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Invited review

Impact structures in Africa: A review

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Dedicated to our students, collaborators and friends in Africa, without whose support we could not have successfully pursued our investigations of confirmed and possible African impact structures.

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ABSTRACT

More than 50 years of space and planetary exploration and concomitant studies of terrestrial impact structures have demonstrated that impact cratering has been a fundamental process – an essential part of planetary evolution – ever since the beginning of accretion and has played a major role in planetary evolution throughout the solar system and beyond. This not only pertains to the development of the planets but to evolution of life as well. The terrestrial impact record represents only a small fraction of the bombardment history that Earth experienced throughout its evolution. While remote sensing investigations of planetary surfaces provide essential information about surface evolution and surface processes, they do not provide the information required for understanding the ultra-high strain rate, high-pressure, and high-temperature impact process. Thus, hands-on investigations of rocks from terrestrial impact craters, shock experimentation for pressure and temperature calibration of impact-related deformation of rocks and minerals, as well as parameter studies pertaining to the physics and chemistry of cratering and ejecta formation and emplacement, and laboratory studies of impact-generated lithologies are mandatory tools. These, together with numerical modeling analysis of impact physics, form the backbone of impact cratering studies.

Here, we review the current status of knowledge about impact cratering – and provide a detailed account of the African impact record, which has been expanded vastly since a first overview was published in 1994. No less than 19 confirmed impact structures, and one shatter cone occurrence without related impact crater are now known from Africa. In addition, a number of impact glass, tektite and spherule layer occurrences are known. The 49 sites with proposed, but not yet confirmed, possible impact structures contain at least a considerable number of structures that, from available information, hold the promise to be able to expand the African impact record drastically – provided the political conditions for safe ground-truthing will become available. The fact that 28 structures have also been shown to date NOT to be of impact origin further underpins the strong interest in impact in Africa. We hope that this review stimulates the education of students about impact cratering and the fundamental importance of this process for Earth – both for its biological and geological evolution. This work may provide a reference volume for those workers who would like to search for impact craters and their ejecta in Africa.

Contents

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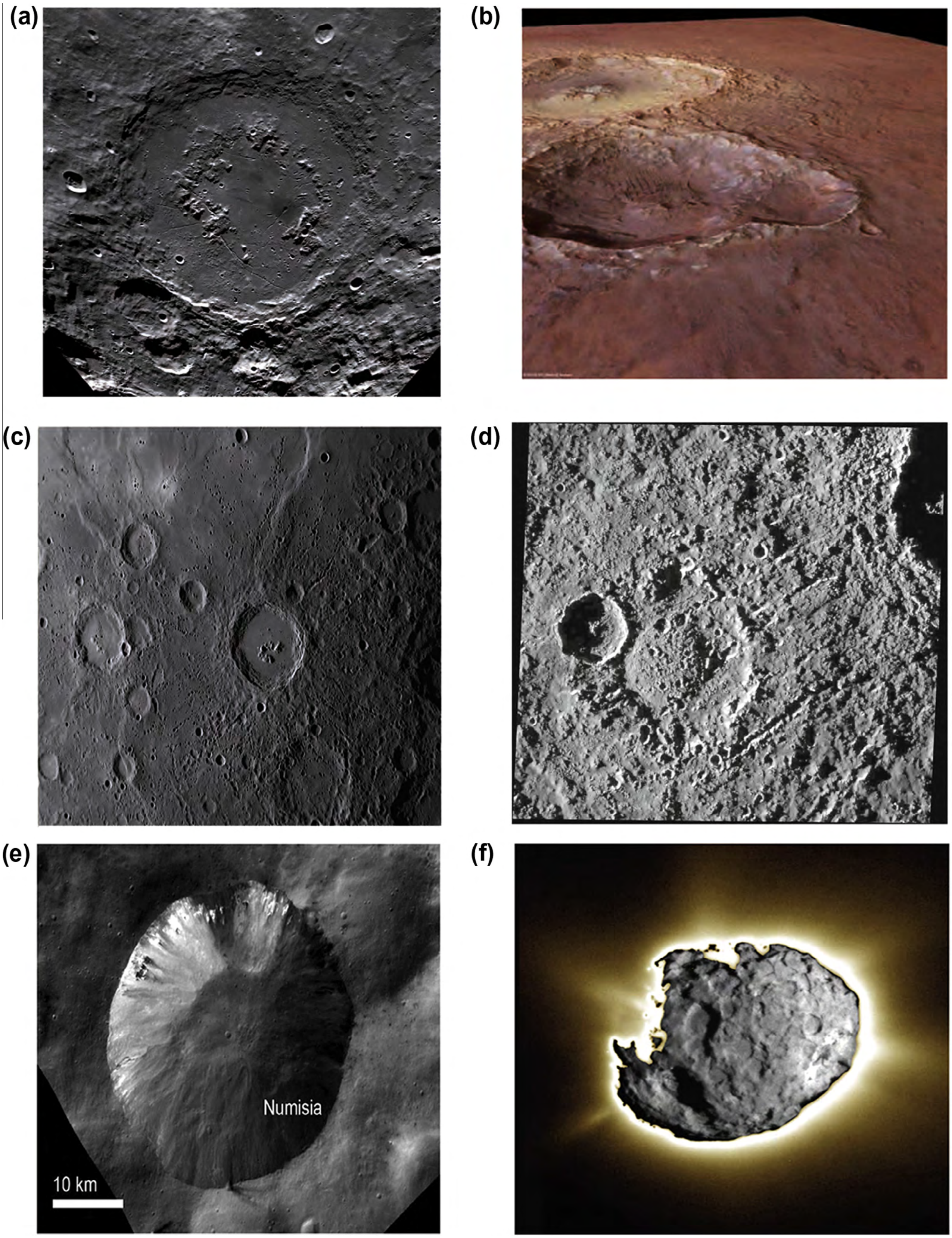
1. Introduction

More than 50 years of exploration of the surfaces of planets, moons, asteroids, and even of comet nuclei, have firmly established the fundamental role that impact cratering has played as a planetary process and surface-modifying agent on nearly all solid bodies in the solar system (Fig. 1a–f), throughout its entire history of 4.56 billion years. The only body in the Solar System lacking an impact record is Jupiter's moon Io – ostensibly because of recent resurfacing due to extensive volcanism. Accretion of planetary bodies today is understood to have been driven by continuous impact of ever larger particles and bodies – from tiniest dust grains to planetesimals (e.g., Taylor, 1992; Morishima et al., 2008; Hirashita, 2012).

However, impact is not purely a process of the distant past; in fact, several recent events have demonstrated that the danger of impact has very much persisted into the present. In July 1994 small fragments of comet Shoemaker-Levy 9 impacted successively into the atmosphere of the gas-giant planet Jupiter (Boslough et al., 1994) – impacts that were followed on TV by millions of humans. That these small, likely merely hundreds of meters sized, fragments packed gigantic punches was demonstrated impressively by the Earth-sized holes formed in the atmosphere of Jupiter. In the 1990s, the Pentagon released seismic records that proved that sizable impactors (up to several tens of meters in diameter) impacted onto Earth's surface repeatedly – albeit the recorded events took all place over the oceans and, thus, did not cause damage to humans. This and current efforts to safeguard Earth from catastrophic impact are, for example, discussed in Task Force on Potentially Hazardous Near Earth Objects (2000) or National Research Council (2009, 2010).

In September 2007 a small bolide, not larger than a meter or two in size, impacted at Carancas in Peru, and formed a 14-m-diameter impact crater (Borovička and Spurný, 2008; Kenkmann et al., 2009a). This relatively small event did cause some concern by the local, rural population, but did not cause any bodily harm. However, even such a small event in a densely populated area (such as a metropolitan area) could have caused the loss of hundreds of lives. Not quite so dramatic, but nevertheless causing injury to 1500 people and extensive damage to housing, are the consequences of the explosion of an originally (i.e., prior to impact) about 17–20 m wide meteoroid that exploded at a height of 23 km above the Russian city of Chelyabinsk (Urals) of 1 million inhabitants on 15 February 2013 (e.g., Borovička et al., 2013; Kohout et al., 2013). The explosion had an estimated magnitude of some 500 kilotons of TNT. The 1908 Tunguska bolide exploded with a force of 3–5 megatons, but was only roughly four or five times larger than the Chelyabinsk bolide. More than 100 kg of meteorites have been recovered around Chelyabinsk, and a large piece of


(a)
(b)
(c)
(d)
(e)
(f)
Numisia
10 km

about 650 kg has been recently retrieved from the bottom of Lake Cherbakul. A popular review of the event was recently published by Durda (2013).

Imagine if an explosion similar to Tunguska, which destroyed some 2000 km^2 of forest, would today occur above a mega-city – the outcome would be an unbelievable disaster. In fact, meteor blasts in the atmosphere occur from time to time – some of remarkable magnitude essentially remaining unnoticed by the public: a 50 kiloton explosion took place over the Indonesian island of Sulawesi on 8 October 2009 (Durda, 2013). Durda reports that explosions of Chelyabinsk magnitude are thought to occur about once per century, with Tunguska-like catastrophes being much scarcer at once every few centuries. A detailed assessment of the hazard from such small impactors was provided by Brown et al. (2013).

1.1. Fundamental importance of impact cratering

A recent issue of the journal "Elements" (Jourdan and Reimold, 2012) was dedicated to the topic of impact cratering (e.g., Grieve and Stöffler, 2012; Reimold and Jourdan, 2012) and does not only provide an in-depth review of the impact cratering process, but also of impact cratering studies and their challenges (Koeberl et al., 2012; Jourdan et al., 2012), and of the environmental consequences of impact (Pierazzo and Artemieva, 2012; for relatively recent reviews of fundamental aspects of impact cratering, see also Reimold, 2007; Pati and Reimold, 2007; Reimold and Koeberl, 2003; Koeberl, 2014).

As is evident on the old surfaces of the Moon, Mars, or Mercury, for example, just about every solid surface within the solar system has recorded numerous impact events, illustrating that impact cratering has been the dominant surface-modifying process since formation of earliest solid crust, more than 4.4 Ga ago. In fact, current knowledge favors a giant impact of a Mars-sized "projectile" into the early Earth – as the "war of the worlds" that lead to the formation of the Moon from a mixture of impactor and terrestrial mantle-derived components (e.g., various papers in Canup and Righter, 2000). Much work is being done to obtain surface ages for selected regions on the terrestrial planets by crater counting (see below; for benchmark papers on this topic: e.g., Hartmann et al., 2000; Neukum et al., 2001; Ivanov et al., 2003).

It is still debated whether or not the impact cratering rate in the inner solar system has decreased in intensity in an exponential way over more than 4 billion years, or whether there has been a broad spike (the so-called Late Heavy Bombardment) in the impact intensity curve around 4.1–3.8 Ga ago (e.g., Ryder et al., 2000; Stöffler et al., 2006; Koeberl, 2004, 2006a,b; Morbidelli et al., 2012; Marchi et al., 2013), and – as recently suggested – perhaps even tailing off till as late as 2.5 Ga ago (Bottke et al., 2012). Gigantic impact events formed the large multi-ring impact basins known from the Moon, but also from other planets and moons (e.g., the Hellas or Argyre basins on Mars, Walhalla Basin on Ganymede; South Pole-Aitken Basin on the Moon) prior to 3.8 Ga ago. While such basins must have formed on Earth in even higher abundance – because of its comparatively larger cratering cross-section and much higher gravitational pull – than on the smaller planetary bodies in the solar system (Grieve et al., 2006; Koeberl, 2006a,b), they did not survive the continuous dynamic evolution of our planet. However, in the early stage of solar system development, enormous impact events at hypervelocity transferred immeasurable amounts of kinetic energy, largely in the form of thermal energy, onto and into the proto-planets. Together with the then much enhanced heat-flow due to radioactive decay of ^{26}Al and isotopes of K, U, and Th, this primordial impact energy component is thought to have led to large-scale melting of early planets (see, e.g., Halliday, 2006, regarding the early evolution of the Earth). Clearly, impact cratering has played a decisive role in the earliest stages of planetary development. The early intense impact bombardment of Earth is also held responsible for the obliteration of earliest terrestrial crust (>3.8 Ga of age), which is only recorded in rare zircon age data from Western Australia (e.g., Wilde et al., 2001; Cavosie et al., 2004; Abbott et al., 2013; Bell and Harrison, 2013).

Much thought has been expanded in recent decades about the development of earliest life on this planet (e.g., Westall et al., 2006; Schopf, 2006; references therein), and the importance of water as a mandatory agent for the development of life has been central to this debate. One hypothesis suggests that water was brought to Earth through Hadean and early Archean impacts of giant comets (Hartogh et al., 2011). What is more, the idea of "Panspermia" (Arrhenius, 1903; Gladman et al., 2005) is based on fertilization of the universe through transfer of primitive life on impactors. Impact experiments with targets that were impregnated with spores or microbes showed recently that these primitive species could survive shock pressures as high as 50 GPa at a significant rate (Stöffler et al., 2007; Horneck et al., 2001, 2008; De la Torre et al., 2010; Grieve and Stöffler, 2012).

Since the recognition that a large-scale impact event at Chicxulub, Mexico, and a subsequent environmental catastrophe of global scale led to a mass extinction that affected some 70% of all life on Earth at the end-Cretaceous (for comprehensive discussions, see Montanari and Koeberl, 2000; Schulte et al., 2010), it has been debated by numerous groups whether other mass extinctions in the Earth's biological record could be related to catastrophic impact. This concerns the so-called "mother of all mass extinctions" at the end of the Permian, as well as extinction events in the late Devonian and at the Triassic–Jurassic and Jurassic–Cretaceous boundaries. To date, however, definite evidence in favor of impact having caused, or contributed, to these extinction events has remained elusive – or in some cases, at least questionable. Consider-

Fig. 1. A series of images of planetary surfaces that were subject to impact cratering – representing planets, moons, asteroids, and comets. (a) Mosaic of Clementine images of the 312 km diameter Schrödinger basin on the far side of the Moon. The structure exhibits marginal terraces on the inside of the rim and a peak ring structure in the interior. The structure itself has been repeatedly subject of later impacts, which testifies to its considerable age. Courtesy NASA/GSFC/Arizona State University. (b) Two impact structures imaged with the High Resolution Stereo Camera on board of Mars Express in the Tyrrhenia Terra region of Mars. This image exemplifies the outstanding imagery that in recent years has allowed to investigate detail (here 15 m per pixel) of planetary impact craters, and has provided much geological information about our planetary neighbor. The upper, ca. 35 km wide, complex impact structure is estimated to be a kilometer deep, and its rim rises by 400 m above the surroundings. A prominent ejecta blanket is visible around the structure. The lower crater is ca. 18 km wide and 750 m deep. These two structures are considered a "double impact crater" created by two fragments of the original single impactor. Credit: ESA/DLR/FU Berlin (G. Neukum). (c) MESSENGER image of a heavily cratered part of the surface of Mercury. The image is about 420 km across. Relatively young, fresh impact craters of both simple and complex morphologies are shown, and so are relatively older, degraded craters. Secondary crater chains are visible as well. Credit: NASA/Johns Hopkins University Applied Physics Laboratory/Carnegie Institution Washington. (d) Callisto, an icy moon of Jupiter: Heavily cratered region near the equator. The large, double ring crater on the middle left is known as Har and measures 50 km in diameter. The small, comparatively fresh crater partially superimposed onto Har is about 20 km wide. This younger crater displays a prominent central uplift, whereas Har is characterized by a broad inner mound. Galileo mission image PIA01054; credit NASA/JPL/University of Arizona. (e) Some amazing impact features have recently been visualized by NASA's Dawn mission to asteroid Vesta. Shown here is crater Numisia, of some 25 km width. This crater displays a bright lithology – which we here use to highlight the fact that crater wall imagery has been widely applied for subsurface stratigraphic analysis, also on Mars. This image is credited to NASA/JPL-CALTECH/UCLA/DLR/IDA/UMD. (f) Even comet nuclei do not escape impact bombardment. This image of the nucleus (ca. 5 km wide) of comet Wild 2 was taken during NASA's Stardust mission of 2004. The many circular features on the surface of this icy core are thought to represent – at least in part – impact structures. Credit: JPL-Caltech.

ing that the number of very large, and thus highly dangerous impactors is seriously limited in the asteroid belt (e.g., De Pater and Lissauer, 2001) and that comets only account for <15% of the Near-Earth Objects threatening our planet (http://www.neo.jpl.nasa.gov), danger to mankind to be eradicated by impact is not very probable – but still not entirely impossible (e.g., Shoemaker et al., 1990; Weissman, 1990). The environmental consequences of large-scale impact events could encompass tsunami, earthquakes, atmospheric disturbances (dust-related "impact winter" leading to global cooling), toxification of the hydrosphere depending on mineralogy of the target rock volume, disruption of food chains, etc. Scientists are debating what kind of magnitude impact would be required to overstep the threshold for extinction of our species. The Chicxulub event, likely involving a 5–15 km size projectile, dependent on its velocity and the density of the impactor, caused the formation of a 200 km diameter impact structure; however that event took place in a target that contained a large amount of evaporites capable of releasing environmentally detrimental gases (such as oxides of carbon and sulfur). Thus, it is not impossible that an impact forming an even larger crater structure in a more benign target might not be able to wipe out mankind entirely. But can we chance this? Can we even afford to suffer a relatively small impact of a 1 km projectile that might result in a 20 km size crater? Especially considering our much enhanced environmental vulnerability (in comparison to the dinosaurs and their contemporaries), for example our rather obvious and all-encompassing reliance on electronic communications, inter alia governing all transport of necessary goods and – above all – food stuffs, and the sensitivity of our race to even small environmental (temperature) changes, it is "safe" to predict that even such a small event of direct regional importance would lead to the demise of the population of a continent or more. Which government can afford to avoid taking measures against such a hyper-catastrophe?

