3.5-Ga hydrothermal fields and diamictites in the Barberton Greenstone Belt—Paleoarchean crust in cold environments

Maarten J. de Wit1*† and Harald Furnes2†

Estimates of ocean temperatures on Earth 3.5 billion years ago (Ga) range between 26° and 85°C. We present new data from 3.47- to 3.43-Ga volcanic rocks and cherts in South Africa suggesting that these temperatures reflect mixing of hot hydrothermal fluids with cold marine and terrestrial waters. We describe fossil hydrothermal pipes that formed at ~200°C on the sea floor >2 km below sea level. This ocean floor was uplifted tectonically to sea level where a subaerial hydrothermal system was active at 30° to 270°C. We also describe shallow-water glacial diamictites and diagenetic sulfate mineral growth in abyssal muds. These new observations reveal that both hydrothermal systems operated in relatively cold environments and that Earth’s surface temperatures in the early Archean were similar to those in more recent times.

INTRODUCTION

Climate variability and paleoenvironmental conditions on early Earth are difficult to quantify. Speculations about climate on early Earth are based on models that trade off elevated concentrations of greenhouse gases for lower solar radiation from an approximately 23% fainter Sun (1–4), as well as on geological observations that record the presence of oceans as far back as 3.8 billion years ago (Ga) (5–7) and the presence of surface water by 4.3 to 4.4 Ga (8). In addition, chemical data derived from some of Earth’s oldest well-preserved sedimentary rocks have revealed insights into possible temperature changes in marine waters over deep time. Specifically, oxygen isotope measurements of ca. 3.5-Ga rocks have been used to estimate the temperature range of Earth’s hydrosphere to be between 26° and 85°C (9–13), suggesting a much greater concentration of greenhouse gases, including significant water vapor, in the atmosphere at that time. However, there is controversy about the interpretations of the data at hand, in particular those related to the isotope variability of seawater over deep time [for example, (7, 14–16)]. Here, we present new field observations and oxygen isotope data from greenstones and cherts from the Barberton Greenstone Belt (BGB) in South Africa that shed light on the physical setting and character of Archean sea-floor hydrothermal systems and their contribution to the geological and geochemical signatures implicated in these controversies.

The BGB in the Makhonjwa Mountains of South Africa (Fig. 1, inset map) provides reliable rock-based benchmarks for early Earth studies because it hosts some of Earth’s oldest, best-preserved sedimentary and volcanic sequences, including those interpreted as fragments of oceanic crust (17–21). The Onverwacht Suite (OS) (Fig. 1) from which lava of peridotitic composition was first described is a tectonometamorphic sequence that predominantly consists of basaltic lavas with a preponderance of ultramafics in the lower parts and more felsic types toward the top. Fine-grained silicified pyroclastic and sedimentary units, in the form of cherts, occur throughout the OS, becoming more frequent upwards and dominating its upper parts (17, 18, 21). We present geological and geochemical evidence from the OS suggesting that hot fluids from both different types of hydrothermal systems—operating in deep oceans and shallow seas, respectively—discharged into cold waters. Because these processes operated at low latitudes (20° to 40°) (23), our new data imply that the temperatures of the global oceans and atmosphere were relatively cool at ca. 3.5 Ga. These new observations described below have implications for reconstructing early Earth’s climate and atmospheric composition.

We first summarize evidence for deep-water hydrothermal systems that altered the basaltic igneous rocks and volcanic ash of the Hooggenoeg Complex (HC) and the Kromberg Complex (KC) within the OS. Both the HC and the KC represent the upper oceanic crust formed in water depths of >2 km (Figs. 1 to 3, figs. S1 and S2, and tables S1 and S2). The HC is an approximately 2.7-km-thick volcanic basalt sequence subdivided into nine major eruptive phases, each from ~100 to 700 m in thickness and each separated by 1 to 15 m of chert representing silicified volcaniclastic muds and air-fall ash. The entire sequence formed within 5 Ma (3467 to 3472 Ma) around three volcanic centers inferred to reside at oceanic depths between 2 and 4 km [(18, 19); Figs. 1 and 2] and from which we describe newly discovered massive pipes of a fossil hydrothermal field that operated below a depth of 2 km [(19); Fig. 1] at temperatures deducible from the new oxygen isotope data presented here.

Second, we describe a pervasive hydrothermal system that operated during the deposition of a siliciclastic and felsic volcanic succession of the Noisy Complex (NC), accumulating close to sea level some 20 to 30 Ma later between 3458 and 3432 Ma (17, 21, 24–26). The lower part of the NC (Fig. 1 and figs. S2 and S3) comprises a regionally diachronous sequence of up to 2-km-thick lacustrine and littoral marine facies, fluvial conglomerates, sandstones, and diamictites that in the type area lie with profound unconformity directly on previously weathered pillow basalts of the HC (17, 18, 27). The unconformity represents the ~30 Ma between 3467 and 3434 Ma, during which the HC was tectonically uplifted from oceanic depths of ca. 2 to 4 km to elevations above sea level, where it became deeply incised (>1.5 km; for details, see geological map in Fig. 4 of de Wit et al. (18)] and buried by the fluvial conglomerates and diamictites of the lower NC [(18, 27); Figs. 1 and 2].

The uppermost part of the NC comprises a 0.5- to 1-km succession of silicified shallow-water sandstones and shales (cherts) and oxide

