of hydrocarbons by postdepositional oils but adds to the numerous examples of carbon seams that occur on paleosurfaces, sedimentary accumulation, and ripple surfaces, all of which suggest their in situ formation as microbial mats.

Whereas there is no doubt that the local presence of carbon seams facilitated higher gold grades (Fig. 4B), the majority of the gold is independent of such carbon seams. The carbon seams are typical of low-energy depositional environments and deflation surfaces, but absent from proximal, high-energy deposits which can be excellent hosts to high-grade ore. A prime example that illustrates the spatial relationship between carbon seams, noncarbonaceous conglomerates, and gold grade is the Carbon Leader Reef in the Carletonville (also known as West Wits Line) goldfield, which has been the second largest gold producer among all Witwatersrand mines with a total production of more than 5,600 t gold and reported resources of about 6,300 t Au with an average grade of 11 g/t Au. The Carbon Leader sensu stricto, which is now largely mined out, was a thin (typically <10 cm in thickness), regionally extensive conglomerate bed with a well-developed kerogen and/or bitumen seam along its basal contact. At the Driefontein mine (now Kloof-Driefontein Complex), a facies change is exposed from the typical thin Carbon Leader conglomerate with the basal kerogen and/or bitumen seam in the west toward a series of different channel facies toward the east. In these channels the heavily mineralized carbon seams appear reworked but without any in situ kerogen and/or bitumen within (multiple) conglomeratic channel fills (Fig. 5; see also Mossmann et al., 2008, fig. 5).

Most of the gold is native gold that occurs in a range of textural positions: (1) as minute inclusions within pyrite, especially in rounded, concentrically laminated, porous pyrite and in secondary, euhedral pyrite overgrowths (Koglin et al., 2010a); (2) texturally associated with bitumen or chlorite; (3) as filamentous, more than 1-mm-long, free gold within carbon seams (Fig. 4B,C); and (4) as discrete micronuggets of typically torroidal to spheroidal shape (Fig. 3C) and longest diameters of around 0.1 mm, rarely as much as 1 mm (Hallbauer,
There is a distinct spatial association between gold, pyrite, and uraninite, whose origin has been a matter of debate, as well as a range of other undoubtedly detrital minerals, such as rounded zircon and chromite; for a complete list see Phillips and Law (2000). The composition of the gold particles varies widely, both between and within reefs (Oberthür and Saager, 1986; Hayward et al., 2005). For example, gold from the Vaal Reef has Ag and Hg contents ranging from 4 to 18 and 0.5 to 4.0 wt %, respectively (Reid et al., 1988). Most importantly, Witwatersrand gold is exceptionally rich in Re and Os. Rhenium and Os concentrations of 2.5 to 11.4 ppb and 4.1 ppb to 4.1 ppm, respectively, have been reported for the Vaal Reef (Kirk et al., 2002) and are 4 to 37 ppb and 2 to 15 ppb, respectively, in gold from the Basal reef (Frimmel et al., 2005a).

A range of different morphological pyrite types in the Witwatersrand ore have been recognized by all workers in the past but interpreted very differently. Whereas some have argued for a postsedimentary hydrothermal genesis of all pyrite (Barnicoat et al., 1997; Phillips and Law, 2000), others have recognized detrital, possibly synsedimentary and postdepositional types (Ramdohr, 1958; Hallbauer, 1986; England et al., 2002; Large et al., 2013). A multiple S and Fe isotope study on pyrite from Witwatersrand conglomerates (Hofmann et al., 2009) indicated unambiguously the detrital origin of the compact rounded pyrite grains. In a more recent, most comprehensive study on the texture and chemistry of pyrite in the nonconglomeratic rocks of the Witwatersrand Supergroup, Guy et al. (2010) distinguished a range of detrital, diagenetic, and epigenetic pyrite types and they established a connection between pyrite morphology and/or composition and depositional environment and tectonic setting. Among the conglomerate-hosted pyrite types, the synsedimentary, porous type yielded the highest Au contents, on average 3.5 ppm (Koglin et al., 2010). A strong increase in the proportion of detrital and syngenetic and/or early diagenetic pyrite marks the change from shallow marine to fluvial sedimentation. Synsedimentary
and/or early diagenetic concentrically laminated pyrite nodules are particularly abundant in the Central Rand Group. In contrast to the proximal pyrite-rich facies, the distal, marine facies in the West Rand Group are characterized by the presence of magnetite instead of pyrite (Frimmel, 1996).

Similar to pyrite compositional and textural variability, different compositional and textural types of U minerals are present in the Witwatersrand ores and include larger (ca. 0.1 mm), rounded to subrounded uraninite, and clearly secondary brannerite and uraninite, typically filling minute fractures (Randolph, 1958; Schidlowski, 1981). The proportion of U minerals in the orebodies makes the Witwatersrand reefs not only the world's largest known depository of gold but also one of the world's largest uranium depositories. The Au content correlates with that of U in a given reef and the U/Au ratio increases systematically down-slope from high-energy proximal to distal paleoenvironments (Frimmel et al., 2005a).

By analogy with the discussion on the genesis of the pyrite, a postsedimentary hydrothermal formation of all of the uraninite has been suggested by those favoring a hydrothermal model for the entire mineralization process (Barnicoat et al., 1997; Phillips and Law, 2000). This is, however, not tenable in terms of the textural and compositional characteristics of the rounded uraninite. It occurs as inclusions within carbon seams (Fig. 5A) and has been held responsible for the polymerization and cross-linking of hydrocarbons to form the present bitumen (Schidlowski, 1981). Consequently, it must be older than the carbon seams. Furthermore, the uraninite composition is characterized by variable, but generally high, Th contents and high total rare earth element (REE) contents that are typical of high-temperature uraninite (Frimmel, 2005; Cuney, 2010; Frimmel et al., 2014) and with REE fractionation patterns that are most consistent with magmatic uraninite (Depiné et al., 2013). Highly variable trace element ratios, such as Y/REE or Ta/Nb, in the rounded uraninite grains from different reefs, but also within a given reef at different localities, provide further strong support for the detrital nature of the uraninite and its derivation from different magmatic sources (Frimmel et al., 2014).

Dominion Group

Preceding the deposition of the Witwatersrand Supergroup sediments by at least 70 m.y., the up to 2,250-m-thick, largely volcanic succession of the Dominion Group was laid down in a continental rift, possibly within an overall arc setting (Burke et al., 1986; Frimmel et al., 2009). It includes a thin basalticlastic unit and represents the oldest of the supracrustal units on top of the Paleo- to Mesoarchean basement of the Kaapvaal craton. The age of deposition is constrained by the youngest age of pre-Dominion basement, that is 3086 ± 3 Ma (Robb et al., 1992), and 3074 ± 6 Ma, the age of volcanism (Armstrong et al., 1991). The siliciclastic unit includes two conglomerate beds (Lower and Upper Dominion reefs), up to 2 m thick, which represent braided river environments with fluvial channels incised into underlying basement granite. The conglomerate beds, which show some foliation defined by metamorphic sericite and chlorite, contain quartz pebbles, with additional detrital pyrite, arsenopyrite, uraninite, Ti magnetite, ilmenite, cassiterite, chromite, columbite, garnet, zircon, monazite, and gold (Rantzsch et al., 2011). Analogous to the Witwatersrand reefs, pyrite occurs in detrital (compact rounded), syngenetic (concentrically laminated), and epigenetic (euhedral, overgrowths) forms (Randolph, 1958). Similarly, primary (detrital) and secondary gold forms have been distinguished (Feather and Koen, 1975) as well as primary rounded uraninite and secondary U-Ti and U-Th phases (Rantzsch et al., 2011). Analogous to the Witwatersrand ores, the rounded uraninite has a chemical composition typical of detrital uraninite from a magmatic provenance. The major difference to the Witwatersrand ores is a much higher U/Au ratio, which led to the intermittent mining of the Dominion reefs for uranium as the main commodity. Typical grades in the most recently mined deposits are 0.5 to 0.9 kg/t U₃O₈ and 0.55 to 1.1 g/t Au (Rantzsch et al., 2011).

Venterdsorp Supergroup

Above the Witwatersrand Supergroup, separated by a major regional angular unconformity, are the predominantly volcanic rocks of the nearly 3,700-m-thick Venterdsorp Supergroup (Fig. 1). Its basal unit, the Venterpost Formation, contains a basal conglomerate, the Venterdsorp Contact Reef. An age of 2714 ± 8 Ma has been obtained for the overlying thick succession of flood basalt of the Klipriviersberg Group (Armstrong et al., 1991). Volcanism was coeval with siliciclastic sediment deposition of the Venterpost Formation as indicated by evidence of lava extrusion onto wet unconsolidated sediment (Hall, 1997). Consequently, the Venterdsorp Contact Reef is at least 70 m.y. younger than the Witwatersrand Supergroup.

In spite of belonging to an entirely different lithostratigraphic unit, representing a different tectonic setting that postdates the most important tectonic and/or orogenic activity on the fringes of the Witwatersrand basin, the Venterdsorp Contact Reef displays a style of mineralization that is in many respects similar to that in the Witwatersrand Supergroup. The Venterdsorp Contact Reef is one of the richest orebodies and has yielded approximately 8% of the entire South African production (updated from Robb and Robb, 1998). Within the 0.1- to 4.0-m-thick Venterdsorp Contact Reef, a quartzitic facies is distinguished from a conglomerate-dominated one. The latter consists of an oligomictic, matrix- to clast-supported pebble to cobble conglomerate with >95% of the pebbles and/or cobbles being well-rounded quartz. Where the reef is thin, it consists of a single pebble lag, but where it is thick, it reflects multiple channel fills. Overall, several terraces, terrace slopes, and channels can be mapped out. Higher gold grades are in domains that are rich in pyrite (as much as 15 vol %), though locally, pyrrhotite is more dominant than pyrite (Gartz and Frimmel, 1999). Pyrite displays a similar variety of morphological types as in the Witwatersrand reefs (sensu stricto), ranging from rounded, nonporous, evidently detrital forms to rounded, porous and concentrically laminated, synsedimentary, and euhedral, epigenetic types. The last type seems to predominate at first glance, but in many cases, euhedral pyrite grains contain a core of an older, rounded pyrite and represent overgrowths around originally detrital grains. These texturally different pyrite types are also marked by differences in their trace element concentrations (Agungi et al., 2013), with the synsedimentary varieties yielding the highest Au concentrations (avg 6.4 ppm). The rounded types range in diameter from <0.5 to >10 mm and appear in
hydraulic equilibrium with the quartz pebbles and/or cobbles. Gold occurs as discrete spheroidal to toroidal grains, most of which are <50 μm in diameter, and to a lesser extent as irregularly shaped grains, many of which are inclusions within secondary pyrite. Contrary to some of the Central Rand Group reefs, carbon seams are not present, but bitumen globules of undoubtedly hydrothermal origin occur within crosscutting veins (Gartz and Frimmel, 1999).