1.2. Beneficial impact

On the positive side, some impact catastrophes have also had highly beneficial "side-effects". A large number of terrestrial impact structures have been recognized as hosts of valuable ore deposits (e.g., Grieve, 2005, 2013; Reimold et al., 2005a; Donofrio, 1998). Thereby, one commonly distinguishes three types of impact structure-hosted ore deposits:

(1) Ore deposits in impact structures that already existed *prior to* the impact event are known as *progenetic* deposits. The importance of impact may be that the event has made these deposits accessible to mining due to the stratigraphic uplift of the central crater region inherent to the modification stage of large-scale cratering (see below). Examples for this type of deposit are the Carswell uranium (Canada) or the iron-ore deposits of the Ternovka (also called Terny) structure in Krivoi Rog, Ukraine (cf. Sharpton et al., 2013). Importantly, the Witwatersrand gold and uranium deposits of the Vredefort impact structure largely fall into this category, although there is an authigenic (epigenetic, see below (3)) component as well (Reimold et al., 2005a; Frimmel et al., 2005, for relatively recent reviews of Witwatersrand geology and ore mineralization).

(2) Ore deposits that were formed as a *direct* result of an impact event are known as *syngenetic* deposits. Here, the tremendous base metal sulfide (Cu, Ni) and platinum group element (PGE) deposits of the Sudbury impact structure (Naldrett, 2003 and references therein) are a typical example. It is thought that this elemental wealth existed already in the target rocks but was enriched to the comparatively concentrated Sudbury ores due to formation of the thousands of km^3 of impact melt and various enrichment factors active within the melt body (differentiation). Many other base metal deposits in impact structures fall into this category of syngenetic deposits (Naumov, 2002; Reimold et al., 2005a). It has also been suggested that the carbonado occurrences of Central Africa (see below, Bangui geophysical anomaly; proposed impact structures at Kogo in Equatorial Guinea or Minkébé and Mékambo in Gabon) and in Brazil could be related to a catastrophic impact event in the west-central region of Africa, prior to the break-up of Gondwana. However, to date no *bona fide* evidence for the existence of a large impact structure in this strategic region has been reported, and there is so far no evidence that the formation of these carbonados was related to impact.

(3) Immediate *post-impact* epithermal/hydrothermal ore-forming processes are considered to lead to so-called *epigenetic* ores that are known from many impact structures (Naumov, 2002; Reimold et al., 2005a). Typical examples are the Pb–Zn deposits in the area of the Siljan impact structure in central Sweden, or the Cu–Pb–Zn and Au occurrences in and around the Sudbury impact structure. Of great economic importance are the authigenic gold mineralizations of the South African Witwatersrand basin. According to a model proposed by Reimold et al. (2005a; see below, section on Vredefort), fluid flow laterally away from the central parts of the Vredefort impact structure was caused by the stratigraphic uplift of hot, mid-crustal rocks of the central uplift and the thermal barrier of the hot impact melt sheet covering much of the impact structure (Ivanov, 2005). This model is consistent with the conclusions of Hayward et al. (2005) that the authigenic gold of the Witwatersrand gold fields was not deposited from a single, basin-wide operative fluid but from local fluid activation that caused dissolution of pre-existing detrital gold and rapid reprecipitation in the immediate environs, at centimeter to meter scales.

A second major point of importance of epigenetic deposits in and around impact structures are the highly economic hydrocarbon deposits known from many impact structures of quite variable size (ranging from a few to hundreds of kilometers – e.g., Ames, Oklahoma, 16 km wide – to the Chicxulub structure, Mexico [Grajales-Nishimura et al., 2000], 180 km diameter), mostly in North and Central America (Donofrio, 1998; Grieve, 2005, 2013). Billions of US$ worth of hydrocarbons have been produced from impact structures in North America annually.

Obviously, impact-related ore deposits represent an important benefit to mankind and undoubtedly there is further potential for other deposits to be discovered. Especially geologists in developing countries should take note of this potential.

Some impact structures have been used as water reservoirs with or without hydroelectric power facilities (e.g., Manicouagan, Quebec, or Boltysh, Ukraine), or as a fishing resource (Bosumtwi, Ghana). Dimension stone was produced for decades from the granitoid basement exposed in the core of the Vredefort impact structure, or from the Rochechouart and Ries impact structures in Europe. The still remaining, albeit dormant quarries in the central part of the Vredefort structure, the Vredefort Dome (see below), provide important 3-dimensional exposures of the upper and middle crust of the Kaapvaal craton and – in the context of this publication – of impact-generated rock deformation (Gibson and Reimold, 2008) that ought to be preserved for educational purposes alone. In the heart of the historic Voortrekker Monument outside the city of Pretoria (now called Tshwane) is a sarcophagus, which has been cut from Vredefort dimension stone. At Rochechouart, France, and Ries, Germany, the impact rock "suevite" (see below) has been quarried and used for the construction of important historic buildings (castle and church of the town of Rochechouart, the medieval St. George church and the town hall in the city of Nördlingen; and the exteriors of numerous buildings in the cities of Munich, Leipzig and even Berlin are clad with this lithology). Ries suevite has also been applied as a component for concrete production, especially the so-called *Trass* cement used widely for the remediation of monuments. Bentonite (Vredefort, South Africa, active mines), trona (Tswaing, South Africa), gypsum (Lake St. Martin, Canada), clay minerals, or coal occur in many impact structures and may still have (or had in the past) economic value.

A number of other positive aspects of impact structures must be listed. Quite a few impact structures are sites of museums, or at least educational displays, with the Ries Crater Museum in Nördlingen, southern Germany, being one of the best known (Buchner and Poesges, 2011) that also provides the central aspect of the Ries National Geopark (Poesges, 2011). Not only the impact process can be discussed in such facilities, but also the inherent planetological aspects (impact cratering in the solar system, nature of impactors: meteorites, asteroids, comets), regional geology, environmental consequences, and danger to mankind, but also reseeding of life after an impact catastrophe, ore formation, local biodiversity and environmental conditions, geography including hydrological circumstances, and anthropogeography). A discussion of the ecotourism potential of the Bosumtwi impact crater is given by Boamah and Koeberl (2007).

In Africa, temporary displays existed at Tswaing crater (South Africa) for several decades and a strong effort has been made there to erect a full crater museum. Such a facility would not only add enormous value to a visit of this well-preserved meteorite impact crater, but it would also allow to educate about the geological and economic importance of the wider region (the Bushveld Complex of South Africa with its gigantic mineral riches) and the socio-political aspects of this densely populated region of long political strife in the immediate environs of the geological landmark.

In July 2005 a part of the Vredefort impact structure (also South Africa) was declared a World Heritage Site by UNESCO (Gibson, 2011). Since then, this area has been developed with new infrastructure, and also a large Visitors' Centre was built in 2008. Unfortunately, the facility is still not completed because of construction-related issues and apparent political incompetence. It is hoped that this facility will soon provide South Africans and international visitors alike with a wealth of information about more than 3 billion years of development of a craton and the formation of the world's largest and oldest known impact structure, and consequences of this event that are still determining the geography and land use of the region. Comprehensive guides to the Vredefort impact structure were published by Gibson and Reimold (2008) and Reimold and Gibson (2010).

Another important aspect of impact structures is that the crater synforms may represent closed basins, with the impact age representing an exact time marker of basin formation and onset of sedimentation therein. This may provide accumulations of very detailed sedimentological records that, by themselves, can represent important paleoclimatic records (e.g., Harms et al., 2007; Oberhänsli and Emmermann, 2011). A number of large impact structures have been drilled in the last decade with studies of paleo-environmental records of this nature forming an explicit objective, besides better understanding of the impact process, impactite formation, and impactite emplacement. An impactite is a lithology that has been created as a direct consequence of an impact event. In Africa, the ICDP drilling of the Bosumtwi impact structure in Ghana must be noted (Koeberl et al., 2007a). Detailed lake sediment records for a location near the equator were recovered. Here also environmental concerns were highlighted in the cause of this project, such as overfishing of the lake, the educational and recreational potential of the lake and its environment, and the opportunity for astronomical, astrophysical, and geological tuition at this regionally unique location (e.g., Boamah and Koeberl, 2007).

1.3. Why a special review in JAES dedicated to impact structures in Africa?

This question has been answered already on the preceding pages where the fundamental, universal importance of impact cratering as a planetological, geological, and environmental process has been emphasized. But it could be argued that this is not the first review of African impact structures, as this work was preceded by a review by Koeberl (1994). Despite extensive work and new discoveries on African impact structures and many other structures that have been proposed as such, only abstract form compilations of confirmed and suspected impact structures in Africa have been contributed since 1994 (e.g., Master and Reimold, 2000; Youbi et al., 2011). Throughout the two decades, not only many new impact structures have been discovered or proposed on the African continent, but a vast body of new work including many important contributions to this record and also of global significance has been published. Impact cratering studies have evolved into a mainstream scientific discipline, with many investigative methods having been added or improved, and a global body of impact knowledge has been assembled that must be referred in any such review effort.

This has drawn attention to the need for continuous promotion of impact cratering as a serious geological process that needs to be part of the university education of every student of geoscience – also in Africa. Which economic geology course can afford to avoid discussing the economic potential of impact structures such as Vredefort or Sudbury? It is a fact that African countries are heavily reliant on revenue from ore resources, and basic knowledge about impact cratering and impact geology should be offered to every student of geoscience. And what better example would be there for the "new geoscience" – integrated *system earth science* – than the multidisciplinary, geological, and planetary impact cratering discipline!

In the absence of a dedicated textbook, this review may serve as both an introduction to general aspects of impact cratering, as well as a comprehensive guide to Africa's impact structures. Thus, it is high-time to update and synthesize the current knowledge about Africa's impact structures, in order to facilitate access to the pertinent literature, to emphasize the important role that investigation of African impact structures has played, and still has to play, regarding our general understanding of the impact cratering process. This undertaking is also based on a strong incentive for us, namely to caution about the serious problem of uncritical promotion of alleged impact structures based on insufficient evidence, which unfortunately has become a widespread problem. Finally, it is hoped that the African geocommunity will take note of the heritage value that many impact structures on this continent represent.

2. The study of impact structures

2.1. The tools of impact cratering science

The main purposes of searching for and studying impact structures are: (1) identification and then confirmation of an impact; (2) improving the terrestrial impact cratering record to evaluate how cratering intensity may have changed over geological time; this includes the dating of impact structures (and, thereby, of the events); (3) investigations related to the understanding of the physical and chemical processes inherent to impact cratering; (4) their dependence on the individual conditions related to the nature (geological composition, stratigraphy, tectonic arrangements such as effects of faults or non-horizontal stratigraphy) of the target rock(s); (5) improving the impact-related scaling relationships such as the function of energy released in relation to crater size, or the variance of stratigraphic uplift to crater diameter for different target rock types and configurations; (6) correlation of impact magnitude and environmental effects; and (7) basic investigations related to *shock metamorphism*, i.e., the response of rocks and minerals to impact-induced pressure and temperature.

A range of tools (e.g., Fig. 2a–c) are available for impact cratering studies, with most investigations having been conducted ever since impact crater analysis commenced in earnest in the late 1950s in an integrated, multidisciplinary fashion. The recognition of the early 1980s that the mass extinction at the Cretaceous-Paleogene boundary could be related to an impact catastrophe, and the debate about possible traces of primitive life in the Martian meteorite ALHA 84001 in the mid-1980s, must be credited with the subsequent elevation of impact science into geoscientific, planetological, and astrobiological mainstream research. A detailed review of the methodology of impact cratering studies has recently been published by French and Koeberl (2010).

Impact studies are conducted on a range of scales, from remote sensing investigations of planetary surfaces, through kilometer- to meter-scale field investigations of impact structures and ejecta horizons, to hand specimen, to sub-millimeter, and even micrometer-to-nanometer-scale laboratory investigations. Mineralogy is at the forefront of analysis of impact-induced deformation phenomena, from the deformation and transformation of target minerals to the formation and differentiation of impact melt rock. Geochemical techniques are vital for the tracing of elemental or isotopic relics of the projectile in impact-generated rocks – and understanding possible fractionation mechanisms that might hinder fingerprinting the meteorite type of an impactor. State-of-the-art geochronological methods allow, in some cases, to determine a precise age for an impact event and can provide information regarding correlation with an ejecta horizon or with environmental change.