---

1Aeon and Earth Stewardship Science Research Institute, Nelson Mandela Metropolitan University 7701, Port Elizabeth 6031, South Africa.
2Department of Earth Science and Center for Geobiology, University of Bergen, Allegt. 41, Bergen 5007, Norway.
*Corresponding author. E-mail: Maarten.deWit@nmu.ac.za
†These authors contributed equally to this work.
Fig. 1. Volcanotectonic and chemical stratigraphy of the upper part of the OS of the BGB, with estimates of water depth. The left column shows tectonostratigraphy at the main shear zones (SZ; purple), and unconformity separating the HC (dark green; with the 10 main chert horizons HC1 to HC10 in dark gray) from its underlying complex (Komati Complex; not marked), and the NC (yellow) from the KC (pale green) and the MC (orange), together with selected radiometric ages from the sequences [modified from de Wit et al. (18)]. Note that the HC is separated from the NC (yellow) by a major unconformity that cuts down from above HC10 to below HC6 [into eruption phase 5 (EP5)]. The two central columns show the $\delta^{18}$O and SiO$_2$ values of the mafic volcanic rocks and of a few intrusive rocks of the HC and KC, as well as 6 of the 10 chert horizons of the HC (HC1 to HC10) that separate its nine eruption phases [EP1 to EP9, as defined in Furnes et al. (20); see also Fig. 2]. Note how the $\delta^{18}$O values in EP1 to EP4 and EP9 trend toward mantle values above and below their respective bounding chert bars (see text for further explanation). The $\delta^{18}$O and SiO$_2$ values of the hydrothermal field site (on HC2), including the material of the hydrothermal pipes (red triangles), are also shown. For detailed sample sites, field relationships, geochemical data, and temperature calculations, see Figs. 2 to 5, tables S1 and S2, and figs. S1 to S3.
Fig. 2. Reconstruction of the volcanic development of the HC with average $\delta^{18}O$ values for chert horizons. Volcanic architecture of the ca. 2.7-km-thick HC reconstructed from the base (above the Middle Marker; HC1) to the highest preserved chert (HC10; below the Etimambeni section of the NC). The magma types of the nine eruptive phases (EP1 through EP9) shown on the two right-hand columns are derived from Furnes et al. (20). The eruption depths of EP1 to EP9 increased from about 2.0 to 4.0 km below sea level [as calculated from the vesicularity of their respective pillow lavas and as described in detail elsewhere (19); see also Fig. 1]. The location of the hydrothermal field site (built on HC2 at the bottom of EP2) is shown as a red framed box. The average $\delta^{18}O$ values for 6 of the 10 chert layers (HC1 through HC10) are highlighted (from table S1), emphasizing the gradual decrease in the $\delta^{18}O$ values of the cherts (from top to bottom) from 18.0 to 15.3‰, corresponding to a temperature increase from about 80° to 170°C (Supplementary Materials). The number of chert analyses of HC1 through HC10 (tables S1 and S2) and the standard deviations are as follows: HC1 (three analyses; SD, 0.47), HC2 (eight analyses; SD, 0.60), HC3 (three analyses; SD, 0.40), HC4 (three analyses; SD, 0.62), HC6 (one analysis), and HC10 (five analyses; SD, 0.78). HV, HV1 through HV15 refer to the 15 measured volcanic sections shown in fig. S1.
facies banded-iron formation underlain by up to 1 km of felsic lavas and porphyries (17, 18, 21, 24–26, 28). The felsic magmatism lasted about 17 Ma between 3462 and 3445 Ma (17–20, 23, 25). Abundant oxygen isotope analyses from the rocks of this complex have been published elsewhere (9, 11–13), supplemented here by new data from directly above the unconformity (Fig. 1 and table S1).

The two different tectonic environments of these hydrothermal systems fuelled by deep oceanic magmatism (the HC) and by shallow continental magmatism (the NC) are clearly distinguished by geological (field) and geochemical data (17, 18, 25, 29–32). However, previous oxygen isotope studies of the BGB have not distinguished between chert samples collected from these two different systems when evaluating their δ18O signatures to derive seawater temperatures [for example, (9, 11–13)]. In addition, in earlier studies of the BGB, the cherts of these Archean rocks were interpreted as primary precipitates from seawater, and temperatures derived from related oxygen isotope analyses were thus interpreted to represent directly the prevailing seawater temperatures (9, 11–13). Subsequently, the cherts were found to be formed predominantly from igneous and sedimentary rocks during secondary silification processes [for example, (21, 30)]. Further, preservation of primary silica precipitates is rare [for example, (25, 33)]. The new field observations thus prompt us to reevaluate these earlier interpretations of early Archean environments with greater attention to precise locations within stratigraphic sequences.

In addition, new field observations from the shallow-water environments of the NC include diamicites, varves, and dropstones that we interpret as glaciogenic deposits. In the HC deep-water cherts, we identify sulfate (gypsum) pseudomorphs, which we interpret as diageneric minerals grown during interaction between hydrothermal fluids and cold seawater. Together, these geological field observations and geochemical data shed new light on the longstanding controversies over Archean environments.

**RESULTS AND DISCUSSION**

**Field evidence for deep-water volcanic sections and hydrothermal pipes**

The deep-water hydrothermal systems are recorded in the mafic sequences and associated cherts of the OS. All cherts in the HC and KC are silicified mafic-ultramafic volcanoclastic rocks that directly overlie pillow lavas and massive flows, or less frequently are intruded by mafic sills [for example, (18, 21, 28, 34); Figs. 1 and 2 and fig. S4].

The volcanic reconstruction of the HC [Fig. 2; (19)] shows the location of its 10 major chert horizons, with the average δ18O of 6 cherts also indicated. Also shown (Fig. 1) are the estimated water depths of the HC and the KC (2 to 4 km), as calculated from the vesicularity of pillow lavas eruption in a water depth of >1 km. Although differences in temperature, major element composition, and volatile content may all influence vesicularity, we assume here that these parameters are similar to the Hawaiian case and are also relatively consistent among the different basaltic units within the BGB sequences such that information on relative depths is preserved. For tholeiitic pillows near Hawaii, which are comparable to the geochemical composition of most pillow lavas of the HC and KC of the BGB, vesicularity is near zero at a depth of about 3 km but is about 5% at a depth of 1.7 km (35). The vesicularities of the pillow lavas of the HC and KC are of that order (19); hence, the above depth range has been adopted here without modification. Details of the BGB pillow shapes and sizes and their vesicle densities are described elsewhere (19).

We have discovered remnants of a hydrothermal field (Figs. 1 and 2) comprising three white and green silicic pipe-like structures (SiO2 > 90%), hereafter referred to as pipes. Each pipe is about 14 m tall, protruding from the second chert layer (HC2) of the HC. Two of the pipes are well exposed, spaced 14 m apart along the ca. 1-m-thick subvertical chert (HC2) that can be traced along strike for hundreds of meters away from the pipes (Fig. 3, A to E, and fig. S5). The pipes must have extended vertically above the chert into the overlying basalt before the OS sequences were tectonically rotated by 90°. Both pipes are about 5 m wide, and both are rooted in brecciated sections of the chert, from which they protrude 14 m vertically up through the overlying pillow lavas (Figs. 3 and 4 and fig. S5). A third pipe of similar height, but less well exposed, occurs 28 m farther on. The pipes consist almost exclusively of fine-grained white quartz and foliated zones of bright-green fuchsite (Cr- and Rb-rich mica) and white