The position beneath a thick cover of flood basalt led to a very different hydrological and rheological behavior of the Venterdorp Contact Reef in comparison with the Witwatersrand reefs that are typically intercalated within a thick siliciclastic succession. During postdepositional fluid flow through the underlying Witwatersrand Supergroup and into the Venterdorp Supergroup, the Venterdorp Contact Reef experienced comparatively intense alteration and it displays more hydrothermal veins than the other reefs. Both the mineralogy and chemistry of these veins were largely buffered by the chemical contrast between conglomerates and overlying basalt (Gartz and Frimmel, 1999). However, large differences in the composition of individual gold grains, even within a single thin section (Frimmel and Gartz, 1997), are not compatible with a fluid-buffered system but rather attest to very local remobilization of originally detrital gold particles of different provenance (and hence different composition). Intrareef quartz veins locally contain gold, next to a series of sulfide minerals, and even these gold grains show a wide range in Au/Ag/Hg ratios (Frimmel and Gartz, 1997). The spatial association of these veins with pseudotachylyte, as well as fluid inclusion data, suggest the local remobilization of gold (and other components) by meteoric waters as a consequence of the 2023 Ma Vredefort impact event (Frimmel et al., 1999).

**Transvaal Supergroup**

After deposition of the volcano-sedimentary Venterdorp Supergroup, the Kaapvaal craton was subjected to large-scale erosion, resulting in a major unconformity at the contact with the overlying Neoarchean to Paleoproterozoic Transvaal Supergroup rocks, which span an age range from probably 2664 (Sumner and Beukes, 2006) to 2054 Ma, the time of emplacement of the Bushveld Intrusive Complex (Scoates and Friedman, 2006). These rocks were deposited in a vast intracontinental basin (Transvaal basin) in the northern and western parts of the Kaapvaal craton. The lower Transvaal Supergroup (2.67–2.46 Ga) reflects the fluvial to marine transition during flooding of the largely peneplained craton, build-up of the first stable carbonate platform, drowning of that platform and subsequent iron-formation deposition, and finally regression (Sumner and Beukes, 2006). The base of the Transvaal Supergroup consists of up to 2-km-thick siliciclastic rocks of the Wolkberg Group (and likely stratigraphic equivalents). These are overlain by the fluvial to shallow marine Black Reef Quartzite Formation (Fig. 1), which, in turn, grades upward into the carbonate platform of the Malmani Subgroup (Chuni-espoort Group). On the basin margin, the Black Reef Quartzite Formation oversteps the Wolkberg Group (or equivalent units) and rests unconformably on rocks of the Venterdorp or Witwatersrand supergroups. The depositional environment of the Black Reef Quartzite Formation was a deeply incised topographic erosion surface. Fluvial channels on that surface were filled with boulder beds, conglomerate, quartz arenite, and minor argilitic (“channel facies”). These channels, which can be followed laterally over many kilometers, are up to 600 m wide and up to 13 m deep. On the channel levees, sedimentation started with conglomerates of the “blanket facies.” Both facies contain Au- and U-bearing, pyrite-rich oligomictic conglomerates that are very similar to those of the Witwatersrand (Barton and Hallbauer, 1996). By analogy with probably Wolkberg-equivalent units elsewhere, the best available age constraint for the Black Reef Quartzite Formation is 2664 ± 6 Ma (Sumner and Beukes, 2006).

In contrast to the Witwatersrand deposits, the Black Reef Quartzite Formation did not experience significant metamorphism and deformation (its metamorphic grade is subgreenschist facies), which excludes the possibility of postdepositional introduction of the gold by any orogenic, metamorphic fluids. Similarly to the Witwatersrand sedimentary rocks, rounded, evidently detrital, as well as synsedimentary and diagenetic and/or epigenetic pyrite types can be distinguished on morphological, chemical, and isotopic grounds (Barton and Hallbauer, 1996). In contrast to the Witwatersrand deposits, however, U is hosted mainly by (probably secondary) brannerite rather than (detrital) uraninite. Other secondary, evidently hydrothermal, phases are sphalerite and (dominantly radiogenic) galena. Gold occurs as discrete grains, <5 to 250 μm in diameter, and contains on average 86.6 wt % Au and 0.53 wt % Hg (Gautier et al., 2011).

The Black Reef has been mined intermittently for more than a century in seven different areas, all of which are located along the southern margin of the former Transvaal basin where the Black Reef Quartzite Formation lies unconformably above Witwatersrand Supergroup strata (Button, 1978). This spatial relationship strongly suggests that Witwatersrand rocks played a critical role as source of the gold in the Black Reef. An exception to this is a thin bed of auriferous Black Reef at the northwestern margin of the Transvaal basin in the so-called Derdepoortrand graben between Thabazimbi and the Botswana border, known as the Batavia goldfield (Wagner and Ross, 1925), which nevertheless yielded insufficient amounts of gold to attract further interest. Currently, the Black Reef is being mined at Modder East, where 9.24 t Au have been recovered from 2009 to 2012, and for which a resource of 63.46 Mt at 2.03 g/t Au (including a reserve of 8.56 Mt at 4.33 g/t Au) has been reported.

**Other Archean Examples from Outside the Kaapvaal Craton**

**Fortescue basin, Pilbara craton, Western Australia**

Western Australia is the only other region, in addition to the Kaapvaal craton, where an almost continuous record of crustal evolution from the Paleoarchean to the Paleoproterozoic is preserved (Hickman and van Kranendonk, 2012). Granite-greenstone terranes, including elastic successes of the 3.53 to 2.83 Ga Pilbara craton are unconformably overlain by a series of 2.78 to 1.79 Ga volcano-sedimentary successions. The latter comprise the siliciclastic fills of the 2.78 to 2.42 Ga Fortescue, Hamersley, and Turee Creek basins (unified as the Mount Bruce Supergroup), and the 2.21 to 1.79 Ashburton basin (Wyloo Group).
The depositional history of the Mount Bruce Supergroup commenced with crustal extension and volcanic plateau volcanism in the 2.78 to 2.63 Ga Fortescue basin, followed by passive margin deposition in the 2.63 to 2.45 Ga Hamersley basin. Of particular interest here is the Fortescue Group (Fig. 6), a 6-km-thick predominantly volcanic succession that unconformably drapes the eroded paleosurface of the Pilbara craton (Thorne and Trendall, 2001). Early continental rift-related basaltic volcanism and late plume-related volcanism were interrupted by the deposition of up to 1,700-m-thick fluvial and alluvial conglomerates and sandstones with lacustrine shales of the 2766 to 2752 Ma Hardey Formation. Auriferous polymictic, poorly sorted, generally clast-supported, pyrite-rich pebble to boulder conglomerates (and subordinately conglomeratic sandstone) of the Witwatersrand-type have been identified at several stratigraphic levels in the Fortescue Group—that is at the base of the group and in the lower parts of the Hardey Formation. In addition, weakly auriferous pyritic conglomerate has been reported from the middle to upper Hardey Formation (Fig. 6). The auriferous conglomerates represent various facies that range from thin topographic-hollow or paleochannel fills, laterally extensive braided fluvial or sheet-wash deposits, to stacked units in thick, laterally extensive alluvial fan sequences.

Gold occurs as “fine grains and larger flakes and rounded particles up to several millimetres across in the conglomerate matrix” (Huxtable, 2013, p. 57), but larger nuggets have also been reported. Uranium occurs mainly as fine-grained, rounded uraninite grains embedded in kerogen and/or bitumen, analogous to the carbon seams of the Witwatersrand. Pyrite, which can make up as much as 40% of the conglomerate matrix, is predominantly in the form of rounded grains, 2 to 65 mm in diameter. Smaller (<2 mm in size) secondary euhedral grains are rare. Notably, known areas of mineralization are restricted to the zone with the least metamorphic and tectonic overprint—that is, subgreenschist facies and not to zones of higher metamorphic grade (Thorne and Trendall, 2001; Fig. 6) as would be expected if the mineralization had been epigenetic (orogenic).

The depositional setting for the auriferous conglomerates was a >60-km-long half graben that formed in response to early Fortescue crustal extension (Thorne and Trendall, 2001). The source of the gold has been speculated to be in orogenic-type auriferous quartz veins in the underlying craton. Although at least 50 m.y. older than the Ventersdorp...
Supergroup in the Kaapvaal craton, the Fortescue Group is commonly considered similar in tectonic setting, lithology, and genesis. The same applies to the style of mineralization in the Hardey Formation, which is very similar to that of the Ventersdorp Contact Reef.

The conglomerates of the Fortescue Group have been targets of many exploration initiatives that yielded repeatedly elevated Au and U concentrations. Most recent exploration activities have focused on two areas (Nullagine and Marble Bar). For the former, an inferred resource of 8.9 Mt at 1.47 g/t Au has been reported, for the latter, a grade of 0.03 to 9.26 g/t Au over a reef thickness of 0.5 to 5 m and a strike length of 2 km is noted (data from http://www.novoresources.com).

**Bababudan Group, Dharwar craton, India**

Similar to the Western Australian example, the Paleo- to Mesoarchean basement of the Dharwar craton is unconformably overlain by a Neoarchean volcano-sedimentary cover, the Dharwar Supergroup. The lower part of this supergroup constitutes the Bababudan Group, which consists of an up to 1,800-m-thick succession of quartz pebble conglomerate, quartzite, minor calcipelite, iron formation, and felsic and mafic volcanic rocks with intercalated pyroxenite and peridotite (Srinivasan and Ojakangas, 1986). The age of the group is only poorly constrained between 2.91 and 2.72 Ga (Devaraju et al., 2008). Although the rocks have been subjected to greenschist-facies metamorphism, sedimentological features are well preserved. Of interest is the basal dolomitic conglomerate, which can be followed along the southern margin of the Bababudan belt for more than 60 km. It is 4 to 12 m thick, is well sorted, clast-supported, and grades upward into quartz arenite. The pebbles are predominantly vein quartz with subordinate quartzite and chert. The matrix consists mainly of quartz, muscovite, and fuchsite. Pyrite is abundant, both in the form of compact rounded and recrystallized, euhedral grains. For more than a century, these conglomerates have been known to be auriferous and uraniferous. Uranium is located in uraninite, which occurs in three different forms, one of which is rounded and Th rich, and consequently, by analogy with the Witwatersrand ores, detrital (Aurora, 1985). A proper economic evaluation of the Bababudan conglomerates is yet to be completed.