Impact (shock) experiments with acceleration of chemically well-defined projectiles onto targets designed from minerals, rocks, or metals, by means of explosive-driven devices or compressed light-gas guns, allow to determine the behavior of materials under the extreme pressure and temperature conditions of impact (hypervelocity shock physics), as well as the shock and temperature calibration of the shock metamorphic effects that are observed in naturally shocked materials from impact structures and ejecta components. Finally, a vital and highly prospective avenue of impact research has become the technique generally known as numerical modelling, which allows to focus on individual aspects (parameter studies) of cratering, from astronomical considerations such as the orbits of projectiles and obliquity of impact, via kilometer to sub-millimeter modelling of target (stratigraphic succession, localized deformation features such as faults or folds, mineral assemblages, effect of pore space, inter alia) and ejecta (dissemination of ballistic ejecta and ejecta curtain materials, origin of ejecta within different levels of the target). The understand-

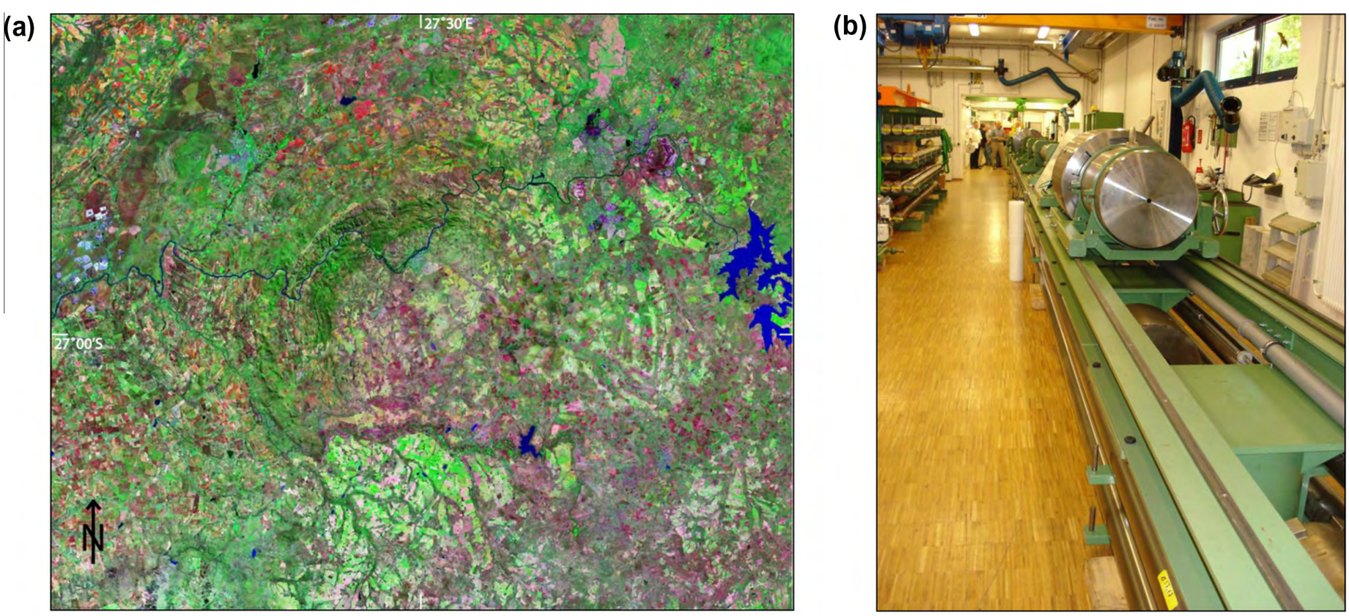


Fig. 2. Some of the tools of impact cratering studies. (a) *Remote sensing*: False-color Landsat thermal mapper satellite image of the region around the Vredefort Dome (roughly 145 km wide; width of the Vredefort Dome ca. 90 km). Courtesy of Mike Phillips, formerly of Cardiff University, United Kingdom. The large water body at the far right of this area is part of the Vaal Dam reservoir. Note the obvious course of the Vaal River through the Vredefort Dome, where its bed is largely determined by topographic and structural (fault lines) aspects. The prominent mountain land of the collar of the Vredefort Dome is also well visible. (b) *Experimental shock*: The extra-large (XXL) two-stage gas gun of the Ernst-Mach-Institute for High-Speed dynamics, Freiburg. The length of the full apparatus up to the sample recovery chamber (red, in background) is about 45 m. Most recently, one application of this gun was to accelerate centimeter-sized projectiles of metal or iron meteorite onto large blocks of porous sandstone to investigate the response of such material, which is of course abundant in the upper crust of Earth, to hypervelocity impact (see Kenkmann et al., 2011a). (c) *Numerical modelling*: Results of a thorough numerical modelling experiment by Ivanov (2005, 2008). Here, results of simulations based on geological and geophysical constraints from the Vredefort impact structure are shown. Top: isotherms after the impact event in the rock volume of the central uplift (left part) and surrounding ring syncline. The post-shock temperatures estimated by Gibson and Reimold (2005) from petrographic observations of the exposed strata are indicated for three locations – clearly the modelling achieved excellent agreement with the observed values. White lines and numbers indicate from which depth (in kilometers) the rocks along these lines have been uplifted. Dashed lines suggest the estimated limits of erosion depth since formation of the Vredefort structure ca. 2 Ga ago. Middle: The profile schematically shows the geology between the central part of the central uplift (left) and the outer Potchefstroom Synclinorium (right), extending from mid-crustal granitoids of the core to the supracrustals of the collar of the Vredefort Dome, and then into the synclinorium comprising mainly Transvaal Supergroup metasediment. Bottom: Here the modelled variation of shock pressures is shown in black lettering, with the limits of known PDF and shatter cone (SH) occurrences schematically illustrated. According to this modelling, the rocks of the inner central uplift should have experienced shock pressures >40 GPa, in good agreement with the estimate of Gibson and Reimold (2005) of >30 GPa.

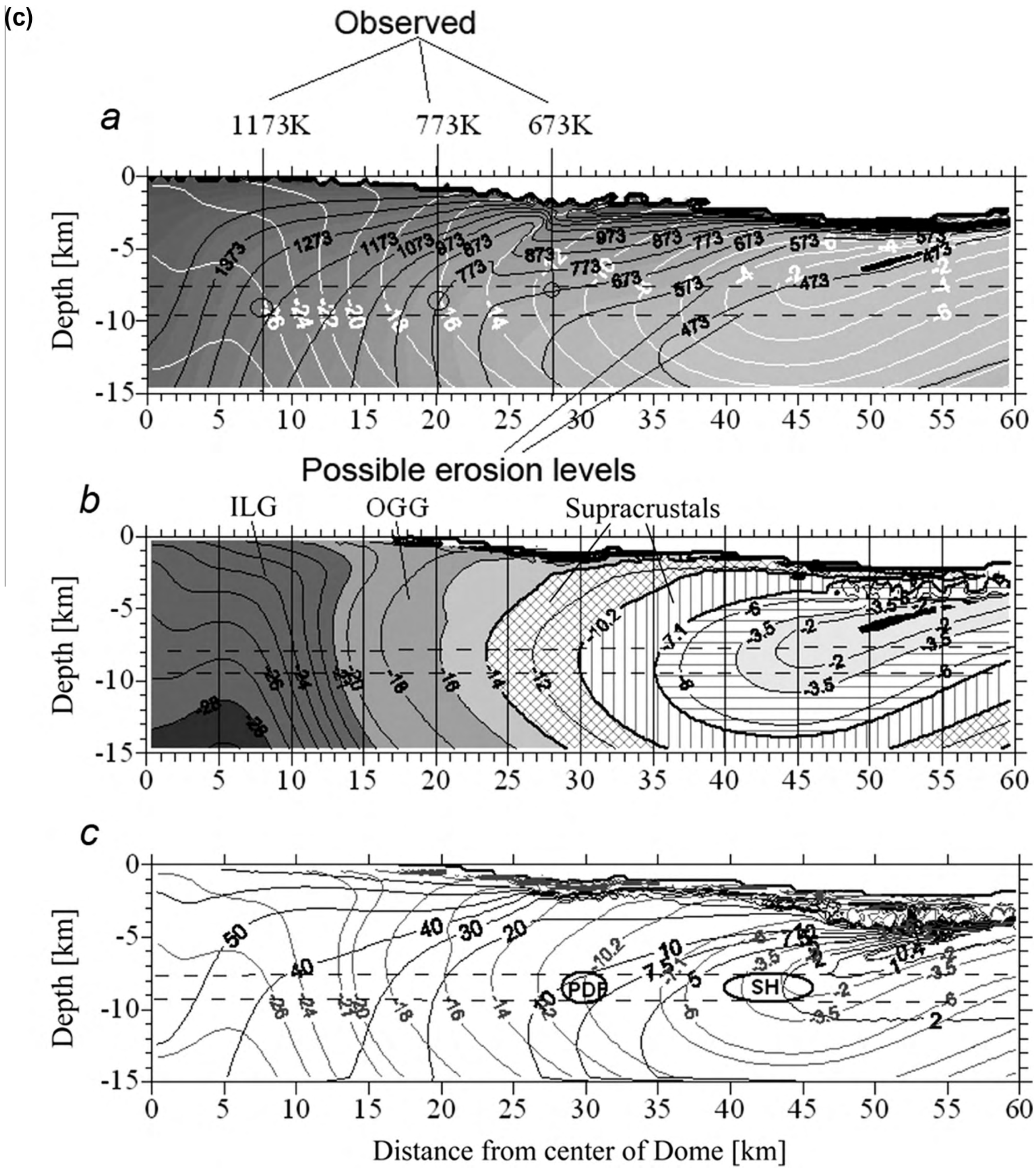


Fig. 2 (*continued*)

ing of physicochemical material behavior and the kinematics (material flow, ejection and displacement) of the phases involved in impact events has been dramatically enhanced by this methodology. Its particular strength will be revealed by comparative modeling and hands-on material analysis, which has been the focus of two important "Bridging the Gap" conferences (Pierazzo and Herrick, 2004; Herrick et al., 2008).

2.1.1. Remote sensing and geophysics

The techniques of both these methodologies have been highly important for the initial recognition of impact structures and as mapping aids on the ground, as well as providing an impact crater density-based means for relative chronology of planetary surfaces, and geological analysis of craters. Satellite and aerial photography, and in recent years more and more the widely-accessible "Google Earth" data, have drawn attention from scientists and laypersons to numerous crater-like structures. In some cases it was possible to indeed verify some of the structures as impact-generated, but this is not possible on the basis of remote-sensed data alone – in every case *in situ* ground-truthing is required through geological assessment of a structure and laboratory-based verification of *bona fide* impact evidence (e.g., French and Koeberl, 2010). Remote sensing data may also provide extremely useful – even essential – support for geological field work, especially in remote areas where other orientation means are not available.

The more or less densely cratered, and thus variably old planetary surfaces of the Moon and terrestrial planets have been evaluated by crater counting techniques (Hartmann et al., 2000; Neukum et al., 2001; Ivanov et al., 2003). Especially the high-resolution imagery provided by spacecraft in the last decade (e.g., SELENE and LROC data for the Moon, the MESSENGER imagery of the surface of Mercury, and the HIRISE data sets for Mars) has provided unsurpassed data sets for the statistical analysis of ever smaller crater diameter classes. The principle of crater-counting chronometry is simple: the older a surface, the more craters should be accumulated. Obviously the technique is not that simple, with parameters such as gravity, atmospheric density, and position of a body in the solar system with respect to the source regions of bolides and, thus, impact flux, or the nature of target lithology that obviously reflects on the size of impact craters produced having to be considered. It is obviously advantageous to have some absolute ages for geological formations to provide benchmark values for the calibration of crater frequency statistical curves, and fortunately some such values are available for the lunar record. The lunar crater production curves have then been adapted for the conditions (gravity, atmospheric density) on other planetary bodies such as Mars.

Impact crater density in certain regions on Earth has also been used to calculate impact flux values for various periods in Earth evolution. For example, based on the relationships established by Grieve (1984) and Shoemaker (1984), already Garvin (1986) estimated that in Central Africa some 5–10 impact structures larger than 10 km in diameter should occur, whereas so far only two (Bosumtwi, Ghana, and Luizi, Democratic Republic of Congo) are actually known.

Obviously, it is very difficult to generalize how many impact structures should be preserved in a given region of the Earth. The geologic history of that region determines how many structures could have been preserved from erosion, are perhaps covered by younger sediment, have been completely obliterated by sedimentation, or could be exhumed at any given time. Ivanov (2008, his Fig. 5) estimated from the lunar cratering record that, on average, one new >10-km-diameter crater would appear on Earth (oceanic plus continental crust) every 1 million years or so, and a >50-km-diameter crater every 5–10 million years. And a 100–200 km diameter crater, likely the minimum crater size thought to be related to a major, mass extinction-type global catastrophe, would form ca. every 100 million years. This author also estimated an average of 220 Ma for the obliteration time required to completely erase a crater >9 km. Naturally, this average value ought to vary significantly from region to region, between oceanic and continental crust. The effects of plate tectonics demand that oceanic crust is subducted at varied rates between 50 and 150 million years. Stable continental platforms may accumulate impact structures and preserve their erosional remnants for periods up to several billion years – the reason why the Scandinavian impact record is quite exceptional (see Earth Impact Database website).

2.1.2. Geophysical data sets

A large number of impact structures has been identified initially because of geophysical, mostly gravity and magnetic, anomalies; in particular, those crater structures refer here that are deeply eroded and/or buried by post-cratering sediments. Just a few examples are the large Chicxulub structure (66 Ma, 180–200 km diameter) in Mexico (Sharpton et al., 1993; Schulte et al., 2010 and references therein), or the economically important Ames Structure (16 km, 470 Ma, Oklahoma, USA) (Carpenter and Carlson, 1992). Several structures were, thus, recognized in the course of economic exploration programs and then confirmed by drilling and subsequent hands-on laboratory analysis of drill core. In Africa, the Kgagodi structure in Botswana was initially identified through gravity analysis as part of a hydrological exploration project (Paya et al., 1999), and geophysical analysis was instrumental during the early investigation of the mostly sediment-covered Morokweng structure of northwest South Africa (Corner et al., 1997).

Typically, impact structures that have not been eroded to, or beyond, the crater floor are characterized by negative gravity anomalies caused by impact-induced fragmentation and brecciation of the target rock and by high-porosity impact-breccia fills of the crater forms – in essence representing circular (simple bowl-shape structures) or annular (complex crater forms with central uplift structure) zones of reduced density. In the case of complex structures, and where craters are deeply eroded, anomaly patterns may be more complex, as they would be largely determined by the subcrater basement geology that can be quite complex both lithologically and with regard to long-term geological evolution (metamorphism, alteration, etc.). Magnetic anomalies may be circular over a simple crater, or ring-shaped in cases of complex crater structures with central uplifts (see below), but not necessarily so. Magnetic signatures of impact structures are determined by the magnetic properties of crater fill and subcrater (target and impact-generated/injected) lithologies, and by impact-induced thermal and chemical remanence, and they, thus, are – more often than not – quite complex (Henkel and Reimold, 2002). Reviews of geophysical signatures of impact structures are found in Pilkington and Grieve (1992), Grieve and Pilkington (1996), and Pilkington and Hildebrand (2003).

Besides potential field studies, seismic investigations have occasionally led to proposals of the presence of an impact structure, when a stratigraphic uplift is indicated in the inner parts of a crater-like feature. For example, the 45-km-diameter Montagnais crater on the continental shelf off the east coast of Nova Scotia (Canada) was recognized from seismic patterns acquired in the course of oil exploration (Jansa et al., 1989). However, also geological investigations of large impact structures may be greatly enhanced by the application of geophysical methods, prominent examples being Chicxulub, Vredefort, and Sudbury (see Grieve et al., 2008, and references therein). Where impact structures are largely or entirely buried, geophysical characterization is a prerequisite prior to selection of drilling sites.

It should be emphasized once again that geophysical patterns are not conclusive in determining the impact origin of a structure. Nowhere has this been better illustrated than in the case of the alleged Bedout impact structure. Becker et al. (2004) suggested that a hundreds of kilometers sized impact structure occurred offshore of northwestern Australia. Their evidence was an alleged resemblance of the gravity pattern in this region to the gravity signature over the Chicxulub structure. They also claimed that they had found impact breccias (with shock deformation features) in a drill core extracted from Bedout, and that this material had the exact age of the Permian–Triassic boundary. This suggestion quickly drew enormous interest in the scientific community and the general public, but it was also severely scrutinized. Glikson (2004) rejected the alleged impact deformation evidence, Renne et al. (2004) demonstrated the improbability of this age coincidence on the basis of the lack of significance of the alleged impact age, and finally, Müller et al. (2005) showed conclusively that the geophysical evidence does not indicate the presence of an impact structure but rather suggests genesis of this feature from endogenic processes. Until now, the alleged shock metamorphic evidence in the form of diaplectic plagioclase glass (maskelynite) has not been confirmed.

Ground penetrating radar has been employed on occasion in impact structures, but the limited penetration depth of this technique restricts its applicability to immediate subsurface studies, for example of the distribution of impact breccia in the environs of a crater (e.g., Grant and Schultz, 1994 – at Meteor Crater; Grant et al., 1997 – at Roter Kamm, Namibia). The high-porosity impact breccias of the crater fill also lend themselves to investigation by electrical methods, such as resistivity mapping (Henkel, 1992), and magnetotellurics has been used to investigate crater depths and maximum depth of subcrater deformation (e.g., at Serra da Cangalha, Brazil, Adepelumi et al., 2005; see also discussion in Vasconcelos et al., 2012a). At this latter structure an intriguing annular pattern of radiometric element signatures (gamma-ray spectrometry) has also been mapped (Vasconcelos et al., 2012b), the interpretation of which is still debated. Radiometric data have also been obtained over the Bosumtwi structure in Ghana, where an annular anomaly outside of the crater itself seems to be associated with widespread K alteration in the environs of the crater, specifically of the ejecta blanket (Boamah and Koeberl, 2002; Koeberl and Reimold, 2005).