---

**Fig. 3. Field location and structures of the hydrothermal field showing two pipes extending from regional chert HC2.** (A to C) Continuous zoom-in shots from a helicopter; Global Positioning System coordinates are rounded to the nearest minute (25°59'S, 30°59'E). Younging directions are indicated by an up-arrow on all images. (A and B) General aerial helicopter view along the white chert layer HC2 (about 1 m thick) looking westward, uphill toward two pipes exposed near the hilltop (a third pipe is exposed but out of view on the far downslope). HC2 and the underlying and overlying lavas of EP1 and EP2, respectively, are in a subvertical position, providing a near-perfect cross section of the sequences. (C) Close-up view of the hydrothermal pipe area showing the two investigated pipes A and B. Both pipes project about 14 m away (upward) from HC2. (D and E) Upward views along pipes A and B illustrating their near-bimodal composition of quartz (white-beige) and chrome-mica (fuchsite; green). Note that the pale-brown color of EP1 directly below HC2 is due to its pervasive silicification, as can be seen from its high SiO2 values in Fig. 1.
I. Field relationships and components of hydrothermal pipes A and B

II. Schematic interpretation

Fig. 4. Field relationships of two silica pipes from the hydrothermal field, with interpretation of their formation and preservation. Schematic representation of silica-fuchsite pipes from the hydrothermal field (for location, see Figs. 1 and 2). (I) Field relationships and components of the hydrothermal field, with the pipes rooted in the chert horizon HC2 at the boundary between eruptive phases EP1 and EP2 of the HC. (II) A schematic interpretation of the development of the pipes through several time stages. At time $T_0$, HC2 appears as a volcaniclastic deposit on top of EP1. At a later stage ($T_{0.5}$), early silicification starts to convert the volcaniclastic rocks of HC2 into chert ($T_2$). Subsequent to the lava deposits of EP2 on HC2, brecciation and focused silica-potassium–rich hydrothermal fluids along fractures develop the hydrothermal pipes until they become inactive at time $T_4$, when buried beneath the next lava sequence of EP2. For detailed sample sites, field relationships, and chemical relationships, see figs. S4 to S9; for chemical data, see tables S1 and S2.
quartz veins, and are surrounded by less-altered brown-green pillows of the overlying lava sequence (20). In places, the pipes contain silicified remnants of pillows with lava-withdrawal structures (fig. S5A) and well-preserved silicified skeletal pseudomorphs of purported olivine and plagioclase (fig. S5B), suggesting that the pipes are vertical columns through in situ lava flows. The pillow lavas of EP1, beneath HC2, are variably silicified to a depth of 5 to 10 m (pale brown in Fig. 3A) and are cut by a stockwork of translucent black-white quartz and dikes that terminate within HC2. The overlying pillow flows of EP2 are relatively unaltered (20). We interpret the pipes as silicified columns formed by focused fluid percolation through the ca. 14-m-thick lava sequence, ultimately fed through the stockwork system below HC2 (Fig. 4). The brecciation of the chert at the foot of pipe B (fig. S5E) most probably occurred during vigorous, focused fluid flow.

All of the rocks of the hydrothermal field area are strongly altered, and the minerals of the pipes, the chert (CH2), and the lavas underneath (EP1), adjacent to and above the pipes (EP2) (Fig. 3), were determined by x-ray diffraction. In EP1, the main minerals of the lavas are quartz and illite (occurring in all samples); about half of the samples contain the mica minerals muscovite and lepidolite (?) and the aluminium silicate hydroxides pyrophyllite and kaolinite (?); and clinochlore (a member of the chlorite group) has been identified in the lavas of EP2 (table S2). The main mineral of the chert samples (HC2) is quartz; illite is common, and muscovite has been identified in one sample (table S2). We interpret the formation of the cherts and pipes within the HC and KC mafic-ultramafic sequences as remnants of fossil hydrothermal fields, similar to those found in the modern ocean crust (for example, (36)).

Shallow water/subaerial hydrothermal system: Field and fluid inclusion studies

In contrast with the above, the hydrothermal system recorded in shallow-water–to–terrestrial environments is associated with felsic volcanic sequences and related cherts of the NC (17, 18, 24–26). The subaerial to shallow-water depth of the NC is based on sedimentological field analyses of many previous workers (9, 17, 21, 27, 28, 33, 37) and is substantiated by the coarse well-rounded fluvial conglomerate and diamictite units described from the Etimambeni section [fig. S3; (18, 22, 27, 28)]. The NC (Fig. 1) is a 1- to 4-km silicified volcanosedimentary and porphyritic sequence intruded by 3.45- to 3.46-Ga tonalitic to trondhjemitic sills and dykes that also cut through the underlying HC and are rooted in tonalitic plutons below the BGB that can be linked to geochemically identical granitoid magmas of overlapping age that intruded the lower parts of the OS some 6 to 8 km below the NC (9, 17, 18, 24–26). In large sections, the upper fine siliciclastics were hydrothermally altered and silicified to cherts during and immediately after their deposition and cut by dense networks of veins, dykes, and sills of black to white cherts (9, 17, 25, 28, 33).

Detailed fluid inclusion analysis (25) reveals that the hydrothermally deposited quartz contains coeval two-phase aqueous inclusions (with variable salinities) and mixed H2O-CO2 inclusions that can be interpreted to be at least partly of felsic magmatic hydrothermal origin (25). Microthermometry shows that unmixing of high-salinity and low-salinity aqueous solutions from mixed H2O-CO2 systems in the inclusions occurred at a relatively low pressure (ca. 100 bars) at temperatures ranging between 100° and 200°C, consistent with the geological field evidence for shallow-water deposition of the sediments and their near-contemporaneous subsurface hydrothermal alteration (25). On the other hand, the homogeneity of the H2O-CO2 inclusions implies their original trapping at a temperature of ca. 270°C at an internal pressure of ca. 2 kbar (25). This suggests that these fluids were issued from a deep magmatic geothermal source along a network of interconnected channels to the NC. This is consistent with geological observations of a network of felsic sills and dykes connecting the NC to granitic plutons (24–26). Thus, the NC cherts likely formed from fluids rooted in a deep granitic source that interacted nearer the surface with relatively cool downward-percolating meteoric and shallow marine waters. This resulted in silicification of the NC sediments and formation of primary surface silica (and iron) deposits at water depths below 70 m and at temperatures between 30° and 120°C, as determined independently from δ18O (9, 11–13, 33) and from extensive fluid inclusions analyses (25, 38, 39). This second hydrothermal system thus resembles terrestrial–to–shallow-water hydromagmatic systems observed in, for example, New Zealand, Japan, and Mexico (17, 25, 31, 38, 39).