**Moeda Formation, Minas Supergroup, São Francisco craton, Brazil**

Similar to the Kaapvaal, Pilbara, and Dharwar cratons, there is a major regional unconformity in the São Francisco craton between a Meso- to Neoarchean basement, consisting of 3.20 to 2.90 Ga gneisses and a 2.78 to 2.72 Ga granite-greenstone terrane (Rio das Velhas Supergroup), and a Neoarchean to Paleoproterozoic supracrustal succession (Minas Supergroup); for further details see Romano et al. (2013) and references therein. The 2.65 to 2.12 Ga Minas Supergroup (Fig. 7) reflects the evolution from fluviodeltaic continental rift sediments to passive margin platform deposits (Caraça Group), pelagic sediments (Batatal Formation), hauled iron formation (Itabira Group) to flyschoid, and molasse sediments in a foreland basin (Sabará Group). The last was syntectonic with respect to the Transamazonian orogeny, which led to large-scale folding and metamorphism.

At the base of the Caraça Group occurs the siliciclastic Moeda Formation in which up to three different stratigraphic units have been distinguished (Minter et al., 1990): (1) a basal polymictic to oligomictic conglomerate and coarse-grained quartz arenite, reaching as much as 180 m in thickness (Unit I); (2) an up to 80-m-thick, probably shallow-marine, very fine grained quartz arenite succession (Unit II), and (3) a ca. 100-m-thick small to large pebble conglomerate in an argillaceous quartzitic matrix (Unit III). Lateral changes in facies and sediment thickness characterize the three units, but the basal conglomerate is a common feature. The age of this continental rift succession is tightly bracketed by a U-Pb zircon age of 2655 ± 6 Ma for a volcanic unit in the overlying Caué Formation itabirite (Cabral et al., 2012) and a U-Pb age of 2646 ± 15 Ma for a detrital zircon population in the Moeda Formation (Hartmann et al., 2006). Alluvial plains and braided rivers are envisaged as the most likely depositional environment (Minter et al., 1990). The basal conglomerate has been known for decades to be auriferous and has been worked intermittently. The basal quartz pebble, cobble or boulder conglomerate varies markedly in thickness, ranging from a few centimeters in pebble lags up to 10 m in deeper channels, but in most places it is a few meters thick. It is rich in pyrite. As in the other localities discussed, pyrite occurs as compact rounded, detrital grains, 2 to 3 mm in diameter, as rounded porous, concentrically laminated, synsedimentary pyrite, and as euhedral, postdepositional overgrowths. These types of pyrite differ not only in their morphology but also in their chemistry (Koglin et al., 2010a). Gold is present in the form of discrete nuggets with various degrees of rounding, ranging from poorly rounded to completely rounded, and exceeding 1 mm in diameter (Fig. 8). In the less rounded particles, the remnants of well-developed crystals are still recognizable and suggest a derivation from hydrothermal gold as expected for orogenic-type deposits in the stratigraphically underlying greenstones of the Nova Lima Group. Apart from these clearly detrital grains, there is also gold associated with hydrothermal tourmaline. The Ag and Hg contents of individual gold grains range from 7.4 to 17.0 and 6.0 to 8.0 wt %, respectively (Garayp et al., 1991; Koglin et al., 2012). Preliminary Re and Os concentrations for the Moeda gold are 0.06 to 4.08 and 0.02 to 0.30 ppb, respectively (J. Kirk, unpub. data, in Frimmel et al., 2005a).

The gold grade is as high as 10 g/t, locally as much as 150 g/t. Elevated gold grades are exclusively within the basal conglomerate. Footwall and hanging-wall contacts are marked by a sharp decrease (by three orders of magnitude) in Au content. Highest gold grades are in the bottom 20 to 40 cm of the conglomerate where they commonly exceed 10 g/t. Higher within the conglomerate bed, the gold grade decreases to a few grains per ton. Crosscutting quartz veins and shear zones do not show, in most places, significantly elevated Au contents.

**Post-Archean Witwatersrand-Type Deposits and/or Occurrences**

**Huronian Supergroup, Superior province, Canada**

Following the formation of Laurentia’s cratonic nucleus between ca. 2.75 and 2.65 Ga, the cratonic land surface was subject to erosion until rifting and passive margin development
set in along the craton’s southern edge, the remnants of which are preserved in the ca. 2.45 to 2.22 Ma Huronian Supergroup (Fig. 9). Pyrite-rich conglomerates, which are locally auriferous and/or uraniferous, occur at several stratigraphic levels in the Huronian Supergroup, notably in the basal Elliot Lake Group and the overlying Hough Lake Group (Mississagi Formation). The sediment record of the sequence preserves three glacial megacycles, each of which commences with a fluvial unit, and is followed by glacial-marine or lacustrine deposits, turbidites, and postglacial basinal to deltaic deposits. Postdepositional alteration includes a low-grade (greenschist-facies) metamorphic overprint during the 1.85 to 1.80 Penokean orogeny and 1.7 Ga old alkali metasomatism (Ulrich et al., 2011).

The Matinenda Formation near the base of the Elliot Lake Group (Fig. 9) reflects regressive cycles during which several pyrite-rich oligomictic quartz pebble conglomerate beds, interbedded with arkose and sandstone, were deposited from braided fluvial systems on to eroded Archean basement. The age of the formation is well constrained by a 2452 ± 6 Ma volcanic unit at its base (Ketchum et al., 2013). The pyritic conglomerates in the lower and middle portions of the Matinenda Formation served as a major source of uranium and yielded a total of 165 kt U₃O₈ in the Elliot Lake–Blind River mining district until the closure of the last mine in 1996. The average U₃O₈ content of the ore was 0.11 wt % and is mainly due to the presence of rounded uraninite grains, whose spatial distribution describes typical sedimentary heavy mineral concentration. Similar to the association in the Witwatersrand, but much rarer, uraninite is locally surrounded by filamentous kerogen (Mossman et al., 1993). Elevated Th (≤ 9 wt %) and REE contents (≤ 8 wt %) attest to the detrital nature of the uraninite grains and to a pegmatitic or granitic provenance (Cuney and Kyser, 2008). To date, the gold grade is subeconomic. As in the Archean examples above, pyrite is present in different morphological forms. Most of the grains are postdepositional, epigenetic, hydrothermal and/or metamorphic, but compact rounded, evident detrital forms, are also common (Koglin et al., 2010a). Archean granite-greenstone terrane(s), consisting predominantly of highly fractionated granites, provided most of the detritus in the fluvial sedimentary rocks (Craddock et al., 2013). This is evident in the geochemistry of the sedimentary rocks (Sutton and Maynard, 1993) as well as in the pyrite chemistry (Koglin et al., 2010a) and explains a relatively high proportion of REE and Y minerals in the conglomerates: REE and Y were recovered as by-products.
Fluvial sedimentary rocks, including pyritic quartz pebble, cobble, and boulder conglomerates, in the overlying megacycle make up the Mississagi Formation in the Hough Lake Group (Fig. 9). Where the formation rests directly on Archean basement or on the Matinenda Formation, its basal conglomerate is not only pyritic but also auriferous and, in places, uraniferous. Again, a variety of rounded, detrital, and postdepositional pyrite types are distinguished, based on their morphology, trace element contents, and S isotope composition (Ulrich et al., 2011). The same authors describe gold occurring as rare nuggets and minute inclusions in porous detrital pyrite grains and report average gold grades of 0.21 g/t, but as high as 9.4 g/t for the conglomerates and 0.13 g/t for interbedded sandstone.

Other heavy minerals in the Mississagi Formation, apart from pyrite and gold, are zircon, titanite, and rare ilmenite-magnetite (Ulrich et al., 2011). Higher up in the stratigraphy, within the upper Huronian Supergroup, whose age is loosely constrained between 2.4 and 2.2 Ga, detrital pyrite and uranium are absent and hematite is abundant.

Another example of Witwatersrand-type mineralization is known from the Padlei Formation of the Hurwitz Group in the Northwest Territories, Canada. Although this group is interpreted as a correlative of the upper glacial diamictite in the Huronian Supergroup, it is rich in detrital pyrite (Aspler and Chiarenzelli, 1997). Little is known about its Au and U tenor.

Iron oxide-bearing auriferous conglomerates in the Amazon-São Luís-West African craton

Due to strong similarities in the lithologic assemblage, tectonic setting, and geologic evolution of the São Luís craton, southeastern Guyana Shield (Amazon craton), and the West African craton, it has been suggested that all of these Paleoproterozoic cratonic fragments constituted a contiguous landmass (Klein et al., 2005), known as Atlantica. They have in common not only a similar history of crustal growth and continental collision tectonics in the course of the Transamazonian orogeny in South America and the Eburnean orogeny in West Africa but also an apparently contiguous Paleoproterozoic greenstone belt that stretches from eastern Venezuela via northeastern Brazil to Mali and Nigeria (Fig. 10). Throughout this greenstone terrane, orogenic-type shear zone-hosted gold deposits are well known. There are, however, also conglomerate-hosted gold deposits and/or occurrences. The best known of these is Tarkwa in Ghana which has produced >280 t gold to date and whose reported resources amount to 345.2 Mt at 1.31 g/t Au (RMG, Version 2014.01.14, ©Intierra Raw Materials Group).