Overall, geophysical analysis is highly useful for the recognition of possible impact structures, in support of geological analysis of crater structures, and particularly where surface geological access is not provided. *However, by itself, neither remote sensing nor geophysical data sets suffice to confirm the presence of an impact structure. Geological ground-truthing, in conjunction with laboratory analysis of crater rocks, is required in every instance – with samples to be acquired either from field work or from drilling.*

2.1.3. Field work and drilling

Ground-truthing of an impact structure is essential. This entails – in both simple and complex impact structures – the search for definite macroscopic evidence of impact, such as *shatter cones* (see below). Also, lithologies that are unique with regard to regional geology – in particular breccia occurrences and melt rocks, could be essential for the further investigation of an impact structure, as they have the highest potential to exhibit diagnostic impact evidence in the form of *shock deformation*. As discussed below, impact melt rock and impact-produced pseudotachylitic breccias provide the best material for dating of an impact event. Several field expressions of impact breccias and of a shatter cone are shown in Fig. 3.

Notably, geological analysis does not stop within the confines of a crater-like structure but entails comparison with the geology and deformation as found outside of the structure. This includes searching for evidence that might indicate a zone of stratigraphic uplift in the inner part of the structure (central uplift, see below), in comparison with regional stratigraphy. Scaling relationships linking the amount of stratigraphic uplift with the diameter of an impact structure (e.g., Melosh, 1989, 2002) can be used to estimate the original size of an impact structure even in cases of structures that are eroded or tectonically dismembered. The regional tectonic situation (rock deformation, presence of fault or shear zones) must be investigated, as local deformation enhancement may be the result of impact but could be due to tectonically induced changes to the crust as well.

A lot of work has been done investigating the possibility that “tectonic” indicators could provide diagnostic evidence for impact. It is well-known that strata at crater rims should be up- or even overturned – showing the characteristic inverse stratigraphy of impact crater rims. Asymmetric tectonic configurations as found in crater rims or in central uplift structures can be indicative of oblique impact (e.g., Kenkmann and Poelchau, 2009; Kenkmann et al., 2010, 2014), but are also dependent on degree of erosion (i.e., different observations have been recorded in poorly and severely eroded crater rims and central uplift structures). In any case, it is necessary to keep an open mind during fieldwork: preconceived ideas that a given structure would have to be of impact origin may lead one onto the wrong track.

Crater-like structures can be produced by many other processes, such as sinkhole formation, volcanic processes (maars, collapsed calderas, volcanic vents including kimberlite pipes), tectonic movements, landslides, karstification, or glacial processes. It is obviously necessary to consider the respective geological situation in its entirety – e.g., whether a crater-like structure occurs in carbonate terrains, in tectonic belts, or regions of volcanic activity in the past or present, or could be the result of glacial overprint (as thought to be the case for the many crater-like “holes” in the Chiemgau region of southern Germany that have been punted by K. Ernstson (Würzburg) and colleagues as a meteorite impact crater strewn field – without ever providing unambiguous evidence for impact! The critical reader may find these allegations in Ernstson et al. (2010); a critical assessment of this alleged Chiemgau impact had been published already before by Heinlein (2009), but notably remained ignored by the proponents of this alleged impact event). Crater structures in volcanic regions hold a particular challenge – considering that it is not impossible that impact cratering might affect volcanic terrains as well. This problematic is highlighted by recent reports of an entire impact crater strewn field in the volcanic Bajada del Diablo area of Argentina, where many crater-like features have been related to impact but, to date, no conclusive pro-impact evidence – what-so-ever – has been recognized (Acevedo et al., 2009). And the thick ice-cap of Antarctica has not protected the impact cratering community from the report that a large, 400 km diameter, Permian–Triassic Boundary impact was located on the Antarctica continent at the site of an alleged gravity anomaly (von Frese et al., 2009). Obviously besides the less than obvious geophysical observation no tangible evidence to validate this allegation has ever been offered. Amazing how it has even been possible to assign an age to this spurious impact event... This case has highlighted a common flaw: misleading disregard for the fundamental scientific principle that requires obtaining proof for a new hypothesis *before* it is reported as fact!

An amazing amount of material has been published in the last years about an alleged impact event at about 12,900 years ago, in the Younger Dryas. This putative event has been alleged to have affected the climate at that time, and thus biodiversity in North America, including living conditions for the then foraging early Americans. Evidence quoted in favor of impact has included abundant nanodiamonds in correlated sediment, widespread soot, and spherules of “cosmic” appearance (e.g., Firestone et al., 2007; Bunch et al., 2012). In contrast, other workers have not been able to confirm the presence of any alleged impact evidence (e.g., Paquay et al., 2009; Surovell et al., 2009). The whole controversy, and a detailed discussion why the “evidence” cited by Firestone et al. (2007), Bunch et al. (2012) and coworkers is either not convincing or contradictory, has been presented by Pinter et al. (2011) and Boslough et al. (2012, 2013).

Various claims about Chiemgau and Antarctic impacts, and an alleged Younger Dryas impact catastrophe, have made it into the secondary literature already, despite the ongoing controversy about these speculative claims. The reader is cautioned to conscientiously evaluate the data for these and other inconclusive but high-profile cases.

Detailed sampling of country rocks and other lithologies is required to allow petrographic analysis, especially in search of shock metamorphic indicators. Special effort should be made to sample all lithologies, also in order to investigate the precursor components and the proportions, at which they occur in the “target area” and at which they could have been incorporated into impact breccias. Should it be possible to utilize impact breccias of the suevite or impact melt rock types (see below) for chemical tracking of an impactor component, care should be taken to analyze representative samples of all possible target rocks. And as it is very desirable to constrain the age of impact events, the geological field work should also be conducted under consideration of searching for relative chronostratigraphic evidence and for lithologies that might allow absolute – or at least relative (i.,e., dating of lithologies, such as tuffs through U–Pb zircon dating, the ages of which might bracket the impact event) dating of the impact.

General geological analysis of an impact structure may involve much more than the above mentioned, especially in the case of large, old, poly-metamorphosed terrains. The case in point is presented by the Vredefort impact structure, as reviewed in all facets by Gibson and Reimold (2008). The origin of Vredefort was controversial for nearly 100 years, but the extended study of this

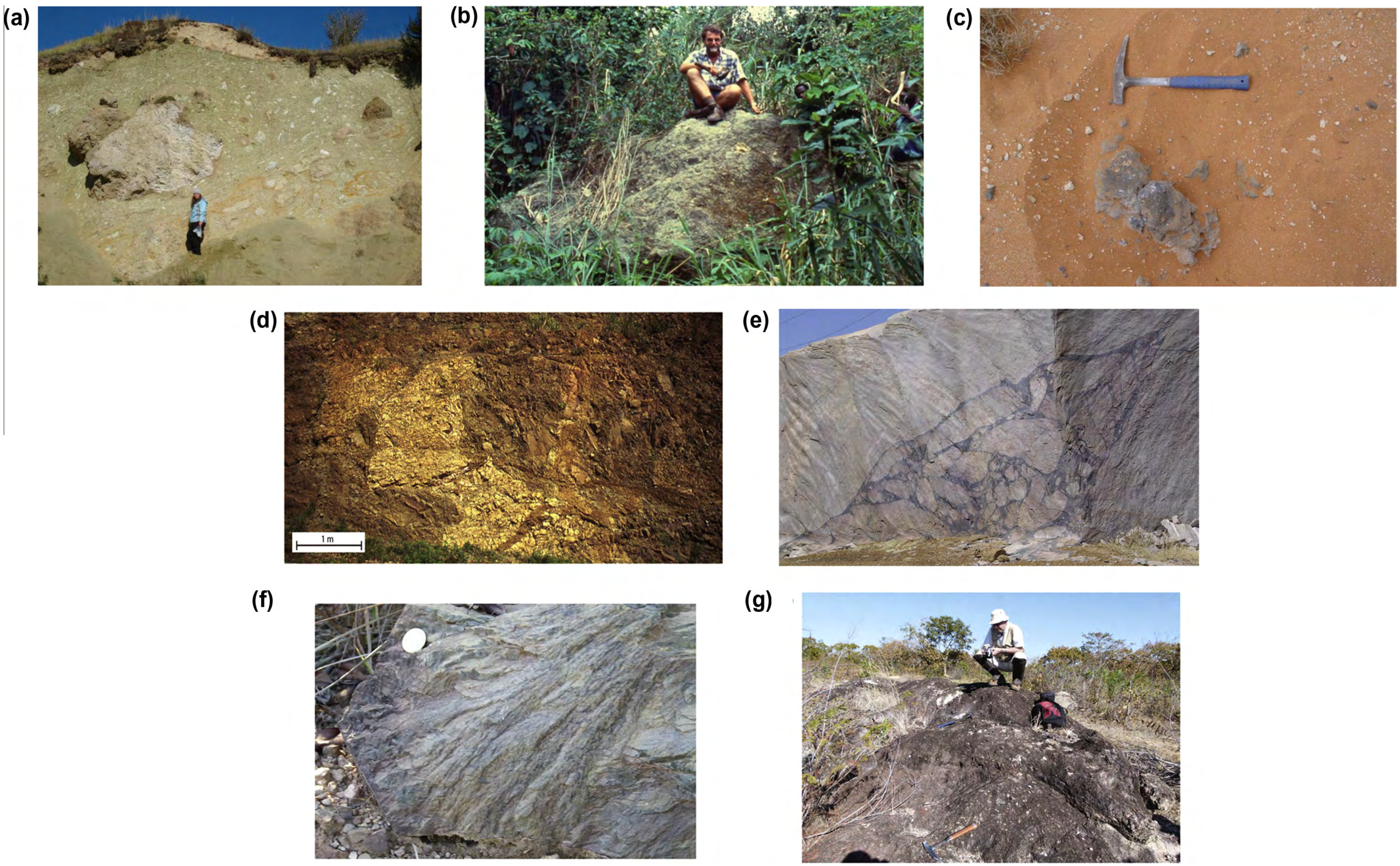


Fig. 3. Exposures of different impact breccias and a typical shatter cone. (a) Part of the quarry face at Limberg Quarry near Unterwilflingen, Ries crater, Germany. This view shows a polymict lithic breccia overlain by a thin layer of Bunte Breccia (which is ballistically emplaced lithic breccia). (b) Relatively large exposure of suevite in the rainforest/cocoa plantations of the outer, northern crater rim of the Bosumtwi impact structure (Ghana). (c) Patch of impact melt rock from the top of the western crater rim of Roter Kamm. Hammer for scale 35 cm long. The melt rock from this exposure was described in Hecht et al. (2004). (d) Road-cut on the inner crater wall of Bosumtwi impact structure (Ghana). Large, dismembered, blocks of different metasedimentary rock types form a polymict megabreccia, which is the result of in- and downward directed slumping at the edge of the impact structure. (e) Exposure of pseudotachylitic breccia, ca. 2.5 m wide, in Archean granite gneiss from Salvamento quarry in the northern core of the Vredefort Dome. (f) A large (ca. 30 cm long) shatter cone from Araguainha (from the contact zone between sedimentary collar and crystalline core. (g) Field work at a large exposure of impact melt breccia in the crystalline core of the Araguainha impact structure (Brazil). The clast content is monomict – only alkali granite clasts occur.

structure that eventually led to its confirmation as a true impact structure has also contributed much to the geological understanding of the wider region in the Kaapvaal Craton – besides the pursuit of impact-related issues. Another detailed review of a comprehensive, multidisciplinary investigation of an impact structure in Africa is the geological record of the Bosumtwi structure in Ghana, by Koeberl and Reimold (2005, also see below). A thorough review of structural geological investigations of impact structures has been published by Kenkmann et al. (2014).

A number of highly successful multidisciplinary drilling operations has been supported in the last 15 years by the International Continental Scientific Drilling Program (ICDP), in the Chicxulub (Urrutia-Fucugauchi et al., 2004), Bosumtwi (Koeberl et al., 2007a,b), Chesapeake Bay (Gohn et al., 2009), and El'gygytgyn (Koeberl et al., 2013) impact structures. Main aspects of analysis involved, *inter alia*, the presence and distribution of impactites, the macroscopic and microscopic analysis of mineral and rock deformation and its possible decrease down hole (i.e., into the crater floor), the presence and volumetric evaluation of melt phases, the origin of breccia components from the growing transient cavity (see below) or during the modification phase (crater collapse), hydrothermal activity in the crater and below in the crater floor, and even the search for evidence of post-impact development of life in the crater. Much has been learned from these projects about impact cratering and post-impact overprint on such structures.

Currently, a further consortium study on ICDP drill cores is underway, whereby drill cores retrieved from the Barberton Mountain Land of South Africa are investigated for information regarding early crustal and mantle processes, including tracking of the processes that led to the development of early life. Another aim of this project has been to retrieve fresh material of Archean spherule layers (see below, section 6.3) to possibly supplement the knowledge about these distal early impact ejecta obtained previously from surface and mine-derived materials.

2.1.4. Laboratory analysis

2.1.4.1. Petrography and petrology. Basic methods for the study of distal impact ejecta are described in detail in Montanari and Koeberl (2000). Optical microscopic analysis on high-quality, polished thin sections of rock samples from possible impact structures is the mandatory, first step of laboratory analysis. The prime objective is, of course, to identify the telltale, characteristic features of shock deformation (shock metamorphism, see below; presence of unusual high-pressure polymorphs) to confirm the existence of an impact structure. The level of deformation, if present, gives an indication of how deeply the structure may be eroded. Where optical microscopy is insufficient to confirm the presence of shock metamorphic features, electron microscopic, and even transmission electron microscopic, analysis may be required. In addition, Raman spectroscopy may be useful to investigate the presence of high-pressure polymorphs of impact-diagnostic value. The combination of optical contrast and SEM-based cathodoluminescence analysis (Hamers and Drury, 2011; Hamers, 2013) is also a powerful technique. Electron-backscatter diffraction (EBSD) analysis, coupled with electron microscopic techniques, has also been employed for detailed shock metamorphic analysis (e.g., Moser et al., 2011; Hamers and Drury, 2011).

Petrographic analysis of melt breccias will show which mineral and rock precursors would have contributed to a breccia "mélange". Evidence for melting is sought after, for the aforementioned importance of melt breccias for dating purposes and because melt rocks may provide a possibility to identify traces of an extraterrestrial projectile therein. Any attempt to determine the nature of breccia formation and emplacement in an impact structure (e.g., the suevite controversy, see below, and in Reimold et al., 2011a, and Stöffler et al., 2013) requires very detailed petrographic analysis. The same holds for the investigation of melt breccia petrology (e.g., Hecht et al., 2008). There has been quite some interest in the formation and emplacement of proximal and distal impact ejecta (e.g., Grieve et al., 2010; Stöffler et al., 2013; Artemieva et al., 2013; and the literature about the K–Pg boundary, and the Archean and Proterozoic spherule layers, recently reviewed by Glass and Simonson, 2013).

It should not be forgotten that a given type of information may provide crucial data for a further, different aspect of research, too. A good example is the multidisciplinary investigation of the ICDP drill core from Chesapeake Bay. Detailed petrographic analysis of the different lithologies provided a means to suggest their likely places of origin in the developing and modifying crater structure. This, in turn, was critical evidence that allowed to construct a numerical model for the multi-stage development of this structure (Kenkmann et al., 2009b).