Oxygen isotope and bulk-rock chemical signatures of hydrothermal alteration

The δ18O signatures of the deep-water sequences (HC and KC combined), including those from the Mendon Complex (MC) (not described in detail here; Fig. 1), are broadly similar to those recorded in modern and ancient ophiolites (see histograms in figs. S6 and S7). Figure 1 shows the relatively good sympathetic relationship between the δ18O and the SiO2 of metabasalts throughout the igneous sequences at all stratigraphic levels of the different complexes of the OS (table S1). Within the eruptive sequences, both δ18O and SiO2 increase from the primary igneous values of +5.5 to +6.0 per mille (‰) at between 20 and 250 m below their chert caps (for example, EP3 and EP2, respectively; Fig. 1) to the SiO2-enriched chert caps that record δ18O values of between +13 and +20‰, similar to those of the HC and KC regional cherts recorded previously [fig. S6C; (9, 29, 30, 32)]. This positive correlation has been recorded elsewhere and is interpreted as a result of enhanced SiO2 precipitation from cooling fluids at lower temperatures (32).

Within the hydrothermal pipes, the δ18O values of the altered basalts and cherts show remarkably systematic correlations with major and trace element compositions (for example, SiO2, Al2O3, Cr, and Rb; Figs. 1 and 2, figs. S8 to S11, and tables S1 and S2). Furthermore, the pipes have a simple mineral assemblage of dominant quartz and fuchsite, with minor pyrophyllite, illite, and kaolinite. The correlations can be explained by the conversion of basalt into a bimineral assemblage of varying proportions of quartz and fuchsite, each with approximately uniform δ18O, suggesting a relatively constant temperature at formation.

The range of δ18O and silica within the pillow lava sequences and the overlying regional cherts is greater, and the correlation between them is more diffuse than that within the pipes (Figs. 1 and 2 and figs. S10 and S11). Although the stratigraphic chemical pattern of each eruptive cycle is distinct, δ18O values in eruptive phases of the HC decrease consistently from their chert caps downward to attain mantle values over a relatively short distance. In the volcanic units EP1 and EP2, directly below and above the hydrothermal pipes that we calculated to have formed at an oceanic depth of some 2 km (Fig. 1), mantle values are reached about 250 m below their respective overlying cherts (HC2 and HC3). This is consistent with model predictions stating that hydrothermal penetration depths close to spreading ridges should be less than 600 m below ocean floor exposed above ca. 2.4 km (16).
Eruptive sequences overlying the cherts capping earlier volcanic cycles also show δ¹⁸O decreasing rapidly upward from the chert horizons. This suggests that the eruptive sequences harbored separate complex and diffuse hydrothermal convection systems, perhaps driven by internal magma pulses away from the volcanic centers, decreasing in temperatures both up and down from within the eruptive sequences. Thin sediment veneers overlying lava sequences were silicified during more diffuse fluid percolation at progressively lower temperatures. We speculate that diffuse convective circulation may have terminated when silica precipitation in the cherts sealed the cap, further restricting the hydrothermal fluid circulation beneath the eruptive units and/or breaking through fractures, creating focused fluid flow (pipes) to influence circulation systems in freshly overlying eruptive sequences (for example, Fig. 4 and figs. S4 and S5).

**Oxygen isotope evidence for temperatures of hydrothermal systems and ambient ocean water**

Most previous oxygen isotope studies of the BGB for which elevated seawater temperatures were derived are from samples collected from the NC rocks [for example, (9, 11–13, 33)]. The temperatures of silicification were from detailed oxygen isotope and fluid inclusion work to range from at least 30° to 200°C [(9–11, 33)] and Supplementary Materials. Our oxygen isotope data are broadly consistent with previous measurements of similar rocks from the HC 

The temperatures of the silica-water interaction for the formation of the silicified shallow-water sequences were determined from detailed oxygen isotope and fluid inclusion work to range from at least 30° to 200°C [(9, 25, 33, 38, 39) and Supplementary Materials].

The temperatures of the silica-water interaction for the formation of the silicified metabasalts and regional cherts of the deep-water sequences can be approximated from their δ¹⁸O by applying the geothermometer of Knauth and Epstein (40) and by assuming δ¹⁸O seawater to be equal to 0 ± 1‰ [for example, (9, 15, 39)]. Although assuming δ¹⁸O seawater to be equal to 0 ± 1‰ (for example, similar to that of the present day) has been considered controversial [for example, (7, 14–16)], this is discussed and justified in Section C of the Supplementary Materials. Our oxygen isotope data are broadly consistent with previous measurements of similar rocks from the HC [for example, (9)]; however, because the sample locations from these earlier studies are not precisely identified, they are not included in our analyses.

The average δ¹⁸O values for six of the regional chert caps are shown in Fig. 2. There is a gradual downward decrease in their average δ¹⁸O values from 18.0‰ at the top of the HC (HC10; deposited at a depth of ca. 4 km) to 15.2‰ (HC4; deposited at a depth of ca. 3 km). For two of the volcanic sections of HC (EP9 and EP2), the silicification of the basalts is extensive, and the SiO₂ and δ¹⁸O values of these two sections define good positive relationships. When the SiO₂ values of these two sections are extrapolated to 100%, the calculated temperatures increase from around 80° to 170°C, respectively (Figs. 1 and 5 and figs. S10 and S11).