The Tarkwa ores are hosted in the 2107 to 2097 Ma Tarkwa Group, which forms part of the stratigraphy of the ca. 2170 Ma Birimian Ashanti belt adjacent to the Archean gneiss complex of the Man Shield to the west. The predominantly siliciclastic Tarkwa Group rests unconformably on the volcano-sedimentary 2154 to 2125 Ma Kumasi Group, which in turn lies on top of mafic and/or intermediate volcanic and volcaniclastic rocks of the >2174 Ma Sefwi Group (Perrouty et al., 2012). The unconformity below the Tarkwa Group is strongly sheared and host to a number of shear zone-hosted, orogenic gold deposits. The Tarkwaian sedimentary rocks, which have experienced lower metamorphic grade (lower greenschist facies) than the underlying Birimian units, comprise an upward-fining succession of conglomerates, quartzite, and phyllite that is subdivided, from bottom to top, into the Kawere, Banket, Tarkwa Phyllite, and Huni Formations. Host to the gold are well-sorted oligomictic to polymictic conglomerates of the ≤600-m-thick Banket Formation. Its maximum age is indicated by the youngest detrital zircon as 2107 Ma (Perrouty et al., 2012). The sedimentary facies reflects a change from...
fluvial channel formation and rapid incision due to uplift and reworking to development of meandering channel bars and conglomerates with interbedded silt beds at times of reduced fluvial flow. Mineralization affected the high-energy, early channel fills as well as later sheet flood-dominated alluvial fan deposits. A total of eight, \( \leq 7 \)-m-thick, conglomeratic ore-bodies (reefs) have been recognized with poorly mineralized quartzitic units in between.

Gold in the Tarkwaian reefs occurs as free, visible gold grains in the conglomerate matrix, but also as inclusions within quartz pebbles. The latter provides strong evidence of orogenic-type auriferous quartz veins having been a significant source of the detrital gold in the Tarkwaian sedimentary rocks. In addition, some of the gold is associated with tourmaline and might be of postdepositional nature. There is no pyrite in the auriferous conglomerates but hematite and...
subordinate magnetite are present. While hematite dominates in the Banket Group sedimentary rocks, magnetite is more abundant in the footwall and hanging wall.

Perrouty et al. (2012) recognized six stages of deformation in the region. Sedimentation in the Tarkwa basin occurred after regional folding in the underlying Sefwi greenstone belt (D1) and syn-Kumasi extension (D2), and just prior to thrust faulting and folding in the course of Eburnean compressional tectonics (D3–D6). Most of the orogenic gold deposits in the region are related to D3 and D5 and thus cannot be a source of the gold in the Tarkwaian sedimentary rocks, which must be derived from an older, so-far unspecified source.

A similar situation applies to the Jacobina gold district in Bahia, Brazil, where intermittent mining of metaconglomerates has so far produced >70 t of gold since the early 1950s. Separated by a thrust fault, an Archean gneissic basement is tectonically overlain by the siliciclastic Jacobina Group. This has been interpreted as having been deposited between 2086 and 1883 Ma, initially in a continental rift that was subsequently inverted to a synorogenic basin at approximately 1.94 to 1.91 Ga (Teixeira et al., 2001). Most of the gold in the Jacobina district comes from a basal oligomictic quartz-pebble conglomerate in the lowermost formation of the Jacobina Group (Serra do Córrego Formation). Two conglomerate beds are distinguished with a quartzitic intercalation. The matrix of the clast- to matrix-supported conglomerates consists of recrystallized quartz, sericite, fuchsite, with subordinate detrital zircon, chromite, Fe oxides, rutile, and tourmaline (Teixeira et al., 2001). Highest gold grades (>3 g/t) are in zones rich in hydrothermal pyrite, pyrrhotite, as well as secondary fuchsite, rutile, tourmaline, and andalusite. This spatial association, in combination with the strong sedimentological control of the gold distribution, led Milesi et al. (2002) to suggest a genetic model that involves the mobilization and redistribution of placer gold in the conglomerates by acidic, synorogenic, magmatogenic fluids along 1.98 to 1.91 Ga shear zones.

Somewhat younger than at Jacobina is the gold mineralization in the Roraima Supergroup near the borders between Guyana, Venezuela, and Brazil. The predominantly siliciclastic supergroup represents an intracratonic foreland basin to the southwest of a 2.3 to 2.1 Ga granite-greenstone belt of the Guiana Shield (Voicu et al., 2001). It comprises quartz-pebble conglomerates, sand and siltstones, and minor shale, representing braided fluvial sedimentary rocks from alluvial plain to subaerial braided delta settings. A thin intercalated ash tuff yielded a U-Pb zircon age of 1901 ± 1 Ma (H.E. Frimmel, unpub. data), which provides the best age constraint on the Roraima Supergroup. The arenitic sedimentary rocks show an elevated background Au content of 10 ppb, whereas intercalated conglomerate beds contain >100 ppb Au. Some conglomerate beds above intraformational unconformities yield as much as 23 g/t Au (Minter et al., 2002). The Roraima sedimentary rocks have not experienced any significant metamorphic, tectonic, or hydrothermal overprint, except for local, very minor formation of postdepositional chlorite and pyrite. Gold occurs in the form of variably rounded nuggets,
with a systematic increase in the degree of rounding with increasing distance from the former basin margin (Frimmel et al., 2005b). This observation, paleocurrent directions, and an imprecise Re-Os age of ca. 2.0 Ga (Minter et al., 2002) that overlaps with the age of orogenic-type gold deposits in the hinterland greenstone belt leave little doubt about the detrital nature of the Roraima gold and its derivation from hydrothermal greenstone-hosted deposits, such as Omai. The timing of magmatism and subsequent hydrothermal Au mineralization at the intrusion-related Omai deposit are well constrained as 2092 ± 4 and 2002 ± 5 Ma, respectively (Norcross et al., 2000). Thus the synorogenic gold mineralization is clearly older than the sedimentation in the Roraima basin.

**Fennoscandian Shield**

A conglomerate-hosted gold occurrence (with no resource estimate) has been reported from Kaarestunturi in the central Lapland greenstone belt in Finland (Eilu, 2007). It is hosted by the Paleoproterozoic (1.88–1.80 Ga) Kumpu Group. Gold is reported to occur as detrital particles, together with magnetite and hematite.

**Tertiary Waimumu district, New Zealand**

Fluvial and colluvial auriferous quartz pebble conglomerates in the Waimumu district in southern New Zealand are particularly noteworthy because they bear strong resemblance to Witwatersrand-type ores but, in contrast to the Archean or Paleoproterozoic examples above, are Oligocene to Pliocene in age (Falconer et al., 2006) and therefore much younger. They lack any metamorphic overprint, are poorly lithified, and have been buried to no more than 100 m, yet display a similar association of detrital and secondary morphological types of gold and sulfides, as observed in the Archean and Proterozoic examples described above. The provenance of the detritus is the Otago Schist Belt to the north, which contains numerous orogenic vein-type gold deposits. The study by Falconer et al. (2006) revealed that detrital gold has been mobilized at the micron scale to form secondary authigenic overgrowths. It also revealed a dominance of authigenic or early diagenetic marcasite among the sulfides with textures that are very similar to those of rounded, porous, concentrically laminated pyrite in many of the older Witwatersrand-type deposits. The presence of detrital pyrite and arsenopyrite in some of these Tertiary conglomerates shows that under certain circumstances, sulfides can survive fluvial transport, at least over short distances, also in an oxygenated atmosphere.

**Syngenic Versus Epigenetic Gold Emplacement**

By comparing all of the above examples of conglomerate-hosted gold deposits and/or occurrences, an all-inclusive approach can be taken to the question of syn- versus epigenetic introduction of Au to the host conglomerates. The comparison shows that this type of mineralization occurred at vastly different times, in a range of different tectonic settings, and in rocks that experienced different degrees of tectonic and metamorphic or hydrothermal overprints. Most epigenetic models that have been discussed over the past 25 years, especially for the Witwatersrand deposits, assumed a metamorphic source of the mineralizing fluids from dehydration reactions at depths of some 15 to 20 km (Phillips and Myers, 1989; Barnicoat et al., 1997; Phillips and Powell, 2011) and imply strong similarities with orogenic-type mineralizing systems. It becomes apparent from the above comparison that only some of the examples were affected by postdepositional orogenic activity, whereas others lack evidence of such overprints (e.g., Black Reef in the Transvaal Supergroup, Hardey Formation in the Fortescue Group, Roraima Supergroup, Waimumu district). However, even among those affected by metamorphism, the spatial distribution of the orebodies alone argues strongly against an orogenic-type model. If orogenic fluids derived from metamorphic devolatilization at depth had introduced the bulk of the Au into the host conglomerates, pathways for these ascending fluids would be observed and these pathways would, in certain places, cut across stratigraphic boundaries, whether in the form of steep quartz veins, faults, or shear zones. There is little evidence of such pathways and, in the case of the Witwatersrand basin, the total number of quartz veins is surprisingly low considering the sheer volume of siliciclastic rocks that make up the basin fill. Where present, crosscutting quartz veins and faults are usually barren or generally have a markedly lower Au tenor than the conglomerate beds crossed by them. This applies not only to the Witwatersrand reefs but also to other deposits of similar style elsewhere (Fig. 11). Rather than dispersion of Au from such channel ways into the conglomerate beds, the opposite effect is observed: a dilution of gold grade in the veins. The few exceptions of local auriferous quartz veins or faults (in the Witwatersrand basin) have been overemphasized by some workers in order to promote an epigenetic model (Phillips and Law, 2000) but these should not be viewed as representative of the entire mineralizing system.

From a practical point of view, orogenic gold deposits are structurally highly complex, requiring hundreds of drill holes to define a resource. This is in stark contrast to the relatively simple structure of the Witwatersrand orebodies. Some protagonists of an epigenetic model suggest that the lack of new discoveries (in the Witwatersrand) might be due to explorationists having utilized the wrong genetic model, i.e., paleoplacer model (Phillips, 2013). In this context, it should be born in mind that those companies who have subscribed to epigenetic models in the past have failed to discover a single orebody that crosses stratigraphy!

Apart from the stark contrast in the overall geometry between stratiform conglomerate-hosted and typical epigenetic, orogenic-type gold deposits, there is a wealth of lithochemical, mineral chemical, and isotopic evidence against significant postdepositional introduction of Au into the host conglomerates. If Au had been introduced by bedding-parallel fluid flow preferentially along conglomerate beds, as suggested for example by Barnicoat et al. (1997), some geochemical dispersion both into the footwall and hanging wall would be expected. In most cases, the lithologic boundaries are also sharp with regard to gold grade. Although some minor dispersion-related alteration is recorded locally in the hanging wall, most of the alteration is restricted to the footwall and explainable by acid weathering in paleosols beneath erosion surfaces (Frimmel and Minter, 2002).