2.1.4.2. Geochemistry. Detailed reviews of geochemical analysis of impact facies have been given by Koeberl (2007, 2014), French and Koeberl (2010), and Koeberl et al. (2012). This involves both elemental and isotopic analysis aimed at determination of the origin of lithologies found in a crater, as well as the identification of an extraterrestrial (i.e., meteoritic) component, a remnant of the projectile. The latter, if successful, serves as a definite criterion supporting the impact origin of a crater structure; and, of course, the nature of the projectile is important for understanding what has impacted Earth at various times throughout Earth evolution. Siderophile element analysis (comparison of elemental abundances in the impact breccia with the chemical compositions of the target rocks – the so-called *indigenous components* of elements that may be enriched in impact breccias due to the presence therein of extraterrestrial material) and especially the analysis of the platinum-group elements have been utilized since the 1970s. In recent decades this has been successfully supplemented by Re–Os and Cr isotopic methods, whereby the Re–Os method has superior sensitivity and may determine meteoritic components as low as 0.1%. In contrast, the Cr isotopic method is less sensitive (meteoritic components ought to be above 1%), but allows to constrain the meteoritic type of a projectile.

2.1.4.3. Chronology. For many years attempts have been made to date impact events (i.e., the structures that result from them) to improve our knowledge of how impact flux may have varied over time. Both terrestrial and extraterrestrial materials (Apollo samples returned from the Moon and lunar meteorites, Martian meteorites) have been investigated. In the last decades, since the onset of the debate about whether or not catastrophic impact events have caused, or contributed to, mass extinction events, a further incentive has been to match individual impact events with such paleo-biodiversity crises – or show that a multitude of impact events at specific times could have caused global – or at least regional – breakdown of the environment. However, biostratigraphic and radiometric dating techniques have, in many cases, not provided very precise ages. As discussed in detail by Jourdan et al. (2009, 2012), only a fraction of the known terrestrial impact structures (21 structures only, out of a total aggregate of known impact structures of about 184 – Jourdan et al., 2012) is currently dated at precisions of 1–2%. And any attempt to correlate impact events with distal ejecta or mass extinction related horizons (e.g., the globally observed K–Pg boundary layer with the Chicxulub structure, or the spherule layer occurrences in the northern hemisphere [Glass and Simonson, 2012, 2013; Huber et al., 2014] with Vredefort or Sudbury) demands excellent age precision. Continuing attempts of high-precision age dating are warranted – in conjunction with sensitive petrographic analysis of best-suitable impactites – which is mandatory for obtaining high-quality chronological results.

One technique of choice is the $^{40}Ar–^{39}Ar$ method, which may allow to separate inherited Ar (from the target rocks) from the signature of reset impact melt rock or another type of impact-generated melt rock known as pseudotachylitic breccia, and may also allow to identify post-impact overprint due to hydrothermal alteration or post-impact thermally-induced loss of radiogenic Ar. Where it is possible to analyze zircon and/or monazite crystals that have grown from impact-related melt phases, the U–Pb isotope technique, especially spot analysis by ion microprobe, may be the technique of choice. Due to the complex systems of precursor rock remnants, impact-related new phases, and post-impact overprint(s), many dating attempts have remained unsuccessful. To combine two or more techniques may improve the chance to be successful (Jourdan et al., 2009, 2012). The recent combination of in situ ion microprobe dating and EBSD micro-structural analysis of the specimen dated (Moser et al., 2011) also holds strong promise.

2.1.5. Experimental impact

Laboratory acceleration of projectiles onto targets of metal, minerals, or rocks is a powerful technique for investigating shock deformation effects produced in these materials under controlled physicochemical conditions, for comparison with those deformations produced in natural impact events. That shock deformation effects have been observed not only in samples from terrestrial impact structures but also in lunar rocks and in meteorites (including those derived from the Moon and Mars) demonstrates that they are characteristic of impact deformation. Shock experiments also allow the calibration of the onset of formation of specific shock effects with precise shock pressures, thus providing a means to investigate the attenuation of shock pressure in natural impact structures. Reviews of the techniques (e.g., Fig. 2b) and results of shock experimentation have been given by Stöffler (1972, 1974), Stöffler and Langenhorst (1994), and Langenhorst and Deutsch (1998). A compilation of shock effects vs. shock pressure for many rock-forming minerals is provided (Fig. 11g, below). Much of the shock experimentation of previous decades was done with single-crystal mineral targets, and there is extensive scope for continuing experimental shock deformation with rocks. Limited studies have been conducted with target materials pre-heated to temperatures typically observed in the upper and middle crust (e.g., Langenhorst et al., 1992 or Huffman and Reimold, 1996). A particularly important field is currently investigated by the MEMIN research group (Kenkmann et al., 2011a, 2013a). Upper crustal target rocks are often porous and wet sediments, and their shock behavior is so far not known very well. One highly important outcome has been that formation of diaplectic quartz glass and actual rock melt could be achieved in porous sandstone targets at low shock pressures of 5 GPa, instead of the shock regime of 30–50 GPa where these phases would appear in non-porous rock. This dramatic lowering of onset pressure for melt phase generation is achieved due to shock front interaction with the pore space, whereby shock pressures can be locally elevated by factors up to 6 times the nominal experimental shock pressure (Kowitz et al., 2013a,b). An application to naturally occurring shock metamorphism is the comparison of these experimental results with the glass-bearing arenites of the deeply eroded central uplift of the Oasis impact structure in Libya (see below, section 6.1.13).

2.1.6. Numerical modelling

Simulation of impact cratering with numerical modelling techniques (e.g., Pierazzo and Collins, 2004; Collins et al., 2012; also several papers in the special issues by Pierazzo and Herrick, 2004, and Herrick et al., 2008) is being done for two reasons: first, to carry out parameter studies of individual phases, i.e., to obtain snapshot-information about the cratering process, with different projectile types and sizes, impact velocities, impact angles, and other parameters. Such modelling is extremely useful in setting baselines for the formation of deformation effects that are actually observed in nature. Both processes related to the target volume and those related to the formation and dissemination of

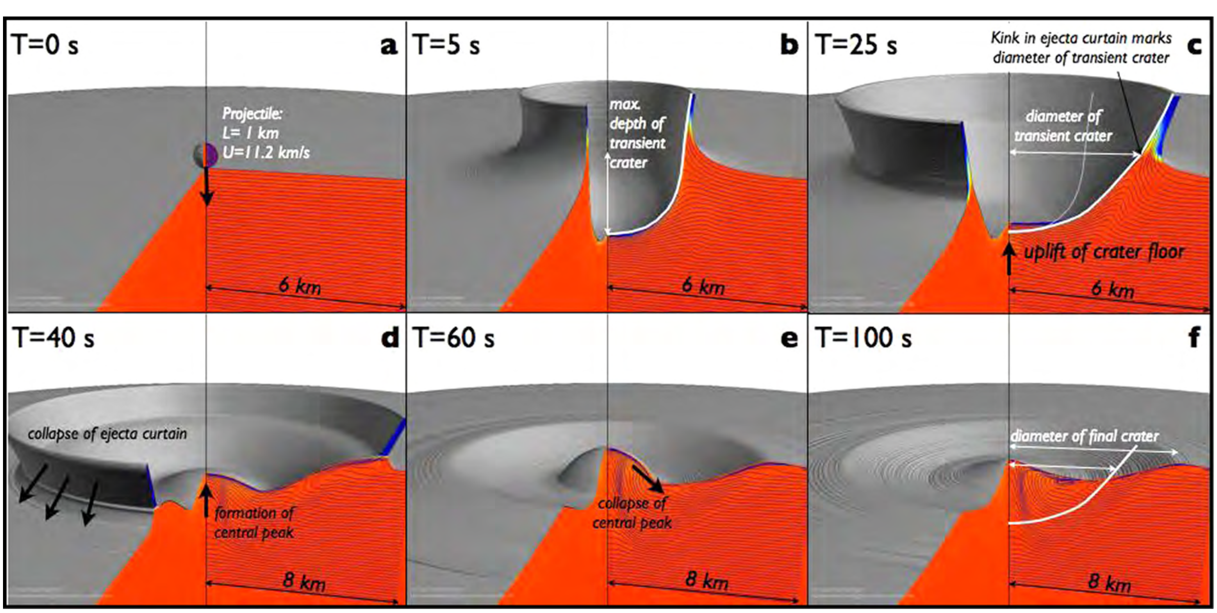


Fig. 4. Snapshot series of impact crater formation (courtesy of Kai Wünnemann, MfN Berlin; see also Wünnemann et al., 2011). (a) Initial stage just pre-impact. (b) Maximum depth of crater excavation attained; this corresponds to the maximum depth of the transient crater. (c) Diameter of transient crater has attained its maximum value, as indicated by the kink in the ejecta curtain. (d) Due to the effect of gravity, the crater floor rises upward; the ejecta curtain is deposited around the crater. (e) Beginning of collapse of the central uplift, with outward directed flow field. (f) Final crater morphology; the radius of the final crater is much larger than that of the transient crater (compare with (c)).

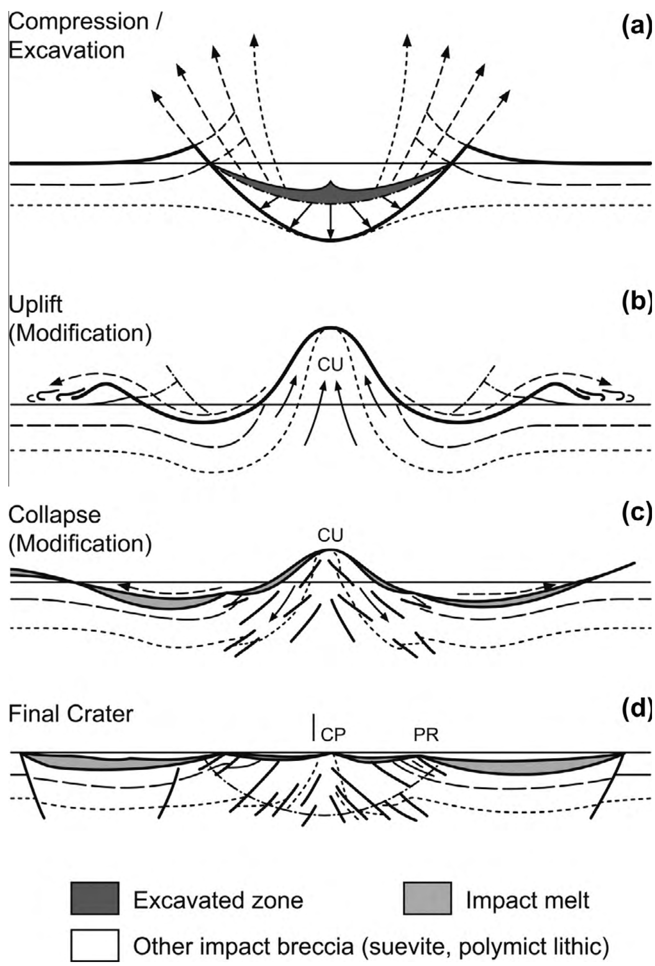


Fig. 5. Schematic representation of the formation of a complex crater (after Grieve, 1983 and Mohr-Westheide, 2011). (a) Compression/excavation stage; (b) uplift, and (c) collapse (modification stage); (d) final crater. The initial stage of excavation and compression relates to the formation of the transient crater (TC). The strength-degraded crater floor rebounds to form a central uplift (CU), while collapse of the transient cavity wall initiates an inward-directed material flow that combines with the upward-directed flow inherent to uplift formation. The CU collapses in stage c to form the central peak (CP) and – in even larger structures – the peak ring (PR) of the final complex crater morphology. As discussed in detail by Grieve et al. (2008, 2010), the Vredefort, Sudbury and Chicxulub structures are the only three terrestrial impact structures that presumably had a CP-PR (i.e., an incipient multi-ring basin) morphology.

impact ejecta (ballistic ejecta and impact plume materials) can be investigated by these techniques. The comparison of modelling results and field and laboratory findings is a powerful technique that has significantly enhanced our understanding of impact cratering, in general. And, what is more – on all scales! It is possible to set the basic parameters of numerical modelling, namely the cell size of target and projectile, to very different values, which allows to model energy distribution and material response at very different scales – from planetary-scale impact events down to the effects on hand specimen sized targets or even further to pore space scales. Ivanov (2005; see Fig. 2c) discussed the general procedure for modelling of impact cratering and then proceeded to review the modelling results for 5 large impact events – at Puchezh-Katunki (40 km diameter), Popigai (100 km), Chicxulub (180 km), Sudbury (200–250 km) and Vredefort (250 km). An example with several progressive steps in a modeled cratering experiment is shown in Fig. 4, courtesy of Kai Wünnemann (MfN Berlin).

3. The impact cratering process

3.1. The three stages of impact cratering

The mechanics of impact cratering has been described in detail by, for example, Grieve (1987), Melosh (1989, 2002), and Melosh and Ivanov (1999), and useful introductions to this topic are provided by French (1998), French and Koeberl (2010), and most recently, by Collins et al. (2012). One generally distinguishes – during the short interval of cratering – (i) contact and compression phase, (ii) excavation phase, and (iii) collapse and modification phase (after Melosh, 1989; French, 1998; see Fig. 5).

(i) Contact and compression phase: The process of impact cratering begins upon contact of the projectile with the target (Fig. 5A). Hypervelocity impacts on planetary surfaces typically occur at bolide velocities of tens of kilometers per second, with the average speed for asteroid impacts estimated at 15 km/s and that for comet impacts at 25 km/s. The main result of this initial phase is the transfer of the kinetic energy of the projectile to the target via a shock wave. The shock wave originating at the point of contact can reach peak pressures of many hundreds of GPa (1 gigapascal = 10 kbar). It propagates hemispherically through the target, as well as backward into the projectile. The pressures produced are much larger than the yield strength of either the target or the projectile, so that most of the projectile and part of the target at the sub-surface are vaporized.

The duration of the pressure pulse depends on the projectile diameter. Pressure release occurs when the shock wave reaches the back-end of the projectile and is reflected back as a rarefaction wave, which travels slightly faster through the target material than the initial shock wave. This causes unloading of the pressures in the material, whereby much of the projectile's kinetic energy is transferred into the target as thermal energy (shock temperature). Naturally, shock pulse durations in natural impact events are much longer than those generated in shock experiments with projectiles orders of magnitude smaller. Consequently, the kinetics of shock propagation are of vital importance and anybody comparing the effects of natural and experimental impact ought to bear this in mind.

The projectile will be transformed, almost entirely, to vapor, with the remainder being incorporated into impact-generated melt. Solid particles from the projectiles involved with natural impacts have been recovered on occasion (some tiny particles at Chicxulub, on the sea-floor from the Eltanin impact, and a decimeter-sized meteorite fragment in the Morokweng impact melt rock – see below), but this is exceedingly rare. Much of the impact melt may pool in the interior of the crater – in large events into crystalline targets resulting in coherent melt bodies (often of sheet geometry), and some is incorporated into ejecta (suevite outside of the crater, impact glass, tektites) in the form of small melt particles. The largest "melt bomb" ever found in suevite of the Ries crater measured some 2.5 × 0.5 m in size (A. Müller, unpublished information courtesy D. Stoeffler (MfN Berlin, 2012).