The temperature of alteration fluid in the pipes (Fig. 4 and fig. S5) can be calculated to be about 200°C, derived by comparing the assumed equilibrium fractionation of δ¹⁸O/δ¹⁴O between quartz and fuchsite (with measured δ¹⁸O values of 14.4 and 8.4‰, respectively) with the experimental calibration of quartz-muscovite fractionation factors (41–43). The fluid driving this alteration has a calculated value of about 0‰ (Fig. 5 and Section C of the Supplementary Materials). We do not know the precise Rb or K content of Archean seawater, but fluid inclusion analyses of hydrothermal minerals in the KC suggest that seawater salinity and chemistry were likely rather similar to the present (38, 39). Either way, the relatively high enrichment of Rb and the precipitation of fuchsite in the pipes indicate a large focused flux of hot seawater-derived Si-rich fluid emanating close to the ocean bottom surface at

---

**Fig. 5. **Al₂O₃ and SiO₂ versus δ¹⁸O relationships of metabasalt and chert samples from the hydrothermal field. Bimodal quartz-fuchsite samples from the hydrothermal pipes and adjacent rocks were used to illustrate the excellent sympathetic relationships between Al₂O₃, SiO₂, and δ¹⁸O (‰) [Standard Mean Ocean Water (SMOW)], and the end-member δ¹⁸O values of quartz (14.4‰) and of the chrome-rich muscovite fuchsite (8.4‰) were used to calculate a temperature range between 170° and 250°C for the fluid in equilibrium with pipe mineralogy, assuming a δ¹⁸O value between –1 and +1‰ for Archean seawater, and a temperature of about 200°C when the δ¹⁸O value of Archean seawater is equal to SMOW (see Supplementary Materials for details).
temperatures up to at least 200°C, until covered by the subsequent lava flow (Fig. 4). Because the likelihood of exposing such vents in vertical geological sections (for example, normal to the paleohorizontal) is small (for example, Fig. 2), their occurrence in our sections suggests that such hydrothermal venting was commonplace between 3.46 and 3.47 Ga, as previously advocated (29, 30, 44). In the following section, we discuss evidence that both the deep-water and the shallow-water hydrothermal systems discharged into cold water environments.

Geological evidence for shallow-water, low-latitude glacial conditions: Polymictic diamicrites and varves with dropstones in the NC

In the NC-type section at Etimambeni, the lower 100 m of the sequence is dominated by 1- to 20-m-thick crudely stratified units of poorly exposed polymictic diamicrites with clasts ranging from boulders (>50 cm) to millimeter-sized fragments set in a fine groundmass of angular grains (18, 27, 28; Fig. 6 and fig. S3). Clasts of all sizes are mostly angular, are very poorly sorted, and range from matrix-supported to clast-supported. Petrographic textures of matrix minerals such as feldspar, quartz, and zircon are angular, indicating relatively immature nature and impact fracturing (27). Occasional well-rounded pebbles suggest fluvial rounding before redeposition in the diamicites. Clasts and matrix minerals have undergone a pervasive alteration before and after deposition, which has changed much of their primary matrix mineralogy into quartz, sericite, and carbonate (23). The clasts are dacitic to granitic, with lesser metabasalt, metabasalt, and metasedimentary rocks (including reworked conglomerates and various cherts). Although many of the clasts were likely derived...
from the underlying HC, occasional rounded granitic clasts and detrital zircon grains also suggest fluvial transport from more distal sources. Such a complex, varied source is supported by a wide range of ages between ca. 3350 and 3600 Ma, determined from the detrital zircons in this sequence, the older of which must have been derived from basement granitoids, indicative of large regional transport (27).

Although these deposits have been interpreted as volcanic breccias in the past (28), they were more recently referred to as diamicrites following detailed descriptions from a continuous drill core through the entire sequence (18, 27). In addition, we have recently discovered that these diamicrites are locally interbedded with rhythmically laminated sections (<50 cm thick) comprising millimeter-scale couplets of pale sandy layers and dark mud layers that resemble varves (Fig. 6A). Occasionally, small isolated clasts occur in the rhythmites, whose laminations are deflected downward beneath the top of the clasts, resembling deflections around glacial dropstones (Fig. 6B and C). Although we acknowledge that such clasts might have been derived from volcanic explosions, the clasts are not volcanic in origin, and their lithologic diversity suggests origins in a variety of geological settings. Thus, we prefer to interpret them as glaciogenic dropstones deposited in glacial varves.

The diamicites are abruptly overlain by a sequence of boulder conglomerates and, in turn, by graded bedded sequences of sandstones, with angular matrix minerals that display slumping, convolute bedding, and flame structures. These have been previously interpreted as turbidites (27, 28, 37), but such features indicative of liquefaction are also commonly associated with tillites and other glaciological deposits.

The sedimentary structures in the lower Etimambeni section (Fig. S3) are consistent with glacially related deposits (diamicton assemblages), suggesting fluvioglacial sorting and fluidization processes in bedded sediments of periglacial environments, with occasional lake deposits (cf. (45, 46)). The angular nature of the lowermost diamictons suggests a supraglacial origin [for example, originating on glacier surfaces; (47)], consistent with our interpreted “land system” for the Etimambeni sequence. Although no striated pebbles have been found, these new observations suggest that the sequence may have been deposited in subaerial glacial environments.

A section sampled from the drill core across the HC–NC unconformity reveals a distinct shift of δ¹⁸O values from −16 to −19‰ in pillow lavas, typically seen directly below the HC cherts, to −13 to −14‰ in clasts, siliciclastic matrix, and interlayered cherts of the NC silicified diamicites directly overlying the unconformity (Fig. 1 and table S1). Because the diamicites represent shallow-water–to–terrestrial deposition, a direct interaction with Archean meteoric water and/or possible glacial melt water during silicification was likely, which may account for the relatively high δ¹⁸O values in these silicified rocks [for example, (15)].

Evidence for cold deep-ocean water: Gypsum pseudomorphs in abyssal cherts of the HC

The upper parts of most of the deep-water HC cherts comprise fine-grained silicified mudstones with silicified pseudomorphs after sulfates (interpreted as gypsum; Fig. 7) (18, 48). Some studies [for example, (49)] argue that there are no known examples of Archean sulfates because the concentrations of sulfate and calcium ions in the deep Archean oceans must have been low as a result of anoxic conditions. In the BGB, however, sulfates occur in the weathering zones of underlying altered ultramafic lavas (50) and are common in barite-bearing oxide-facies jasplilites of the NC and as clastic barite in sediments immediately overlying the KC (21, 31, 39, 51). HC cherts that contain pseudomorphs after the purported gypsum are enriched in K-feldspar, calcite, dolomite, barite, chlorite, and iron oxides (34). Thus, sulfate and calcium ions could have been liberated via hydrothermal reactions within the sediments that likely controlled diagenetic reactions during the diffuse hydrothermal percolation through the porous deep-ocean volcanoclastic muds before and during silicification (29, 30, 32, 34). In deep water, gypsum growth would have required temperatures below 42°C because the gypsum-to-anhydrite transition temperature is ~42°C at the surface in pure H₂O, and this transition is lowered by increased salinity and pressure (52, 53). The presence of possible pseudomorphs after gypsum in the deep sea-floor sediments of HC is inconsistent with the ocean temperatures (40° to 85°C) estimated from oxygen isotopes on these same cherts (9, 11, 34). As an alternative, we suggest that the gypsum may have formed as diagenetic gypsum assemblages in muds, analogous to that observed along the continental slopes (>600 m) and deep-sea basins of the modern South and North Atlantic, where gypsum crystals grow in muds during influxes of cold Antarctic/Arctic bottom water. Such cold waters dissolve calcium carbonate on the sea floor, producing a source of Ca ions to react, respectively, with local sulfate (54, 55) or pyrite in cold-water muds and carbonate mounts (56). Similar cold oceanic regimes could have provided the physical and