Witwatersrand-type deposits typically exhibit a broad correlation between grain size, Au content, and the concentration of other elements that indicate detrital phases (e.g., Zr, Cr,
in the ore paragenesis, in all Witwatersrand-type deposits studied so far are not compatible with postdepositional epithermal formation, but instead reflect chemical variations that can only be explained by detrital particles derived from different sources in the hinterland. The trace element contents and S isotope ratios in the rounded pyrite forms differ systematically from those of the euhedral, postdepositional types. The elevated Th and REE contents as well as considerable variations in certain trace element ratios, such as REE/Y or Nb/Ta, in the rounded uraninite are incompatible with crystallization from an epithermal fluid. Similarly, a strong case can be made for the carbon seams representing in situ microbial mats and not remnants of migrating postdepositional oils (Mossman et al., 2008). Age data based on different isotope systems applied to different ore components (including the gold) are older than the maximum age of sedimentation as summarized by Frimmel et al. (2005a). Finally, the exceptionally high Os concentrations of the Witwatersrand gold (and pyrite) contradict any kind of hydrothermal model because of the very low solubility of Os in aqueous hydrothermal fluids. Hydrothermal transport of Au by crustal fluids (as in an orogenic model) would invariably increase the $^{187}\text{Os}/^{188}\text{Os}$ ratio of the gold (and rounded pyrite) due to mixing of Os from the primary mantle source and crustal sources. This is not reflected by the data available. Both gold and rounded pyrite have low initial $^{187}\text{Os}/^{188}\text{Os}$ ratios that correspond to mantle values at ca. 3.0 Ga (Kirk et al., 2001, 2002) and show no evidence of crustal Os. Taking into account all of the above arguments, a purely epigenetic model with postdepositional introduction of Au into the host conglomerates by crustal (metamorphic) fluids has to be rejected and a primarily syngenetic origin for the gold mineralization is the only tenable option available.

There is no doubt, however, that many of the examples of Witwatersrand-type gold deposits and/or occurrences discussed above show signs of metamorphic and/or hydrothermal overprint and contain strong evidence of secondary postdepositional mineral growth, such as euhedral pyrite overgrowths, secondary gold (in places as overgrowths), secondary uraninite or brannerite, and others. This has laid the foundation for the “modified” paleoplacer model (Ramdohr, 1958; Frimmel et al., 1993; Robb and Meyer, 1995), in which the short-range (μm- to dm-scale) mobilization of originally detrital gold by postdepositional hydrothermal fluids largely within the host conglomerates is assumed. Recently, this model has been challenged by Large et al. (2013) who, while agreeing with a modified paleoplacer model, postulated long-range movement of Au by postdepositional hydrothermal fluids and questioned the source of the detrital gold.

The presence of detrital gold particles in support of a syngenetic model is now well documented for several of the other conglomerate-hosted examples described above, specifically for the Moeda, Harvey, and Banket Formations, and the Roraima Supergroup. In these cases too, true nuggets are similar in morphology to recent placer gold, leaving little room for doubt of a paleoplacer genesis. In the Witwatersrand, however, nuggets have only rarely been observed and these are generally <0.1 mm in diameter (Hallbauer, 1996; Minter et al., 1993). Most of the “micro-nuggets” in the Witwatersrand are of torroidal or spheroidal shape (Fig. 3C) but subrounded crystals or crystal aggregates, as typical of the Moeda or

U). For example, a strong correlation between Au and Zr is expected if all the gold is concentrated together with detrital zircon. In some cases in the Witwatersrand, however, this correlation is poor, which has led some workers to erroneously reject a paleoplacer model (Fox, 2002). All of the examples with poor correlation result from a predominance of Zr-rich but Au-poor samples. This is to be expected in heavy mineral concentrates derived from gold-poor sources. Significantly, very few samples (i.e., only a single analysis out of close to 600) have been reported that are Au rich but Zr poor as one would expect for hydrothermal systems (Frimmel et al., 2005a). Thus one of the main arguments held up against the paleoplacer model (Fox, 2002) is in fact supportive of this model.

The mineral chemical and isotopic characteristics of the compact rounded pyrite (Hofmann et al., 2009; Guy et al., 2010; Koglin et al., 2010a; Ulrich et al., 2011; Agangi et al., 2013; Large et al., 2013) and uraninite grains (Cuney, 2010; Depiné et al., 2013; Frimmel et al., 2014), both critical phases of the ore paragenesis, in all Witwatersrand-type deposits studied so far are not compatible with postdepositional epithermal formation, but instead reflect chemical variations that can only be explained by detrital particles derived from different sources in the hinterland. The trace element contents and S isotope ratios in the rounded pyrite forms differ systematically from those of the euhedral, postdepositional types. The elevated Th and REE contents as well as considerable variations in certain trace element ratios, such as REE/Y or Nb/Ta, in the rounded uraninite are incompatible with crystallization from an epithermal fluid. Similarly, a strong case can be made for the carbon seams representing in situ microbial mats and not remnants of migrating postdepositional oils (Mossman et al., 2008). Age data based on different isotope systems applied to different ore components (including the gold) are older than the maximum age of sedimentation as summarized by Frimmel et al. (2005a). Finally, the exceptionally high Os concentrations of the Witwatersrand gold (and pyrite) contradict any kind of hydrothermal model because of the very low solubility of Os in aqueous hydrothermal fluids. Hydrothermal transport of Au by crustal fluids (as in an orogenic model) would invariably increase the $^{187}\text{Os}/^{188}\text{Os}$ ratio of the gold (and rounded pyrite) due to mixing of Os from the primary mantle source and crustal sources. This is not reflected by the data available. Both gold and rounded pyrite have low initial $^{187}\text{Os}/^{188}\text{Os}$ ratios that correspond to mantle values at ca. 3.0 Ga (Kirk et al., 2001, 2002) and show no evidence of crustal Os. Taking into account all of the above arguments, a purely epigenetic model with postdepositional introduction of Au into the host conglomerates by crustal (metamorphic) fluids has to be rejected and a primarily syngenetic origin for the gold mineralization is the only tenable option available.

There is no doubt, however, that many of the examples of Witwatersrand-type gold deposits and/or occurrences discussed above show signs of metamorphic and/or hydrothermal overprint and contain strong evidence of secondary postdepositional mineral growth, such as euhedral pyrite overgrowths, secondary gold (in places as overgrowths), secondary uraninite or brannerite, and others. This has laid the foundation for the “modified” paleoplacer model (Ramdohr, 1958; Frimmel et al., 1993; Robb and Meyer, 1995), in which the short-range (μm- to dm-scale) mobilization of originally detrital gold by postdepositional hydrothermal fluids largely within the host conglomerates is assumed. Recently, this model has been challenged by Large et al. (2013) who, while agreeing with a modified paleoplacer model, postulated long-range movement of Au by postdepositional hydrothermal fluids and questioned the source of the detrital gold.

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Numerous sources have been suggested to explain the huge amount of gold in the Witwatersrand basin, depending on the favored genetic model. A placer model requires sources in the hinterland to supply an adequate amount of particulate gold. The intuitively most likely source presents itself in the form of several greenstone belts in the vicinity of, and beneath, the Witwatersrand basin. Aauriferous quartz veins are well known in today’s exposed parts of these greenstone belts and at least some of them (in the Barberton greenstone belt) are of an age that is in perfect agreement with the most likely age of the Witwatersrand gold and associated heavy minerals—that is, 3.08 and 3.04 Ga (Dziggel et al., 2010). However, differences to the composition of greenstone-hosted gold (Barberton gold contains orders of magnitude less Os; Frimmel et al., 2005a), the lack of larger nuggets and gold-bearing quartz pebbles, as well as mass-balance problems discussed by Robb and Meyer (1990) all cast doubt on such a source.

Elevated Au contents, consistent gold composition, and similar alteration mineral assemblages in hydrothermally altered granites of suitable age in the hinterland led to the suggestion that erosion of these 3.1 to 2.9 Ga granites provided the main source of the gold (Hallbauer and Barton, 1987; Robb et al., 1992). This hypothesis lost favor, however, when it became evident that the low-temperature alteration in these granites is most likely related to basinal fluids that are younger than Witwatersrand sediment deposition (Klemd, 1999). Since then, a high-temperature potassic, REE-enriching alteration has been recognized in the basement directly adjacent to the Witwatersrand basin. The age of this alteration is within error of that of 3062 Ma granitic and gabbroic calc-alkaline magmatism, and the alteration has been considered a product of late magmatic autometasomatism (Frimmel et al., 2009), possibly comparable with porphyry- or IOCG-style mineralization. While some kind of hydrothermal alteration in the immediate hinterland is thus documented, it remains unclear whether this alteration also led to any Au enrichment. Furthermore, the regional extent of this alteration remains unknown.

As an alternative to eroded gold-bearing deposits in the pre-Witwatersrand basement and/or hinterland, Hutchinson and Viljoen (1987) proposed synsedimentary pyritic exhalites along the basin margin as the dominant gold source, but this hypothesis remained purely speculative because no direct evidence of gold concentration by such exhalative activity has been recorded to date. Moreover, it would not explain the fact that Au mineralization occurred at different times in very different types of tectonic setting, most of which are not conducive for exhalative activities.

All of the above hypotheses share the assumption that the hinterland must have contained elevated Au contents, a view that has been challenged by Loen (1992) who calculated that an average background value typical of Archean granite-greenstone terranes (0.4–6.8 ppb) would be sufficient to explain the world’s largest accumulation of gold. Those workers who prefer an altogether epigenetic model for the Witwatersrand deposits suggest the source of the Au within metamorphosed rocks beneath the reefs. Two principal source regions have been proposed in this regard. One scenario calls for the dissolution of background Au (at the ~2-ppb level) in the course of progressive devolatilization reactions during greenschist- to amphibolite-facies metamorphism of underlying greenstone belts (Phillips and Powell, 2011). Such metamorphic fluids are typically of moderate salinity and are dominated by H2O-CO2-H2S. Their capacity to carry Au, whether as AuHS(aq) and Au(HS)2− at lower temperatures and respectively low and neutral pH, or as AuOH(aq) or AuCl₃− at higher temperature, is thermodynamically well constrained (Stefánsson and Seward, 2004). Equally well documented is the efficiency of such fluids in the formation of gold deposits as evidenced by the numerous orogenic-type deposits worldwide (Goldfarb et al., 2005).