Generally, impacts into crystalline targets generate proportionally more melt than impacts into porous sedimentary rock, where much of the impact energy is expanded in closing the pore space while producing thermal energy. Detailed modelling of the generation and dissemination of melt from an impact into a composite (sedimentary/crystalline) target such as the Ries impact was recently presented by Artemieva et al. (2013). As highlighted above, the onset of transformation from crystalline material to diaplectic glass and/or melt is lowered in porous targets, as the interaction of

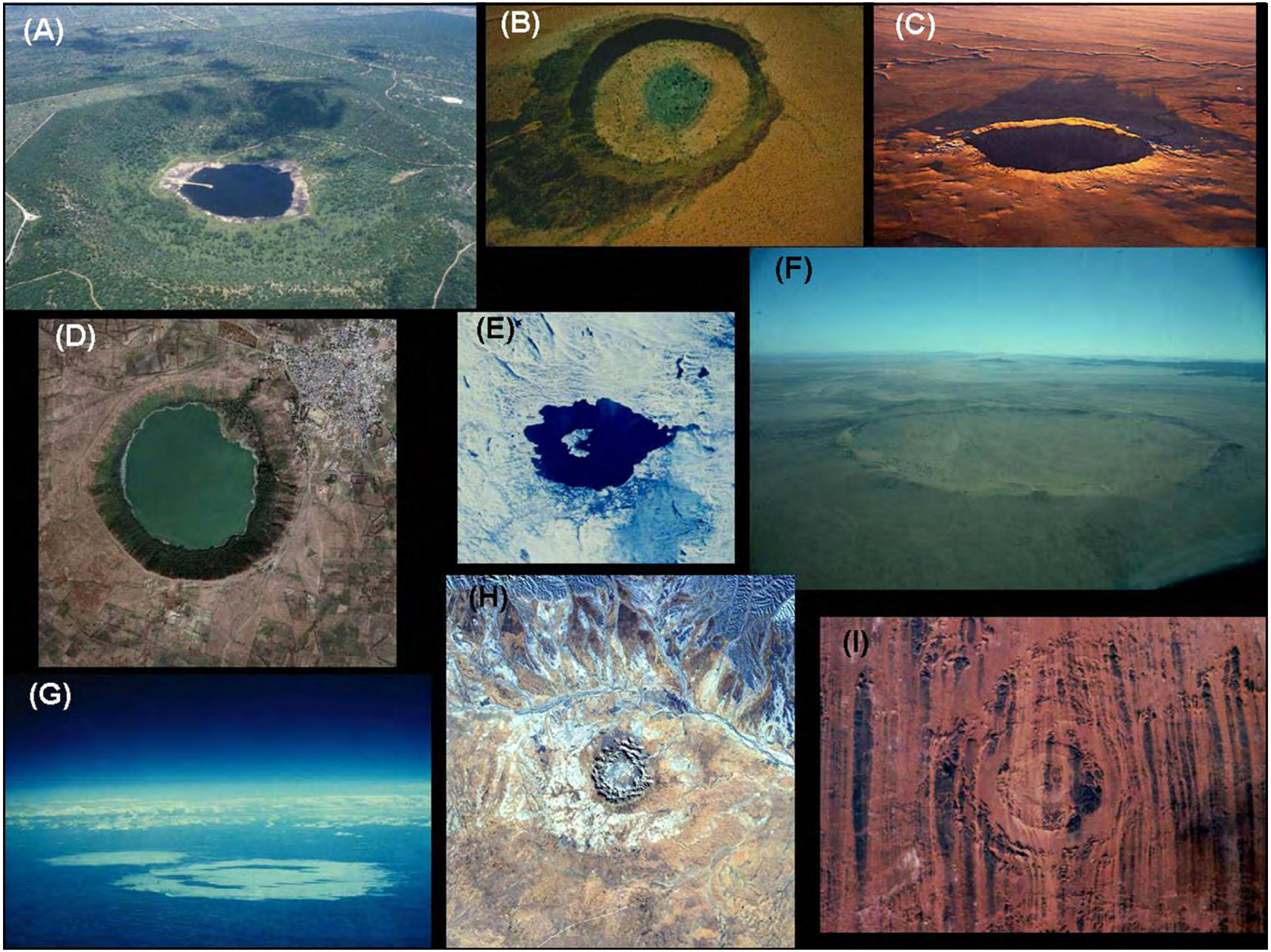


Fig. 6. Images of selected simple bowl-shape and complex terrestrial impact structures. The ones in the upper row and the left and right images in the center row are simple craters, the others are complex structures. Upper row: (A) Tswaing (Saltpan) crater in South Africa (1.2 km diameter, 250,000 years old); (B) Wolfe Creek crater in Australia (1 km diameter, 1 Ma old); (C) Meteor Crater, Arizona, USA (1.2 km diameter, 50,000 years old); (D) Lonar crater, India (1.8 km diameter, age ca. 0.5 Ma); (E) Mistastin impact structure, Canada (28 km diameter, age ca. 38 Ma); (F) Roter Kamm crater, Namibia (2.5 km diameter, age ca. 4 Ma); (G) Clearwater double impact structure, Canada (24 and 32 km diameter, age ca. 250 Ma); (H) Gosses Bluff structure, Australia (24 km diameter, age 143 Ma); (I) Aorounga structure, Chad (18 km diameter, age unkown).

shock wave and pore space may cause local shock pressure enhancement resulting in melt formation at comparatively low shock pressures. The unloading of the projectile from high pressure constitutes the end of the contact and compression stage.

(ii) Excavation stage: The excavation stage covers the interval during which the shock wave is expanded, along paths that transgress along a hemispherical front through the target rock, with the shock front weakening along its path down- and outward into the target. The combination of shock front and trailing rarefaction wave sets the target rock in motion in the form of a subsonic excavation flow; where the flow lines intersect the free surface (original target surface), they are bent upward, which leads to material ejection from the crater (Fig. 5b) into the atmosphere and onto the adjacent terrane. The final result of the excavation flow is a bowl-shaped "transient crater" of a diameter many times that of the projectile (typically 10–20× for terrestrial craters – Melosh, 1989), and an ejecta blanket of excavated rock surrounding the crater. The excavation stage lasts until the stress wave and rarefaction (or release) wave dissipate and the maximum diameter of the transient cavity is reached.

(iii) Modification stage: At the onset of the modification stage, the transient crater begins to collapse (Fig. 5c) under the effect of gravity, which produces a final crater morphology (e.g., Melosh, 1989; Wünnemann and Ivanov, 2003). Loose material along the rim of the transient crater slides downwards into the cavity. Simple craters are relatively small,

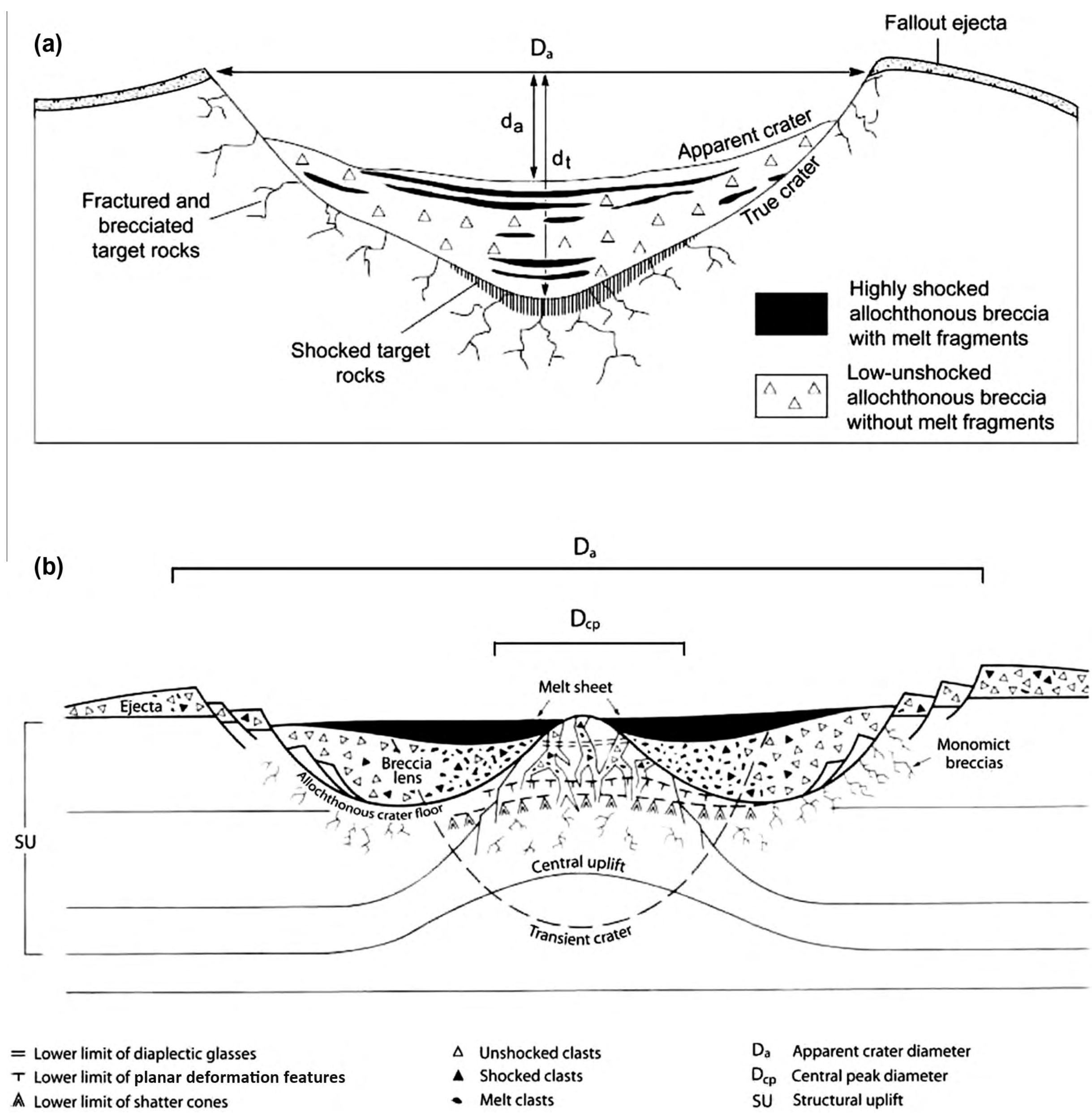


Fig. 7. (a) Schematic cross section through a simple, bowl-shaped impact crater (modified after French, 1998). D_a – apparent diameter; d_a – apparent depth; d_t – true depth. Note that the distribution of melt-bearing and purely lithic breccias in the crater interior is only exemplary and schematic. (b) Schematic cross section through a complex impact structure with central uplift, terraced crater rim, and extensive melt sheet. D_a – apparent crater diameter, after collapse of the transient cavity (crater); D_{cp} – diameter of central peak structure. Thin black lines indicate the warping of target stratigraphy in/below the inner crater. Also indicated are estimated occurrences of diaplectic glasses, planar deformation features, and shatter cones. Crater filling breccias are composed of a mix of unshocked and shocked clasts, with melt particles. Based on a diagram kindly provided by Dirk Elbeshausen (Museum für Naturkunde Berlin).

with a relatively smooth bowl shape. Thought to have had an original close-to hyperbolic shape, the slopes of the crater wall become more gentle during modification. Larger craters (see next section) depart further from gravitational stability, which causes the initially steep crater walls to collapse downward and inward, while a central uplift forms from a combination of the effects of this inward directed flow and elastic rebound of the crater floor (Fig. 1c), leading to complex crater geometries. The rim sections of large, complex craters are frequently characterized by large slump features, often appearing like multiple terraces on the inside of the final crater wall. Even small, simple (e.g., Tswaing – Brandt and Reimold, 1995a, 1999) or complex (such as BP – Koeberl et al., 2005a, or Bosumtwi – Reimold et al., 1998a) craters display complex structural geology in their rim sections. This may involve, inter alia, rim-parallel, radial or oblique (with respect to the center of the impact structure), and tangential faulting, with dips on fault planes variably inward or

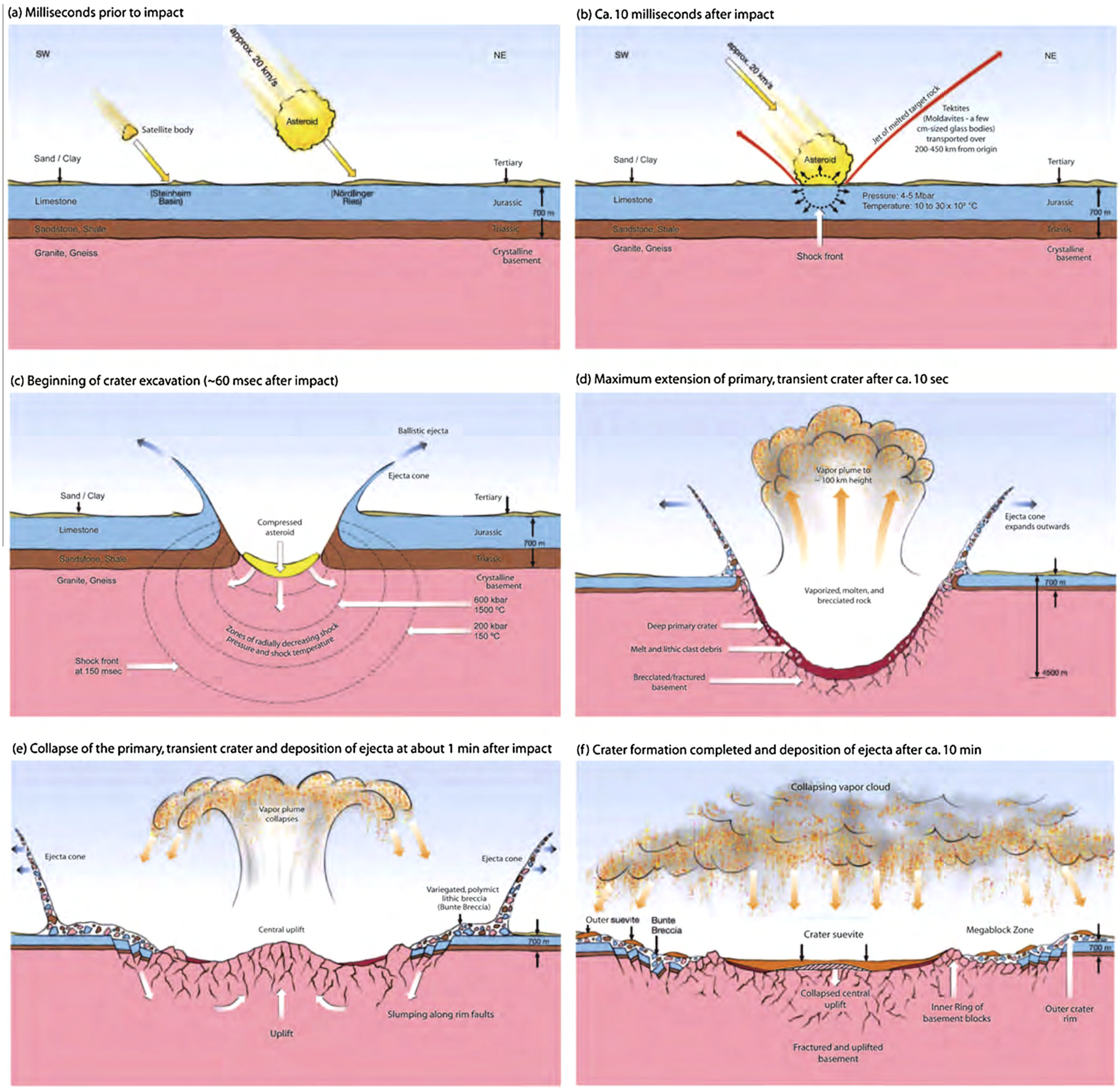


Fig. 8. A series of cartoons representing time steps in the early development of a moderate size, complex crater and its twin – similar to the Ries Crater and Steinheim Basin situation in southern Germany. (a) Just a few milliseconds prior to impact, showing the schematic target stratigraphy and the dual impactors. (b) Immediately (!) after contact with the target surface – the projectile penetrates the target volume and a shock wave develops. A jet of tektites is ejected from the horizon just below the target surface. (c) Crater formation has begun in earnest, and the subsurface target is becoming subject to the radially attenuated shock pressure and temperature conditions. (d) Some 10 s after impact a deep transient crater has been established, and the ejecta cloud/vapor plume has risen and begins to travel outward. (e) The transient crater has collapsed, material flow is inward directed from the edge of the crater, and upward in the inner sector. The vapor cloud begins to collapse while the ejecta curtain still travels outward. A central uplift is formed. (f) Only a few minutes after impact the impact crater is essentially completed. The central uplift has collapsed and – in this scenario – has been replaced by a peak ring. Deposition of vapor plume material is still ongoing. The crater is filled with debris of all shock degrees. Cartoons based on a series of drawings by Dieter Stöffler (Museum für Naturkunde Berlin).

outward directed. Furthermore, sometimes intense folding of strata, including stratigraphic duplication or elimination, can be a response to the dynamic and variably compressional or extensional forces involved with the different stages of cratering (compression to final modification). Further complexity and structural asymmetry is caused by oblique impact (e.g., Kenkmann et al., 2013b). Note that statistically a vertical impact would occur much rarer than an oblique impact. The final crater depth is extremely small in comparison with its width (on average, depths are no more than 1–3 km for a 100 km diameter crater; e.g., Ivanov, 2005). During formation of a central uplift, both extensional and compressional forces will be active; upon collapse of the uplift structure, extensional forces will be widely active but may, in turn, result in local compressional regimes.