Fig. 7. Deep-sea silicified muds with hydrothermal fractures and pseudomorphs of purported gypsum. (A) Cross section (polished slab) of a silicified volcanoclastic unit, now chert (HC1). A lower section (ca. 15 cm) of low-angle wavy laminations and rippled siltstones (pale gray) is overlain by a ca. 15-cm middle section of homogeneous silicified muds cut by an interconnected network of small translucent chert veins and tabular crystals of silicified sulfates, most of which are monoclinic [see also Fig. 10 in Lanier and Lowe (48)] and resemble gypsum crystals (dark gray), and by a ca. 15-cm top section of homogeneous pale-gray silicified muds with dark-gray silica pseudomorphs of purported diagenetic gypsum with a large variety of primary crystallization habits; rectangle enlarged as (B) shows four orthorhombic to pseudohexagonal sections of the silicified crystal forms, each connected by thin radiating chert veins (dark gray), suggesting a volume increase in the crystals, perhaps during transformation from anhydrite to gypsum. (C1 to C3) Further enlarged images of the three crystals. Note that the crystallization habit of the monoclinic gypsum crystals is often such that it shows a pseudohexagonal form, and that C1 represents several intergrown crystals. The host rock represents abyssal oceanic volcanoclastic muds deposited ca. 2 km below sea level (Fig. 1; 18). Similar diagenetic crystals and zones of fine-grained gypsum have been documented in deep-sea muds in which gypsum precipitation is initiated by episodic influx of cold ocean currents. Further detailed sedimentology and diagenesis of this and other chert units with associated sulfate pseudomorphs have been described previously (9, 18, 34, 48).
chemical conditions necessary for the precipitation of gypsum in the Barberton deep-water sediments. This is consistent with our interpretation that the inferred glacial deposits formed at sea level and suggests that cool conditions prevailed in the ocean and atmosphere between 3.45 and 3.47 Ga in the Barberton region.

CONCLUSION

We have presented new field data suggesting that the BGB may contain remnants of glaciogenic materials deposited in cold environments. Glacial conditions and cold deep-ocean water circulation at ~3.4 to 3.5 Ga during the deposition and hydrothermal alteration of volcanic and sedimentary rocks of the BGB are inconsistent with hot Archean oceans [for example, >40°C; (9, 11, 34)] but are compatible with the high temperatures recorded by oxygen isotopes and fluid inclusions in cherts formed during silification by hot fluids emanating from both the deep-water and the shallow-water hydrothermal systems that we have documented. The presence of dropstones in the NC is at odds even with seawater at 26° to 32°C [for example, (12, 13)], as neither tide-water glaciers nor winter sea ice is possible under such conditions. This implies at least a moderately cool climate with local glacial conditions and circulation of relatively cool deep-ocean water at around 3.4 to 3.5 Ga. Paleomagnetic measurements from these—the world’s oldest recorded glacial deposits—show that they formed at a latitude between ~20° and 40° (23), further implying that globally, at that time, the ocean and atmosphere temperatures were cool and that greenhouse gas concentrations were relatively moderate to 32°C [for example, (12)].

Supplementary Materials

Supplementary material for this article is available at http://advances.sciencemag.org/cgi/content/full/2/2/e1500368/DC1.

Volcanon stratigraphic framework of the HC and KC and associated cherts
Hydrothermal field and associated chert and volcanic eruption sequences
Temperatures of hydrothermal fluids
Fig. S1. Sample locations for the different sections of the HC.
Fig. S2. Sample locations for the different sections of the KC.
Fig. S3. Sample location for the Etimambeni section of the NC.
Fig. S4. Field relationships of silification processes in pillow lavas and cherts.
Fig. S5. Sketch map, location of analyzed samples, and pictures of the most typical lithologies of the pipes of the hydrothermal field of chert horizon CH2.
Fig. S6. Summary of the oxygen isotope values (δ18O; ‰) of mafic rocks and cherts of the OS.
Fig. S7. Summary of the oxygen isotope values (δ18O; ‰) of mafic rocks of the upper parts of the OS.
Fig. S8. Harker diagrams showing the relationships of SiO2 and other major and trace elements.
Fig. S9. Mid-ocean ridge basalt normalized multielement diagrams of various lithologies within and adjacent to the hydrothermal field.
Fig. S10. Relationship between SiO2 and δ18O and estimated hydrothermal alteration temperatures for the mafic lavas of the HC.
Fig. S11. Relationship between SiO2 and δ18O and estimated hydrothermal alteration temperatures for the mafic lavas of the KC and the MC.
Table S1. Geochemical data (δ18O and SiO2) of metabasalts and cherts from the HC, NC, and KC of the OS.
Table S2. Mineralogical and geochemical data from the hydrothermal field site of chert horizon H2.
References (61–69)

MATERIALS AND METHODS

Field work and geochemical analyses were carried out between 2004 and 2010, and drilling was completed in 2009. Details on sample collection and subsequent laboratory preparation of samples are described elsewhere (18–20, 57). For this work, 376 whole-rock samples were analyzed for δ18O (tables S1 and S2 and figs. S1 to S3), of which 309 were from HC and KC basalts and 30 were cherts from the HC and KC. Representative samples from all basalt sequences (varying from pillow rims to interiors and massive flows) were analyzed. Sampling was carried out at regular intervals through each of the sections logged; where good exposures allowed, the stratigraphic interval between samples was ca. 5 m or less. A further 11 samples were volcanoclastics, siliciclastics, and clasts, as well as cherts of the NC, from the drill core (DH 2a; fig. S3) through the NC [see Grosch et al. (57)]. An additional collection of 26 samples was analyzed from the hydrothermal pipes and related cherts, silica dykes, and breccias (table S2 and fig. S6B). The oxygen isotope compositions of the samples, along with their lithological descriptions and stratigraphic positions, are given in table S1, fig. 2, and fig. S1. To get the best possible view of the chert horizon H2 with the hydrothermal field and pipes, we used a helicopter (fig. 3).