The location of orogenic gold deposits is typically controlled by large structures, shear zones, and/or steep, crosscutting quartz (-carbonate) veins. In the Witwatersrand basin (as in all other examples of Witwatersrand-type deposits described above with the possible exception of Jacobina), such large-scale pathways for ascending mineralizing metamorphic fluids are not documented and the geometry of the conglomerate-hosted deposits is in stark contrast to any known orogenic gold deposit elsewhere in the world. Other arguments against a synorogenic model have been already summarized in the previous chapter.
Last but not least, if Witwatersrand-type mineralization was the product of metamorphic fluids leaching Au from source rocks with average background Au values (on the order of a few ppb), this type of mineralization would be expected in conglomerates of all ages. Similarly, based on the secular variation of orogenic gold deposits (Barley and Groves, 1992; Goldfarb et al., 2005; Groves et al., 2005b) that shows several prominent peaks of formation throughout the Precambrian and Phanerozoic, one would not expect the secular distribution of Witwatersrand-type deposits to be as limited as it is. Interestingly, no examples of Meso-, Neoproterozoic, or Phanerozoic Witwatersrand-type deposits are known anywhere in the world, with the exception of the economically insignificant Tertiary gold and Fe sulfide-bearing quartz pebble conglomerates of the Waimumu district in Southland, New Zealand (Falconer et al., 2006). Collectively, the above arguments suggest that a metamorphic source of the mineralizing fluid for this type of ore deposit is not considered a viable option.

The second option suggested for a possible source of Au in an epigenetic model is pyritic shales within the Witwatersrand basin (Large et al., 2013). While agreeing with the modified paleoplacer model, these authors suggested long-distance post depositional fluid flow and introduction of Au (plus U, As, S, Te, and many others) to explain the chemistry of secondary, euhedral pyrite overgrowths and gold. According to their hypothesis, the detrital pyrite is of “intrabasinal origin and sourced from the West Rand Group... or lower Central Rand Group” (Large et al., 2013, p. 1239), and the gold was introduced in two stages, once by sedimentary reworking of older sediments within the Central Rand basin and then by organic-rich basinal brines whose circulation was triggered by the Bushveld event. The hypothesis is based on the study of samples from a single reef, the Carbon Leader Reef, which limits its applicability to all Witwatersrand deposits. Whereas some of the rounded, evidently detrital pyrite in the Carbon Leader may well be derived from diagenetic pyrite in shales of the underlying West Rand Group, the above hypothesis does not explain the widespread occurrence of rounded compact pyrite in effectively all Archean and >2.4 Ga Paleoproterozoic fluvial conglomerates on all cratons. For example, detrital pyrite in conglomerates of at least the lower West Rand Group and the Dominion reefs cannot have an “intrabasinal” origin because no significant volumes of underlying diagenetically lithified mudstones existed at the time of sedimentation. A further shortcoming of the hypothesis of Large et al. (2013) is their inferred fluid source. The 2054 Ma Bushveld event had a great thermal effect on the central Kaapvaal craton and this has been documented in some of the Witwatersrand rocks (Gibson and Wallmach, 1995; Frimmel, 1997). However, it was unlikely to have triggered major basinal fluid flow because the Bushveld event occurred several hundred million years after sediment deposition and diageneric, and long after the peak metamorphic conditions had been attained; therefore, it would have affected a Witwatersrand basin fill that was already relatively dry with very little, if any, basinal fluid preserved. This is well reflected by the lack of Bushveld-event ages in postdepositional xenotime age spectra from the Witwatersrand strata (Kositcin et al., 2003). Much of the fluid inclusion data referred to by Large et al. (2013) relate to locally gold-mobilizing fluids that percolated through the Witwatersrand (and other) rocks due to the generation of a secondary permeability in the course of the 2023 Ma Vredefort impact (Frimmel et al., 1999; Hayward et al., 2005) and not to the Bushveld intrusion. Last but not least, the hypothesis of long-range metal transport by basinal brines does not explain the same style of mineralization in so many other pyritic conglomerates elsewhere in the world, where the same morphological and compositional variety of pyrite types exist with no geologic evidence of post depositional basin-wide fluid flow.

Although a primarily syngenic model, with locally variable post depositional remobilization, is indicated for all examples of conglomerate-hosted gold deposits, significant differences exist in the texture and morphology of the gold grains, which pertain to the likely source of the gold. Critical in this regard is the presence or absence of “proper” nuggets. Although gold nuggets are regarded by most geologists as masses exceeding 1 g in weight or 4 mm in diameter, many gold particles in modern placer deposits are much smaller (Hough et al., 2007). Fluvial placer gold varies in size, depending on transport distance. It enters the fluvial system either as free gold particles, which are generally small (<1 mm) or more commonly as inclusions in ore clasts. Communion of these clasts in the first few kilometers of the fluvial system releases much of the gold in the clasts, some of which are larger (several mm) than the free gold particles. Consequently, gold particle size increases in the first few kilometers and then decreases again farther downstream (Youngson and Craw, 1999). Placer gold that is hosted by till in glacial terranes of Canada has been described to be predominantly (85%) of silt to fine sand size (Averill, 1988). Most gold particles in the chemically weathered zones in semiarid, subtropical to tropical regions have a diameter between 0.1 and 0.4 mm (for more details on lateritic gold deposits see Freyssinet et al., 2005). Consequently, for practical purposes a lower limit of 0.1 mm is arbitrarily set for the diameter of a rounded or flattened, mechanically eroded gold grain to qualify as a placer particle. The presence of such placer particles and/or quartz pebbles with primary gold inclusions or even clasts of greenstone with primary gold inclusions (as observed in the Roraima, W.E.L. Minter, pers. commun., 2014) are taken as evidence of gold derivation from free gold in quartz veins within orogenic, greenstone-hosted deposits. Good examples are those from the Moeda, Hardey, and Mississagi Formations, Tarkwa Group, and Roraima Supergroup. In all of these cases, one or more specific point-source(s) existed in the form of eroded gold deposits with gold grains large enough to form proper nuggets in the detritus.

The lack of proper nuggets has been the biggest obstacle for general acceptance of a placer model for the Witwatersrand deposits. Indeed, Witwatersrand gold, if not present in secondary textural positions, occurs in grains that are in most cases <0.1 mm in diameter. If rounded, spheroidal, or toroidal in shape, they are referred to as “micronuggets.” The relative proportion of such micronuggets in the total gold budget of the Witwatersrand reefs is not well constrained because of difficulties in identification in hand specimen or even under a petrographic microscope. In the past, a proper assessment of
the morphology of the gold particles required the dissolution of the host rock in hydrofluoric acid and subsequent characterization of the thus released particles under a petrographic or an electron microscope. Only few studies of this kind have been carried out, such as that by Minter et al. (1993) on the Basal reef within the Central Rand Group, which showed that approximately 75% of all the gold therein occurs as micronuggets. More recently, microfocus X-ray computer tomography has become a powerful tool, not only for the identification and quantification of minerals in a given rock, but also for the visualization of the three-dimensional distribution and shape of these minerals. Application of this technique to some Witwatersrand reefs (including the Ventersdorp Contact Reef) shows that micronuggets are far more common than previously perceived (Nwaila, 2014).

Several Witwatersrand reefs (those within the Witwatersrand Supergroup) and other reefs in younger units in the Kaapvaal craton show sedimentological evidence of reworking of older gold-bearing conglomerates. This applies specifically to those higher up in the stratigraphy, that is, the reefs in the upper Central Rand Group and younger units, such as the Beatrix reef, the Ventersdorp Contact Reef, and the Black Reef. In particular, the spatial distribution of gold in the Ventersdorp Contact Reef and the Black Reef on top of post-Witwatersrand erosion surfaces strongly suggests a gold source in underlying adjacent Witwatersrand reefs. Even within the upper Central Rand Group, there is sedimentological and structural evidence (low-angle unconformities) that suggests gold derivation from reworked older reefs in the West Rand Group and especially the lower Central Rand Group. What was, however, the primary source of all this reworked, very fine grained detrital gold? As argued above, erosion of former auriferous, greenstone-hosted quartz veins or any other form of hydrothermal gold deposits, such as high-temperature magmatogenic gold of porphyry or IOCG affinity, as seemingly suggested by the high Os concentrations and Re-Os isotope data for the gold, is unlikely to have provided the detrital gold. Both the size and shape of the micronuggets can be explained, however, by reworking of the filamentous gold that is omnipresent in the kerogenous “carbon” seams. Erosion of these fragile microbial mats would have invariably released the filamentous gold particles therein. Subsequent transport and saltation, whether fluvial or eolian, would have led to the kind of overfolded rims recorded in many of the torroidal gold particles (Minter et al., 1993). The fact that the majority of the gold in some reefs, especially in the lower Central Rand Group, occurs within former microbial mats, suggests that this gold was the principal source of the reworked detrital gold in the Witwatersrand reefs. Derivation of placer gold from both microbial mats and hydrothermal vein-type deposits in the hinterland is feasible for most other examples of Witwatersrand-type deposits, because biogenic carbonaceous material has been reported from all, except for the Tarkwaian deposits (Minter, 1991). However, judging from the low relative amounts of both carbonaceous matter and the presence of proper gold nuggets in these other examples, microbially mediated gold must have played a lesser role. This raises the question of why there has been so much more gold in the Witwatersrand deposits compared to all the other paleoplacer deposits.

Why So Much Gold in the Mesoarchean?

A Mesoarchean gold superepisode

The secular distribution of known gold in different types of gold deposits is strongly bimodal with the main peak reflecting Meso- and Neoarchean gold addition to the upper crust and a second peak in the Cenozoic (Frimmel, 2008). The latter is essentially an artifact of better preservation of younger crust, which applies specifically to gold deposit types that formed at shallow crustal levels, such as epithermal and placer deposits. Nevertheless, some 95% of all placer gold is Archean in age, with only 3% from Cenozoic deposits. Updated past production and reported resource data indicate that about 30% of all known gold that is concentrated in deposits worldwide is Mesoarchean. An even stronger bias toward the Archean is obtained if one takes into account the crustal preservation effect as many Archean deposits have been removed from the rock record through tectonic recycling. Interestingly, very few Palearchean gold deposits are known. The bulk of gold produced is from Mesoarchean deposits, primarily those in the Central Rand Group. The Neoarchean gold component is essentially greenstone-hosted orogenic gold and the paleoplacer deposits of the Ventersdorp Contact Reef. The latter, which alone constitutes about 8% of all known placer gold, is shown above to be derived largely from the reworking of the underlying Central Rand Group reefs and consequently should also ultimately come from a Mesoarchean source.