The central uplift or peak ring structure is often more resistant to weathering than the surrounding crater fill of impact breccias and may, thus, be the only remnant of deeply eroded impact structures – e.g., Oasis, Libya – see below, or Gosses Bluff in Australia (cf. Fig. 6H).

In order to explain the phenomenology of complex craters, theories of hydrodynamic response of the target material to hypervelocity impact have been invoked. Thereby, rheological properties of Bingham fluids are assumed for material beneath collapsing craters. The principle thought is that shocked rock volumes are set into vibration – equivalent to a material flow. This model is known as "acoustic fluidization" (Melosh, 1983, 1996; Melosh and Ivanov, 1999), a process that allows the material to behave perfectly plastic with negligible internal friction and low cohesion during transient cavity collapse and uplift formation. Impact cratering workers have been debating how this fluidization process works at the atomic to grain scale in the affected rocks. For example, it has been discussed how it is possible for the Archean fabrics in the gneisses and migmatites of the core of the Vredefort Dome to be largely (with the exception of zones of pseudotachylitic breccia development and fracturing at meso- to micro-scales) preserved. On the micro-scale, however, pervasive microfracturing is noted in the rocks of the Dome. Cross sections through final simple-bowl-shape and complex impact structures are shown in Fig. 7.

Subsequent to the cratering process that, even in cases of formation of very large impact structures, may only last for 15–20 min from projectile contact with the target to completion of the collapse phase (e.g., Henkel and Reimold, 1998; Kenkmann et al., 2009b), the final crater structures become subject to normal geological degradation – erosion, tectonic overprint and even truncation, and finally burial by sediment or tectonic slices. Impact-induced thermal overprint of rocks (e.g., due to the vicinity of massive and superheated impact melt, or prolonged residence in hot impact breccia), tectonic adjustment of the wider crustal volume, and hydrothermal overprint may continue for many hun-

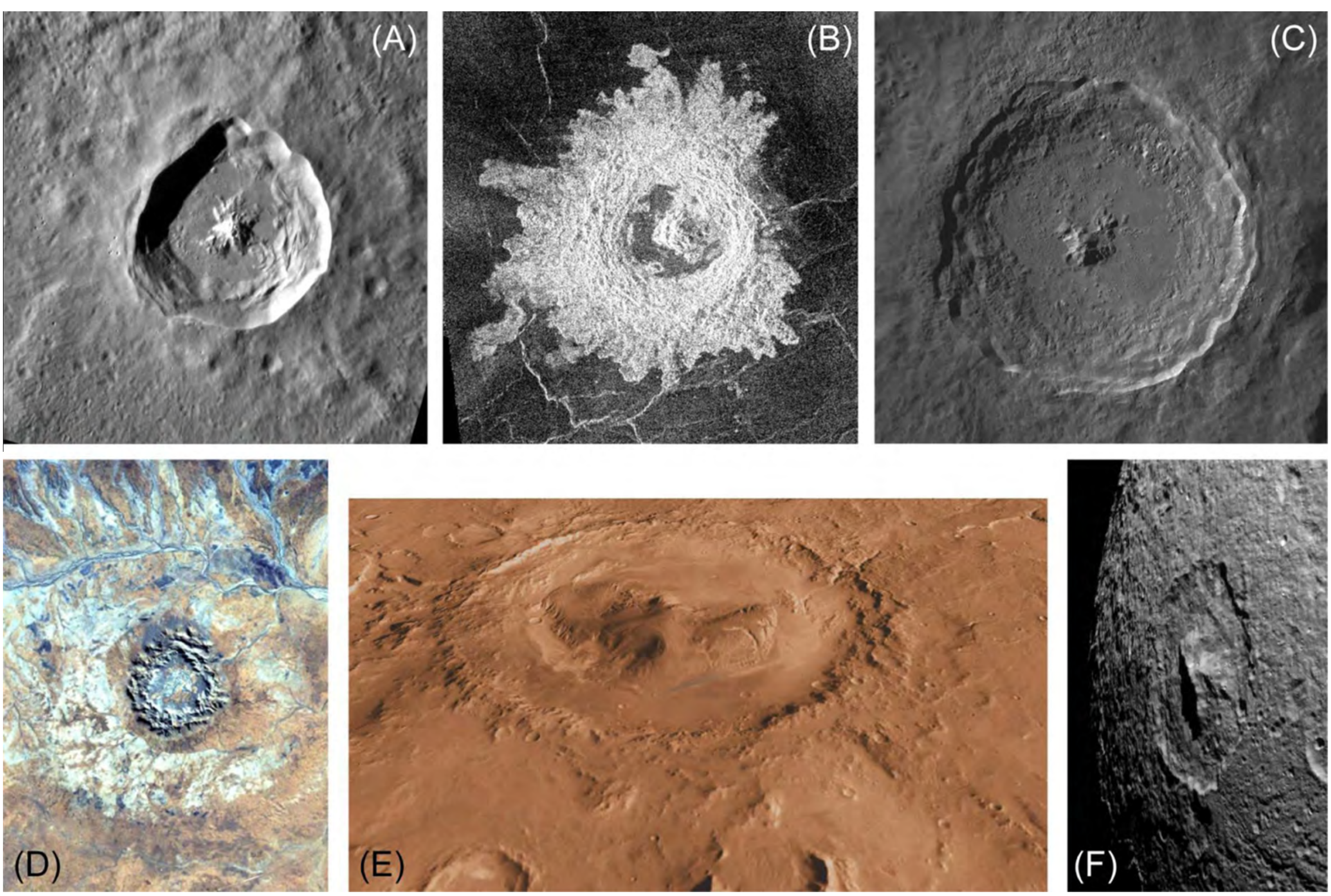


Fig. 9. Selection of complex impact structures on various bodies in the solar system. (A) the 35-km-diameter impact crater Popova on Mercury (Magellan image, NASA). (B) The complex crater Buck in the Navka region of Venus (22 km diameter) in a radar image by the Magellan spacecraft (NASA, showing also the ejecta surrounding the crater structure. (C) The 86-km-diameter impact crater Tycho on the Earth's Moon (NASA Lunar Reconnaissance Orbiter image). (D) The 24-km-diameter Gosses Bluff impact structure, Australia, showing mostly the 6-km-diameter deeply eroded central uplift as a ring-structure in its center (Landsat image). (E) The 154-km-diameter Gale crater on Mars, where the Curiosity rover landed in 2012 (Mars Odyssey orbiter imasge, NASA). (F) Erulus impact structure (120 km diameter) on the heavily cratered surface of Dione, one of the moons of Saturn (1120 km diameter), taken by the Cassini spacecraft narrow-angle camera (NASA).

dreds of thousands, even millions, of years, obviously with the duration being a function of the size of the impact structure, i.e., a function of impact energy, the volume of melt rock, and of fluids, available.

A series of cartoons (Fig. 8) depicting the general development of a complex impact crater – specifically developed for the case of the Ries crater of southern Germany – were recently prepared for the Ries Geopark by D. Stöffler and colleagues at the Museum für Naturkunde Berlin (see also Stöffler et al., 2013). Based on these time steps we are showing how such a moderately sized complex crater could have evolved. In the African context, the Bosumtwi or Luizi structures are broadly comparable to (although somewhat smaller than) the Ries crater of about 25 km diameter, although the different target geologies (sediment above crystalline basement at Ries; metasediment intruded by small granite stocks at Bosumtwi: Karoo sedimentary rocks above crystalline basement at Luizi) might have affected crater development somewhat differently at each of these structures.

Note that the processes concerning the final time step – development of the vapor plume and deposition of fallback into the crater and material mixing within the crater – are quite strongly debated at this time (e.g., Grieve et al., 2010; Stöffler et al., 2013; Artemieva et al., 2013). A most recent view (Artemieva et al., 2013), which is based on results of numerical modeling of crater growth and ejecta formation in a Ries sized impact event, is that the early vapor plume may be much bigger than shown in Fig. 8f and that it may only contain projectile-derived and sedimentary material. Maybe already at this time, and through the crater collapse phase, there may be a secondary plume – the result of additional processes within the crater, perhaps indicating *fuel–coolant interaction* between an impact melt body and superheating water. This secondary plume may be much denser than the primary, and may only reach a height of a few kilometers, spreading outward. According to N. Artemieva's calculations, the primary plume is buoyant from about 2 min after impact – but its collapse may last as much as several hours, during which time fallback is sedimented continuously and incorporated into the crater-fill impact breccias. The secondary plume propagates and collapses quite quickly, within minutes, but a major issue of discussion is when exactly the fuel–coolant inter-activity occurs, immediately after, or even during, crater collapse, or many years later. Also, possible sources of volatiles have remained debated. These models have been calculated for a typical Ries Crater size impact, and the authors have attempted to correlate modelling results and observations on the actual Ries impact breccia deposits. Clearly, there are still uncertainties based on partial incongruence. Like many aspects of the highly dynamic impact process, especially formation and emplacement of polymict impact breccias (see below) remain to be fully understood.

3.2. Morphological considerations

A detailed discussion of impact crater morphologies and the inherent nomenclature is given by Turtle et al. (2005). Impact crater morphology can be very different. It ranges from simple, bowl-shaped (Fig. 7a) to complex geometries with central uplift (peak) – Fig. 7b, with peak ring, with peak ring and central basin, and even with a central pit, to multi-ring basin morphologies (Fig. 9). In contrast to volcanic crater structures, impact produces generally circular, shallow (i.e., upper crustal), and rootless structures. Even at rather low angles of impact, as low as 20–30°, circular crater forms prevail (Elbeshausen, 2012). Due to erosion and post-impact tectonic overprint, the primary morphology can, of course, be significantly changed. A case in point is the Vredefort impact structure that extends over the entire Witwatersrand basin. The NE–SW extension of the basin is clearly the result of post-impact tectonics, likely due to the collision of southern Africa and Antarctica in Kibaran (Grenvillian) times. Also the Sudbury impact structure in Canada is extended from its original, more or less circular geometry, in NE–SW direction – again due to orogenic overprint during post-impact Grenvillian times. A distinctive feature of fresh impact craters is – in contrast to volcanic crater structures – that they have overturned (after some erosion, still strongly upturned) rim stratigraphy. The rims of volcanic craters are generally characterized by flat stratification or, at best, weak upturning of the uppermost strata.

Small (i.e., <2–4 km diameter on Earth, but on the Moon, at a gravity that is only one-sixth of Earth's gravity, <15–20 km) impact craters have simple, bowl-shape geometry. Their morphologies are frequently modelled as hyperbolae, although they, too, suffer to a degree from wall collapse so that the final crater shape, in comparison to the early transient crater form, is characterized by a somewhat wider and flatter geometry. The onset size for formation of complex craters is a function of target composition, with 2 km on Earth being the limiting value for sedimentary targets (e.g., the ca. 2-km-wide BP crater of Libya has a distinct central uplift structure of complexly deformed sandstones) and 4 km for crystalline targets. In contrast to BP that is entirely formed in sedimentary strata, the 3.6-km-wide Brent Crater in Ontario (Canada), which has been extensively explored by a series of drill holes, is formed entirely in crystalline basement and characterized by a simple, bowl-shaped geometry.

For complex craters, the nature of the target seems to influence how wide and high a central uplift structure may become. The 10.5 km wide Bosumtwi crater in Ghana, formed in metasedimentary and metavolcanic terrain (Birimian and Tarkwaian strata, respectively – see below) with minor, locally occurring granite intrusions, has a significant central uplift about 1 km wide and several hundred meters high, whereas the presence of a central uplift in the 25 km wide Ries crater of southern Germany remains unconfirmed despite extensive geophysical analysis (e.g., Wünnemann et al., 2005), although the work by Wünnemann et al. (2005) may indicate limited basement uplift underneath the crater, which then would imply that the uplift structure became seriously reduced in size due to collapse flow. The target comprised a >400 m sequence of Mesozoic sediment on top of older crystalline basement. Instead of a well-defined central uplift structure, the Ries crater features a so-called "inner ring" (also termed "crystalline ring") that is seemingly made up of crystalline blocks intercalated with impact breccias and sedimentary blocks, as recently seen in a shallow borehole into this ring at the Erbisberg location (Jung et al., 2011; Kruppa, 2013).

In the environs of variably sized crater structures, such as the 2 km wide BP impact structure in Libya or the larger Bosumtwi structure, conspicuous ring faults may occur. It has been discussed that they could have formed due to the downward and outward directed compression vectors, or alternatively, that they are the result of inward-motion of material in the course of central uplifting (Wagner et al., 2002). The low, inward-directed dip of the BP ring fault would conform to the latter hypothesis (Koeberl et al., 2005a).

Very large craters may exhibit concentric ring structures that allow classifying them as multi-ring impact basins (Spudis, 1993). Terrestrial candidates for this structural type are only Vredefort in South Africa, Sudbury in Canada, and Chicxulub in Mexico (Grieve and Therriault, 2000; Grieve et al., 2008). The nature of such multiple rings is still a matter of debate, with faulting or warping, or décollement structures having been prominent hypotheses. It has also been suggested that so-called 'superfaults' could be characterized by abundant pseudotachylitic breccia (Spray, 1997), but this has remained controversial and, in fact, unproven.

4. Recognition criteria for impact structures

As discussed above, morphology and geophysical anomalies may provide hints at the possible presence of an impact structure, but they do not suffice as proof. Contrary to this, evidence of *shock metamorphism* is diagnostic, and *chemical evidence* that demonstrates the presence of traces of an extraterrestrial projectile is similarly conclusive.

4.1. Shock metamorphism

Physical expressions of shock wave compression and immediately subsequent decompression are irreversible deformation effects, e.g., planar deformation features and high-pressure polymorphs (e.g., Stöffler, 1971, 1972, 1974; Stöffler et al., 1991 – for an introduction to shock effects in meteorites; Stöffler and Langenhorst, 1994; Langenhorst and Deutsch, 1998, 2012; Grieve et al., 1996; French, 1998; Gratz et al., 1992, 1996), in many rock-forming minerals. These deformation and transformation effects are collectively known as *shock (or impact) metamorphism*.

In Fig. 10a the pressure–temperature regimes of normal crustal metamorphism and of impact metamorphism are compared. It is important to remember that impact structures are mostly formed in upper crust, with only very large structures reaching into the middle or even lower crust. Normal crustal metamorphism rarely exceeds temperatures of 1000 °C and pressures of several kilobars (<1 GPa, i.e., <10 kbar), whereas impact metamorphic conditions can extend to many hundreds of gigapascals and thousands of degrees centigrade.