All samples analyzed for oxygen isotopes were processed by K. Muehlenbachs at the University of Alberta following standard sample preparations. Oxygen from the silicate whole-rock powders was extracted by the BrF5 nickel tube method, converted into CO2, and analyzed on a Finnigan-MAT 252 mass spectrometer (15, 58–60). The yields were always above 95%. The data were reported with respect to the Vienna Standard Mean Ocean Water standards in the usual δ notation. The reproducibility of replicate oxygen analysis was ±0.1‰.

REFERENCES AND NOTES

1. C. Sagan, C. Chyba, The early faint sun paradox: Organic shielding of ultraviolet-labile greenhouse gases. Science 276, 1217–1221 (1997).
2. I. Crossen, J. Sanz-Forcada, F. Favata, O. Witasse, T. Zegers, N. F. Arnold, Habitat of early life: Solar X-ray and UV radiation at Earth’s surface 4–3.5 billion years ago. J. Geophys. Res. 112, E02008 (2007).
3. E. T. Wolf, O. B. Toon, Fractal organic hazes provided an ultraviolet shield for early Earth. Science 328, 1266–1268 (2010).
4. D. Waltham, On the absence of solar evolution-driven warming through the Phanerozoic. Terra Nova 26, 282–286 (2014).
5. H. Furnes, M. de Wit, H. Staudigel, M. Rosing, K. Muehlenbachs, A vestige of Earth’s oldest ophiolite. Science 315, 1704–1707 (2007).
6. S. Maruyama, T. Komiya, The oldest pillow lavas, 3,8–3.7 Ga from Isua supracrustal belt, SW Greenland: Plate tectonics had already begun at 3.8 Ga. J. Geophys. Res. 120, 869–876 (2011).
7. E. C. Pope, D. K. Bird, M. T. Rosing, Isotope composition and volume of Earth’s early oceans. Proc. Natl. Acad. Sci. U.S.A. 109, 4371–4376 (2012).
8. S. A. Wilde, J. W. Valley, W. H. Peck, C. M. Graham, Evidence from detrital zircons for the existence of continental crust and oceans on the Earth 4.4 Gyr ago. Nature 409, 175–178 (2001).
9. L. P. Krauth, D. R. Lowe, High Archean climatic temperature inferred from oxygen isotope geochemistry of cherts in the 3.5 Ga Swaziland Supergroup, South Africa. Geol. Soc. Am. Bull. 115, 566–580 (2003).
10. J. Karhu, S. Epstein, The implication of the oxygen isotope records in coexisting cherts and phosphates, Geochim. Cosmochim. Acta 50, 1745–1756 (1986).
11. F. Robert, M. Chausssidon, A palaeotemperature curve for the Precambrian oceans based on silicon isotopes in cherts. Nature 443, 969–972 (2006).
12. M. T. Heen, M. M. Tice, C. P. Chamberlain, Oxygen and hydrogen isotope evidence for a temperate climate 3.42 billion years ago. Nature 462, 205–208 (2009).
13. R. E. Blake, S. J. Chang, A. Lepland, Phosphate oxygen isotopic evidence for a temperate and biologically active Archean ocean. Nature 464, 1029–1032 (2010).
14. J. Veizer, J. Hoesf, D. R. Lowe, P. C. Thurston, Geochemistry of Precambrian carbonates: II. Archean greenstone belts and Archean sea water. Geochim. Cosmochim. Acta 53, 859–871 (1989).
15. K. Muehlenbachs, The oxygen isotopic composition of the oceans, sediments and the sea-floor. Chem. Geol. 145, 263–273 (1998).
16. J. F. Kasting, M. T. Howard, K. Wallmann, J. Veizer, G. Shields, J. Jaffrés, Paleoclimates, ocean depth, and the oxygen isotopic composition of seawater. Earth Planet. Sci. Lett. 252, 82–93 (2006).
17. S. T. de Vries, W. Nijman, P. L. de Boer, Sedimentary geology of the Paleaeoarchaean Buck Ridge (South Africa) and Kittys Gap (Western Australia) volcano-sedimentary complexes. Precamb. Res. 183, 749–769 (2010).
18. M. J. de Wit, H. Furnes, B. Robins, Geology and tectonostratigraphy of the Onverwacht Suite, Barberton Greenstone Belt, South Africa. Precamb. Res. 186, 1–27 (2011).
19. H. Furnes, M. J. de Wit, B. Robins, N. Sandstå, Petrology of lavas in the upper Onverwacht Suite, Barberton Mountain Land, South Africa. S. Afr. J. Geol. 115, 171–210 (2012).
20. H. Furnes, B. Robins, M. J. de Wit, Geochemistry and petrology of lavas in the upper Onverwacht Suite, Barberton Mountain Land, South Africa. Precamb. Res. 186, 28–50 (2011).
21. P. L. de Boer, C. J. Bray, E. T. C. Spooner, Fluid chemistry of Archean seafloor hydrothermal vents: Implications for the composition of circa 3.2 Ga seawater. Geochim. Cosmochim. Acta 61, 4025–4042 (1997).
22. M. J. Viljoen, R. P. Viljoen, Evidence for the existence of a mobile extrusive peridotitic magmatic system from the earliest part of the Barberton Orogen, South Africa, and its bearing on the characteristics and evolution of the early Earth. Geol. Soc. Spec. Publ. 201, 87–112 (1995).
23. A. J. Biggin, M. J. de Wit, C. G. Langereis, T. E. Zegers, S. Voûte, M. J. Dekkers, K. Droste, Palaeomagnetism of Archaean rocks of the Onverwacht Group, Barberton Greenstone Belt (southern Africa): Evidence for a stable and potentially reversing geomagnetic field at 3.5 Ga. Earth Planet. Sci. Lett. 302, 314–328 (2011).
24. M. J. de Wit, R. Armstrong, R. J. Hart, A. H. Wilson, Felsic igneous rocks within the 3.3- to 3.5-Ga Barberton Greenstone Belt: High crustal level equivalents of the surrounding Tonalite-Trondhjemite Terrain, emplaced during thrusting. Tectonics 6, 529–549 (1987).
25. S. T. de Vries, J. L. R. Touret, Early Archean hydrothermal fluids; a study of inclusions from the ~3.4 Ga Buck Ridge Chert, Barberton Greenstone Belt, South Africa. Chem. Geol. 237, 289–302 (2007).