The prominence of Mesoarchean gold is even more pronounced if one follows the contention that most of the post-Mesoarchean gold deposits worldwide resulted from the (repeated) reworking of Archean crust and gold therein, whether by mechanical sedimentary processes at shallow levels (as in the younger paleoplacer deposits) or by remobilization through fluids and melts in the course of large-scale, plate tectonically driven crustal recycling (Frimmel, 2008). The age and crustal provenance of the primary source rocks for Mesoarchean gold in the Witwatersrand basin is constrained by U-Pb and Lu-Hf isotope data on detrital zircon grains from particularly well-mineralized conglomerates in the Central Rand Group (Kositcin and Krapez, 2004; Koglin et al., 2010b); most of the detrital zircon grains come from 3060 Ma juvenile crust. Remnants of this crust have been identified as a former magmatic arc along the northern and western margin of the Central Rand basin (Poujol et al., 2003; Frimmel et al., 2009). Incidentally, available Re-Os ages for the Witwatersrand gold and detrital pyrite (3010 ± 110 and 2990 ± 110 Ma, respectively) are Mesoarchean (Kirk et al., 2001, 2002) and thus are not only older than the maximum age of conglomerate deposition but also overlap with the dominant age of the above magmatic arc(s).

The Mesoarchean gold trap

There is no doubt that the time period of about 3.0 Ga was critical for the gold enrichment of the continental crust. All of the younger examples of Witwatersrand-type placer deposits show evidence of derivation from point sources, such as greenstone-hosted orogenic gold, or from the resedimentation of older placer deposits along erosional and angular unconformities. In contrast, no specific sources of particulate gold have been identified for the older deposits in the West
Rand Group and especially in the lower Central Rand Group. Microbial mats, as suggested above, might have provided an effective trap but the source of the gold is unclear. The lack of proper nuggets in these older deposits is crucial. It removes the necessity for the presence of discrete gold deposits in the hinterland and instead allows for input of Au during sedimentation, either in colloidal form and/or dissolved in meteoric surface waters (Fig. 12). This is predicted by current understanding of the Archean environment and by thermodynamics.

An overall O₂-deficient, relatively CO₂-, CH₄- and H₂S-rich Archean atmosphere with <0.1 % of present-day atmospheric O₂ levels (Canfield, 2005) is generally accepted and supported by the omnipresence of pyrite and uraninite in Archean fluvial deposits (Frimmel, 2005). The pH of rain water at that time was most likely mildly acidic, estimated at approximately 4 (Krupp et al., 1994). Intense chemical weathering under such an acidic atmosphere is documented for the Witwatersrand paleosurfaces (Frimmel and Minter, 2002) and would have caused an increase of the pH in corresponding river water to near-neutral conditions because of alteration of feldspars to clays. Concurrent volcanogenic SO₂ and H₂S emission invariably led to the concentration of H₂SO₄ and H₂S in the river water. Some variation in the H₂S content of the river water across the Archean Eon is indicated by the temporal and spatial distribution of siderite and pyrite in Archean sedimentary rocks. Detrital siderite, indicative of lower atmospheric H₂S concentration, has been described from both Meso- and Neoarchean fluvial sandstones in the Pilbara craton (Rasmussen and Buick, 1999). Far more common in coeval sedimentary rocks is, however, fresh detrital as well as synsedimentary pyrite in effectively all Archean to early Paleoproterozoic fluvial deposits, especially the Witwatersrand, where siderite is absent.

The presence of pyrite instead of siderite implies a generally higher atmospheric H₂S fugacity of >10⁻⁵ in the postulated O₂-deficient atmosphere (Rasmussen and Buick, 1999). Under such conditions, Au would be dissolved in meteoric (and most likely shallow seawater) as AuHS⁻ or as Au(HS)₂⁻, depending on the proportion of dissolved S²⁻ and Cl⁻ (Stefánsson and Seward, 2004). The solubility of gold in aqueous sulfide solutions has been determined experimentally at elevated temperatures between 150° and 500°C (Shenberger and Barnes, 1989; Hayashi and Ohmoto, 1991; Stefánsson and Seward, 2004). Extrapolation of their results to lower temperatures does not change the overall topology of gold solubility contours in fO₂-pH space and suggests gold solubility in Mesoarchean river water to be several orders of magnitude greater (possibly as much as 0.1−1 ppb) than in modern river water. This, together with intense chemical weathering and much higher erosion rates on a vegetation-free land surface, would have resulted in a very high fluvial Au flux from the Mesoarchean land into fluviodeltaic and coastal environments (Fig. 12).

There is convincing evidence for cyanobacteria having been present from as early as 2.7 Ga, but the switch from anaerobic to aerobic respiration probably took place much earlier (Noffke et al., 2008). Indirect evidence for oxidizing conditions at least in the near-surface ocean comes, for example, from U enrichment in >3.7 Ga sedimentary rocks of the Isua supracrustal belt in West Greenland (Rosing and Frei, 2004). The atmosphere at that time could not have been oxygenated because of ample evidence of mass-independent S isotope fractionation in contemporary sedimentary rocks. Local oxidizing conditions in seawater and an overall reducing atmosphere do not have to be mutually exclusive, as discussed by Canfield (2005). Consequently, stromatolite-like cyanobacterial

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**Fig. 12.** Schematic block diagram of the perceived depositional environment of Witwatersrand reefs, using the Steyn and Basal reef paleoplacers in the Welkom goldfield as examples (modified from Minter et al., 1993), also highlighting the four principal stages involved in the formation of the gold deposits.
mats most likely existed already at around 2.9 Ga, when the bulk of the Witwatersrand gold was deposited. Most recently, independent evidence of oxygenic photosynthesis having taken place already at around 3.0 Ga has been presented by Planavsky et al. (2014). These authors argue, based on Mo isotope data, for the presence of Mn oxides in the water column, which requires a certain level of free oxygen. In such an environment, photosynthesizing microbes are likely to have served as large-scale fixator of Au that was input by Mesaoarchean meteoric waters. As predicted by thermodynamics, already a very small increase in \( f_{O_2} \) would reduce the solubility of gold by orders of magnitude. It is suggested, therefore, that local photosynthetic production of \( O_2 \) led to the precipitation of gold on the surface of the microbial mats according to the reaction:

\[
4\text{Au(II)} + 15O_2 + 2H_2O = 4\text{Au} + 8SO_4^{2-} + 12H^+.
\]

Whether biochemical processes also played a role in the precipitation of gold, as suggested by Horscroft et al. (2011), remains unclear.

Further evidence of Au transport by Mesoarchean river waters comes from pyrite chemistry. It has been shown that among all the various morphological pyrite forms, the rounded, concentrically laminated pyrite has the highest concentrations of Au (Koglin et al., 2010a), which occurs as minute inclusions within the concentrically arranged laminae. This pyrite type, which is present in all of the coarser grained siliciclastic units of the Witwatersrand Supergroup, the Ventersdorp Contact Reef, and even the much younger Black Reef, has long been suspected to be of synsedimentary origin (Barton and Hallbauer, 1996; England et al., 2002; Guy et al., 2010). The finding of anomalously high Au in this pyrite type has since been confirmed by further studies on pyrite chemistry by Large et al. (2013) and Agangi et al. (2013) and the interpretation of this pyrite to have formed by in situ growth in the wet sediment has found support from S isotope studies (England et al., 2002; Hofmann et al., 2009; Guy et al., 2012).

The only plausible explanation for this observed accumulation of gold in syngenetic pyrite that has grown in fluvial to fluvial-deltaic environments is the trapping of Au from river water on the surface of the growing pyrite grains, whether by purely chemical processes or by microbial mediation.

The postulated higher run-off of Au from the Mesoarchean land surface should be also reflected in contemporaneous marine sedimentary rocks. Dissolved Fe(II) in the seawater acted as sink for some of the early microbially produced \( O_2 \). This is evident from the prevalence of pyrite in fluvial and fluvio-deltaic deposits and that of magnetite in contemporaneous marine deposits, such as in widespread magnetite-rich shales (akin to iron formations) in the West Rand Group (Frimmel, 1996; Horscroft et al., 2011). Furthermore, Archean iron formations are enriched in gold by a factor of approximately 50 relative to the continental crustal background (Meyer and Saager, 1985; Rudnick and Gao, 2005).

The Mesoarchean crustal gold endowment

Apart from the above paleoenvironmental considerations, another factor may have contributed to the unparalleled gold endowment of the Mesoarchean continental crust. The exceptionally high Os concentrations in Witwatersrand gold (0.003–4 ppm) may indicate an ultimately magmatic origin. Although several greenstone belts in the surroundings of the Witwatersrand basin (Murchison belt, Kraaipan-Amalia belt) host meso- and epithermal gold deposits, these deposits were unlikely to have contributed a significant proportion of the Witwatersrand gold for reasons discussed above. Instead, finely dispersed physical gold within magmatic rocks in the hinterland, in whatever mineralogical association, may have been leached by Mesoarchean meteoric waters. This Mesoarchean hinterland most likely had higher Au background levels than at any other time because it consisted of largely juvenile crust that formed when the Earth’s mantle reached its maximum temperature (Labrosse and Jaupart, 2007). The degree of partial melting in the upper mantle would have been at its maximum, which would have favored production of sulfur-undersaturated melts. Such melts are known to be richer in Au than sulfur-saturated ones, as evidenced by elevated Au contents of komatiites (Pitcairn, 2011).

Those Witwatersrand gold and pyrite samples that plot close to an Re-Os isochron have initial \( ^{187}\text{Os}/^{188}\text{Os} \) ratios that correspond roughly with the chondritic mantle composition in the Mesoarchean (0.105) but other samples display considerable scatter of the Re-Os isotope ratios (Kirk et al., 2001, 2002). Calculated Re-Os model ages of approximately 3.0 to 3.1 Ga (Kirk et al., 2002; Schaefer et al., 2010) and chondritic initial \( ^{187}\text{Os}/^{188}\text{Os} \) ratios indicate a mafic and/or ultramafic mantle domain as ultimate source, but significant and variable deviation from 3.0 to 3.1 Ga isochrons of many data points suggests not only Re-Os isotope fractionation during postdepositional gold mobilization (Schaefer et al., 2010) but possibly also fractionation in the course of Os transport from the source rocks to the sites of gold concentration in the microbial mats.