26. A. Kröner, J. E. Hoffmann, H. Xie, F. Wu, C. Münker, E. Hegner, J. Wong, Y. Wan, D. Liu, Generation of early Archean felsic greenstone volcanic rocks through crustal melting, with implications for greenstone belt studies. Tectonics 30, 1–11 (2011).
27. R. P. Viljoen, M. J. Viljoen, The geological and geochemical significance of the upper formations of the Onverwacht Group. Geol. Soc. S. Afr. Spec. Publ. 2, 113–151 (1969).
28. M. J. de Wit, R. Armstrong, R. J. Hart, The Jamestown Ophiolite Complex, Barberton Mountain Land, South Africa. Geol. Soc. S. Afr. Spec. Publ. 2, 113–151 (1969).
29. A. Hofmann, C. Harris, Silica alteration zones in the Barberton Greenstone Belt: A window into subsurface floor processes 3.5–3.3 Ga ago. Chem. Geol. 257, 221–239 (2008).
30. M. J. de Wit, R. Hart, A. Martin, P. Abbott, Archean abiogenic and probable biogenic structures associated with mineralized hydrothermal vent systems and regional metasomatism, with implications for greenstone belt studies. Econ. Geol. 77, 1783–1802 (1982).
31. K. Abraham, A. Hofmann, S. F. Foley, D. Cardinal, C. Harris, M. G. Barth, L. André, Coupled silicon–oxygen isotope fractionation traces Archean silification. Earth Planet. Sci. Lett. 301, 223–230 (2011).
32. E. J. T. Stefurak, W. W. Fischer, D. R. Lowe, Texture-specific Si isotope variations in Barberton Greenstone Belt cherts record low temperature fractionations in early Archean seawater. Geochim. Cosmochim. Acta 150, 26–52 (2015).
33. V. Rouchon, B. Orbergen, Origin and mechanisms of K-Si-metasomatism of ca. 3.4–3.3 Ga volcaniclastic deposits and implications for Archean seawater evolution: Examples from craters of Kittys Gap (Pilbara craton, Australia) and Musaibi (Barberton Greenstone Belt, South Africa). Precamb. Res. 165, 169–189 (2008).
34. J. G. Moore, Petrology of deep sea basalts near Hawaii. Geol. Soc. Special Publ. 112, 1969–1977 (1977).
35. J. G. Moore, Petrology of deep sea basalts near Hawaii. Geol. Soc. Special Publ. 112, 1969–1977 (1977).
36. J. G. Moore, Petrology of deep sea basalts near Hawaii. Geol. Soc. Special Publ. 112, 1969–1977 (1977).
37. J. G. Moore, Petrology of deep sea basalts near Hawaii. Geol. Soc. Special Publ. 112, 1969–1977 (1977).
38. J. G. Moore, Petrology of deep sea basalts near Hawaii. Geol. Soc. Special Publ. 112, 1969–1977 (1977).
39. J. G. Moore, Petrology of deep sea basalts near Hawaii. Geol. Soc. Special Publ. 112, 1969–1977 (1977).
40. J. G. Moore, Petrology of deep sea basalts near Hawaii. Geol. Soc. Special Publ. 112, 1969–1977 (1977).
41. J. G. Moore, Petrology of deep sea basalts near Hawaii. Geol. Soc. Special Publ. 112, 1969–1977 (1977).
42. J. G. Moore, Petrology of deep sea basalts near Hawaii. Geol. Soc. Special Publ. 112, 1969–1977 (1977).
43. J. G. Moore, Petrology of deep sea basalts near Hawaii. Geol. Soc. Special Publ. 112, 1969–1977 (1977).
44. J. G. Moore, Petrology of deep sea basalts near Hawaii. Geol. Soc. Special Publ. 112, 1969–1977 (1977).
45. J. G. Moore, Petrology of deep sea basalts near Hawaii. Geol. Soc. Special Publ. 112, 1969–1977 (1977).
46. J. G. Moore, Petrology of deep sea basalts near Hawaii. Geol. Soc. Special Publ. 112, 1969–1977 (1977).
47. J. G. Moore, Petrology of deep sea basalts near Hawaii. Geol. Soc. Special Publ. 112, 1969–1977 (1977).
48. J. G. Moore, Petrology of deep sea basalts near Hawaii. Geol. Soc. Special Publ. 112, 1969–1977 (1977).
49. J. G. Moore, Petrology of deep sea basalts near Hawaii. Geol. Soc. Special Publ. 112, 1969–1977 (1977).
50. D. R. Lowe, G. R. Byerly, Stratigraphy of the west-central part of the Barberton Greenstone Belt, South Africa, in Geologic Evolution of the Barberton Greenstone Belt, South Africa, D. R. Lowe, G. R. Byerly, Eds. (The Geological Society of America Inc., Colorado, 1999), vol. 329, pp. 1–36.
69. J. Marin-Carbonne, F. Robert, M. Chaussidon, The silicon and oxygen isotope compositions of Precambrian cherts: A record of oceanic paleo-temperatures? Precamb. Res. 247, 223–234 (2014).

Acknowledgments: K. Muehlenbachs performed all stable isotope measurements, and J. Touret and M. van Zuilen performed fluid inclusion analyses of the silica pipes; discussions with them and with P. Hoffman and D. Bell improved the manuscript. This paper also benefited from the comments of six exceptional reviewers. 

Funding: This work was funded through the National Research Foundation of South Africa and the Norwegian Research Council.

Author contributions: This paper was based on data and numerous discussions involving both authors. The paper version was written by M.J.d.W., the supplementary text was written by both authors, and the illustrations and tables were prepared by H.F.

Competing interests: The authors declare that they have no competing interests.

Data and materials availability: All data needed to evaluate the conclusions in the paper are present in the paper and/or the Supplementary Materials. Additional data are available from the authors upon request. The AEON contribution number is 154.

Submitted 21 March 2015
Accepted 7 January 2016
Published 26 February 2016
10.1126/sciadv.1500368

Citation: M. J. de Wit, H. Furnes, 3.5-Ga hydrothermal fields and diamictites in the Barberton Greenstone Belt—Paleoarchean crust in cold environments. Sci. Adv. 2, e1500368 (2016).