Looking at the PGE, Re-Os, and Mo isotope characteristics of sedimentary rocks of different age, Siebert et al. (2005) suggested PGE transport to the sediments mainly in particulate form in the Paleo- to Mesoarchean as opposed to dissolution and/or precipitation in the Neoproterozoic and Phanerozoic and explained this by the lack of \( O_2 \) in the Archean. Interestingly, some of their data, specifically those from the 2.7 Ga Belingwe greenstone belt in Zimbabwe and from the 2.95 Ga West Rand Group, display a stronger PGE fractionation and exceptionally high Re and Os concentrations compared to those of the 3.23 to 3.10 Ga samples from the Barberton greenstone belt, with Re/Os ratios similar to those of younger sedimentary rocks. Although speculative, this may indicate that the emergence of aerobic microbes during the Mesoarchean might have provided conditions conducive for trapping Os, similar to that for Au.

Why So Much Gold in the Kaapvaal Craton?

The discussion above provides a plausible explanation for the extraordinary amount of gold transferred into the crust during the Mesoarchean, however, it does not answer the most crucial question from an economic point of view, and that is whether the gold endowment of the Kaapvaal craton is the exception or the rule. So far, no equivalent to the Witwatersrand basin has been discovered in terms of gold endowment. This might not be due to poor exploration in other areas, but rather due to geologic reasons. Because Au is a
highly siderophile element (HSE), together with Re and the platinum group elements (PGE), these metals should effectively all be stored in the Earth's core. Metal and/or silicate enrichment factors in the core of >800 and 500 have been calculated for Re and PGE, and for Au, respectively, resulting in a theoretical proportion of 98% of Au residing in the Earth's core (McDonough, 2005). Yet, elevated Au (together with the other HSE) contents in the upper mantle must be invoked in order to explain the extraordinary addition of Re- and Os-rich gold to the Mesoarchean crust.

It has long been known that the concentrations of the HSE in the mantle are considerably higher than expected from metal-silicate equilibration at conditions perceived for the formation of Earth, and several hypotheses, reviewed and critically evaluated by Walker (2009), have been suggested to explain this phenomenon. These hypotheses embrace (1) incomplete core separation with some HSE-rich metal retained in the mantle, (2) greatly reduced metal-silicate fractionation due to elevated sulfur content in the segregating metal or elevated pressure-temperature conditions at the base of a deep magma ocean, and (3) continued accretion to Earth of extraterrestrial material with chondritic abundances of HSE after core-mantle separation. As discussed by Walker (2009), all of these hypotheses have shortcomings and none of them in isolation can account for the HSE distribution in Earth, Moon, and Mars. Instead, a hybrid model is favored that includes the combination of continuous or periodic metal-silicate segregation over the first 100 m.y. with continuous addition of HSE by late accretion of cosmic material. Maier et al. (2009) noted an increase in PGE contents of komatites from the Paleoarchean to the Neoarchean and Protorezoic and interpreted this as evidence of almost complete PGE depletion in the lower mantle after core formation. This was followed by HSE addition to the lower mantle after extraterrestrial material from 4.5 to 3.8 Ga meteorite bombardment had sunk into, and mixed with, the lower mantle. According to their hypothesis, contamination of the mantle with HSE commenced at 3.2 Ga and was largely completed by 2.9 Ga, thus overlapping in time with the postulated Mesoarchean giant gold episode.

Even today the mantle is not homogeneous within a given depth zone as indicated by seismic tomography (Burke et al., 2008; Mosca et al., 2012). Such mantle heterogeneity has existed most likely from at least the Mesoarchean to present time. Mantle domains enriched in HSE, whether due to incomplete or inefficient core formation and/or addition of cosmic material, could have provided suitable source regions for Au-enriched melts. The Kaapvaal craton, together with the Pilbara craton, represents the oldest well-preserved succession of Archean crust, much of which formed over the time period for which elevated transfer of Au from the mantle into the crust can be assumed. It is probably not a coincidence that the same cratonic domain that hosts the world's largest accumulation of gold is also host to the world's largest concentration of PGEs in the form of the Bushveld Layered Intrusive Complex. The latter currently contributes about 54% of global PGE production and reported resources constitute close to 70% of the global PGE resources (RMG, Version 2014.01.14, © Intierra Raw Materials Group). Although the emplacement of the Bushveld ultrabasic and basic melts is considerably younger, 2054 ± 1 Ma (Scoates and Friedman, 2008), these melts might have tapped a similar mantle domain as the earlier komatiitic melts in the Mesoarchean. Thus, the anomalous HSE endowment of the Kaapvaal craton might be due to a dynamically emplaced Mesoarchean mantle heterogeneity that remained in place at least until the time of Bushveld emplacement.

Conclusions and Consequences for Future Exploration

Comparison between the various conglomerate-hosted gold deposits and/or occurrences currently known across the world shows important similarities, but also systematic differences, all of which make it possible to develop a holistic view on the genesis of this type of deposit. Moreover, it addresses the enigmatic question of gold source and provides an explanation as to why the Kaapvaal craton appears relatively more endowed in gold than any other region. Specific inferences that can be drawn from this study are as follows:

1. The style of mineralization is independent of metamorphic and/or orogenic overprint, and the ore mineral paragenesis is mainly controlled by paleoenvironmental constraints, such as composition of the atmosphere.

2. Witwatersrand-type mineralization occurred repeatedly throughout the Mesoarchean to the Paleoproterozoic from at least 3.1 to 1.9 Ga, excluding the Tertiary example of the Waimumu district in southern New Zealand, and in a range of different tectonic settings.

3. All Witwatersrand-type deposits and/or occurrences, with the exception of the Waimumu district, had a hinterland at the time of sediment deposition that was dominated by Archean or Paleoproterozoic greenstone belts. Those with an Archean hinterland are better endowed.

4. A purely epigenetic model, whether in the sense of orogenic-type gold or any other form of postdepositional introduction of gold, and other critical ore constituents, such as pyrite, uranium, and carbonaceous matter, contradicts the spatial and temporal distribution of the deposits as well as the abundance of sedimentological, geochemical, mineral chemical, textural, and geochronological data available that can only be explained by syndepositional introduction of gold, together with detrital pyrite and uraninite, into the host conglomerates.

5. Diagenesis, postdepositional hydrothermal overprint along shear zones, and/or metamorphism led to variable intensity mobilization of some of the ore components, including gold, which in most cases took place only at a small, micro- to decimeter scale. In some cases, such as in some of the Witwatersrand reefs or the Jacobina deposits in Brazil, postmineralization overprinting almost completely obliterated original detrital features.

6. Detrital gold in post-Archean Witwatersrand-type deposits and/or occurrences can be traced back to specific point sources, typically in the form of orogenic vein-type deposits in the hinterland, or to reworked older, Archean paleoplacer deposits. In contrast, detrital gold in the exceptionally endowed Mesoarchean placer deposits was largely derived from the sedimentary reworking of fine-grained gold particles interpreted to have been entrapped originally on the surface of microbial mats.

7. The composition of the Archean atmosphere and hydrosphere expedited chemical weathering and a high fluvial Au
run-off from the Archean land-based crustal regions. Prolif-
eration of early aerobic life forms (probably cyanobacteria) in Mesoarchean wetlands (flood plains and shallow coastal regions) made it possible for Au that was dissolved in river waters to be trapped on the surface of microbial mats.

8. Maximum addition of gold into the Mesoarchean (ca. 3.0 Ga) crust coincided in time with maximum mantle tem-
peratures and thus maximum degrees of partial melting in the mantle. Heterogeneities in the Mesoarchean mantle with domains enriched in HSE are suggested and might explain higher background Au values in juvenile Mesoarchean crust.

In combination, all of these individual inferences lead to the conclusion that the Mesoarchean Era, for a multitude of reasons, was by far the best time for concentrating, or at least preconcentrating, gold in the crust. This has major implica-
tions for future exploration efforts in the search for new Witwatersrand-type deposits. Any fluvial to fluviodeltaic con-
glomerate succession that obtained its detritus largely from Archean crust and/or younger hinterland that hosts orogenic gold deposits has the potential for Witwatersrand-type miner-
alization. This potential increases radically if the source region is Mesoarchean in age and if the tectonic setting is such that repeated sediment recycling of the underlying strata could upgrade the heavy mineral concentrations on top of angular unconformities and erosion surfaces. Foreland basins are particularly advantageous in this regard. Younger deposits are likely to yield only small amounts of gold, comparable to mod-
ern placer deposits. In contrast, Meso- and Neoarchean con-
glomerates, in which microbially mediated syngenetic gold accumulations over larger areas became reworked, have the potential of containing giant deposits. No currently economic Witwatersrand-type deposit with a source younger than 2.0 Ga is known. Thus, the main limiting factors for the size of a Witwatersrand-type deposits are (1) age of the conglomerate beds, (2) age of the sediment source region, (3) overall sedi-
ment thickness and more importantly the number of angular unconformities beneath and within the siliciclastic sediment pile, (4) sedimentary facies suitable for heavy mineral concen-
tration, and (5) preservation of an ancient sediment succession, which is strongly aided by underlying buoyant Archean subcontinental lithospheric mantle (Groves et al., 2005a).

Whether the primary gold endowment of the Archean crust was more or less uniform across the various cratons or whether some cratons became significantly better endowed because of lateral heterogeneities in the Archean mantle and a subsequent tectonic history that helped preserve the deposits is still speculative. The latter is suspected in the view of the known worldwide distribution of gold and PGE. This would leave the Kaapvaal craton and possibly the Pilbara craton as the best sites for (super-) giant Witwatersrand-type deposits. It does not reduce, however, the potential for additional eco-
nomic deposits of the same style on other cratons. Current exploration efforts in conglomerates of the Huronian Super-
group (Superior province, Canada) or the Fortescue Group (Pilbara craton, Western Australia) are promising indications that more Witwatersrand-type deposits will be discovered in the years to come. However, the probability of discovering a new supergiant cluster of deposits, such as in the richest Wit-
watersrand goldfields, is considered remote.

Efforts to drag the dwindling gold mining industry in South Africa out of its current dilemma should depend, therefore, on new and innovative ways of recovering the gold from depths beyond 4 km, rather than on exploration for major new deposits or goldfields. After all, there is an estimated resource of not less than 44,000 t Au reported to be still in the ground. Conventional mining methods will fail to recover much of this resource, and the challenge is to develop (rapidly!) alternative, more effective methods, possibly based on in situ leach-
ing techniques, in order to fully utilize what appears to be a unique concentration of gold in the Earth’s crust.

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