Shock metamorphism in an impact crater has a ‘‘progressive’’ character (compare Fig. 10b). Progressive refers to the continuous increase of shock and temperature conditions in direction towards the point of impact (i.e., the location where the projectile quasi explodes). Close to the point of impact, at generation of the shock front, temperature conditions are comparable to those on the surface of the Sun. Shock pressure may reach many hundreds, perhaps thousands of GPa. The bulk of affected target material will be vaporized instantaneously, as is much of the projectile. Somewhat further from the ‘‘epicenter’’, shock pressures and associated shock temperatures are sufficient to cause bulk melting of target rock, and still further away mineral melting (i.e., partial melting of target rock) will be possible. Finally, formation of diaplectic glass (syn.: thetomorphic glass – a phase that still has the long-range order of the precursor mineral but is isotropic due to abundant disruption of bonds at the atomic scale), of planar deformation features (PDFs) and high-pressure mineral polymorphs is reached, before the Hugoniot Elastic Limits (HEL) of minerals are reached, below which only elastic (brittle) mineral deformation is possible. Clearly the bulk of the impact affected rock volume is not shocked above the HEL of the mineral constituents. For this reason, extensive investigations are carried out on experimentally and naturally weakly shocked (i.e., <10 GPa) rock for possibly diagnostic impact deformation features (e.g., Kowitz et al., 2013a,b investigated sandstones of varied porosity shocked between 2.5 and 20 GPa). To date, compression features such as impaction of grains onto each other with associated radial microfracturing (concussion fractures), local cataclasis, planar fractures (PFs – fracturing akin to imperfect cleavage), and so-called ‘‘shock extension fractures’’ or ‘‘vermicular microfractures’’ (Buchanan and Reimold, 2002) have been described from this low shock regime. The impact-diagnostic value of these phenomena is still unclear. For example, it is not impossible that single sets of PFs be formed under tectonic conditions, but multiple sets of PFs of different crystallographic orientation formed in a quartz host grain would be atypical for tectonic deformation and may well be shown to signal an effect of impact.

In addition, so-called feather features (FFs) have been noted in quartz in rocks from many impact structures but have only been rarely produced experimentally (Poelchau and Kenkmann, 2011; Kowitz et al., 2013a,b). Note that the definition of FFs has been used differently in the literature: some workers refer to the combination of a longer planar fracture and the short, narrow-spaced ‘‘feathers’’ as FFs – others only mean these short and narrow fractures. No information of FFs found in tectonically deformed rock has been reported yet, which makes this a very promising shock-characteristic deformation phenomenon, but it has also not yet been fully investigated whether they may form under normal tectonic conditions. Whereas so far only one-sided ‘‘feathers’’ have been described in the literature, Zaag (2013) recently observed in a thin section from a shatter cone from the Serra da Cangalha impact structure (Brazil) a two-sided ‘‘feather’’ – resembling a feathered arrow-head. This is impossible to reconcile with the Poelchau and Kenkmann (2011) hypothesis that FFs are formed under shear stress. In Fig. 11, several important micro-manifestations of shock metamorphism, in different minerals, are illustrated.

A further important observation has been reported in a number of papers on impact structures formed in sedimentary targets. For example, French et al. (1974) reported tiny pockets of glass from shocked sandstone samples of the BP and Oasis structures in Libya. Kowitz et al. (2013a,b) made similar observations in experimentally shock deformed sandstone subjected to nominal shock pressures of 5–12.5 GPa. Shock melting was noted preferably in pockets originally filled with phyllosilicate minerals. Further investigation of these glasses is in progress, but already now it can be surmised whether such glass formations may be diagnostic for low-shock overprint of porous sedimentary rocks. Kowitz et al. (2013a,b) also observed the formation of diaplectic quartz glass in their weakly shocked sandstone specimens and the numerical modeling results in Kowitz et al. (2013b) provided an explanation in that interaction of a shock wave with pore spaces can cause local shock pressure increase by factors up to 4, thus reaching the normal pressure regimes for formation of diaplectic quartz glass (>30 GPa) and silica glass (>45 GPa) even at shock pressures as low as 5–10 GPa. A further type of melt produced in the low-shock experiments with porous sandstone concerns silica glass melt along shear fractures in the shock experimental sample assemblies, and are thought to reporesent the result of friction melting (Kowitz et al., 2013b).

PDFs – planar deformation features formed in a range of important rock-forming minerals, including quartz, feldspars, olivine) – are the most widely applied recognition criterion for shock metamorphism. Texturally they are absolutely straight (planar), crystallographically controlled features of 1–2 μm width and 2–10 μm spacings that may occur in parts of a crystal or traverse it penetratively. PDFs have been experimentally produced in quartz above 8–10 GPa (Huffman and Reimold, 1996; Stöffler and Langenhorst, 1994; and many others). Detailed scanning electron microscopy (e.g., Hamers, 2013) and transmission electron microscopy (e.g., Stöffler and Langenhorst, 1994; Gratz et al., 1996) have shown that these narrow features may be amorphous (glass) lamellae, high-density dislocation bands, basal Brazil twins, Dauphiné twins, or annealed equivalents. When PDFs are thermally overprinted (post-impact annealing), the original glassy phase becomes annealed and remains only recognizable due to the straight trails of fluid inclusions exsolved from the primary phase, which still mirror the original locations of PDFs.

Other impact-diagnostic thermal alteration features (Fig. 12a and b) are the presence of ballen quartz (Ferrière et al., 2009a, 2010a) and of toasted quartz (French and Koeberl, 2010; Ferrière et al., 2009b; Whitehead et al., 2002). Checkerboard feldspar

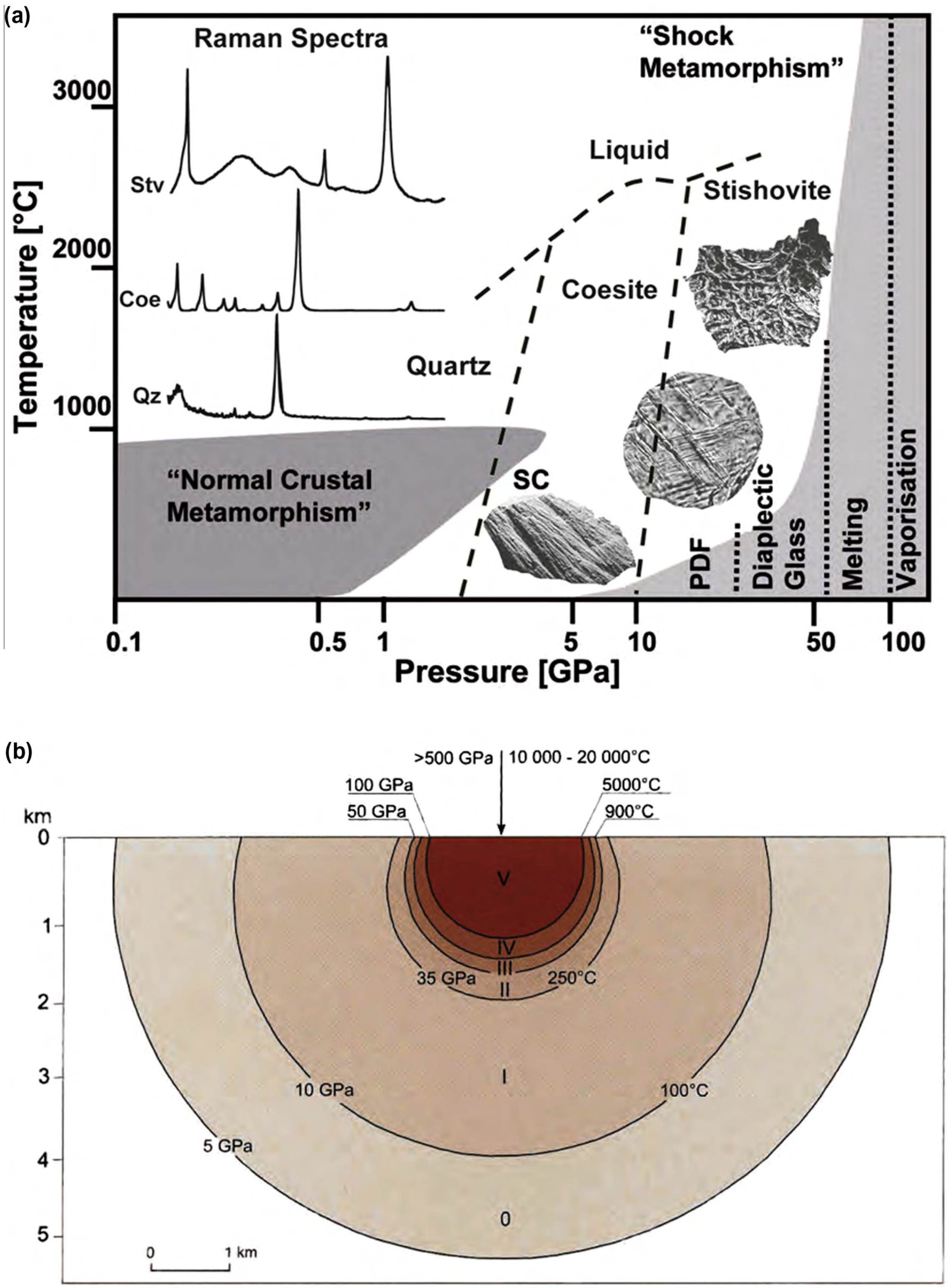


Fig. 10. (a) P–T diagram for normal crustal as well as shock metamorphism (modified after Reimold and Jourdan, 2012). See text for detail. (b) Schematic representation of progressive shock metamorphism along a radial cut through a simple bowl-shaped impact crater. Away from the point where the shock front originates from the quasi-"explosion" of the projectile, shock pressures and temperatures decrease with increasing radial distance. Different shock stages (0-V, according to Stöffler, 1971) are characterized by specific deformation effects: 0 – elastic deformation only, I – PDF development, II – diaplectic silicate glass and incipient melting, III – feldspar fusion, IV – bulk-rock fusion, and V – vaporization. Estimated shock temperatures are also marked at boundaries between shock zones. Based on a diagram in Reimold (2006), used with permission of the Executive Manager of the Geological Society of South Africa, Johannesburg.

(Fig. 12c and d), when occurring in abundance, may hint at impact deformation as well. An equivalent to checkerboard feldspar, occurring in quartz, was described by Buchanan and Reimold (2002). They called this crystallographically controlled micro-melting of quartz, according to the optical appearance of these narrow melt veins, "vermicular quartz". Based on the different shock textures produced at different shock pressure levels, shock metamorphic schemes have been set up for different minerals (e.g.,

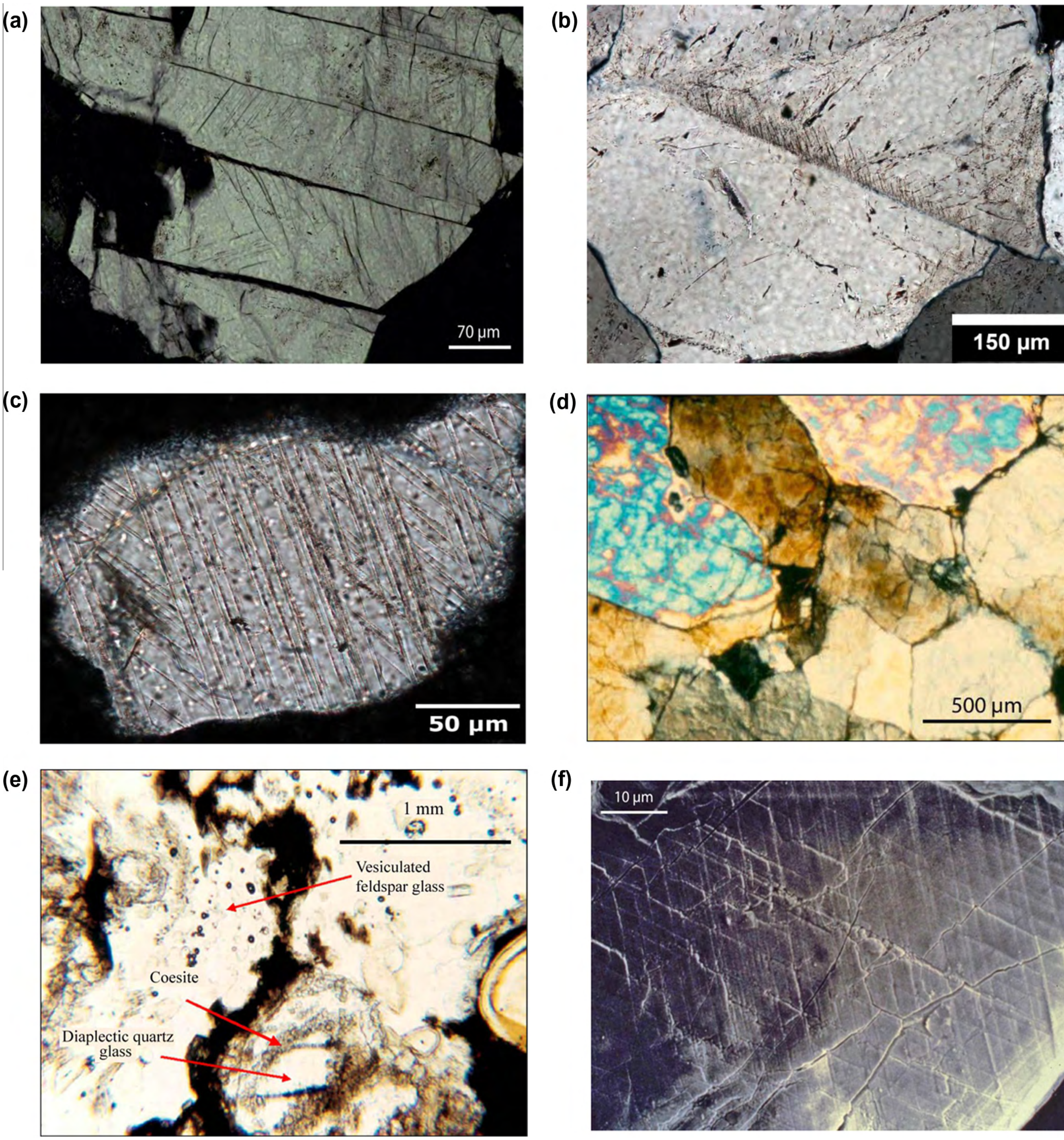


Fig. 11. Shock micro-deformation features. (a) Planar fractures and (short) planar deformation features of two orientations (NNE–SSW, ENE–WSW) in quartz of a granite sample from Alerheim, Ries Crater (Germany). Cross-polarized light. Image courtesy of Ralf-Thomas Schmitt (MfN Berlin). (b) Feather features (comprising a relatively long planar fracture and "feathers") in quartz of a specimen from Haughton impact structure (Canada). Courtesy of Ludovic Ferrière (NHM Vienna). (c) Two sets of PDFs in a quartz crystal of a specimen of impact breccia from Mien (Sweden), Cross-polarized light. Courtesy of Ludovic Ferrière (NHM Vienna). (d) Mosaicism in experimentally shocked (29.3 GPa) dunite (Reimold and Stöffler, 1978). Note the irregular extinction in the individual olivine grains that have obtained this "mosaic" look due to slight re-orientation of tiny lattice domains. (e) Microphotograph of an impact glass particle in suevite from the Ries Crater, Germany. The area shown displays light-colored, locally vesiculated feldspar glass and dark-brown oxidic remnants after a mafic precursor mineral. A clast of diaplectic quartz glass contains aggregates of tiny coesite crystals. (f) Backscattered electron image of a shock-metamorphosed zircon crystal from a melt rock sample from the Vredefort impact structure. Two distinct systems of planar fractures are clearly recognizable. This kind of shock deformation in refractory minerals such as zircon or monazite has proven invaluable to identify bona fide shock evidence in very old impact structures.

quartz – Stöffler and Langenhorst, 1994) – and then for different rock types (e.g., Stöffler and Grieve, 2007). Table 1 (after Reimold, 2006) shows such generalized schemes for quartz and feldspar, and in Fig. 11g progressive shock metamorphism is reviewed for a large number of rock-forming minerals.

The mineral quartz is the preferred study object for the search of shock metamorphic evidence. Not only is quartz amongst the most abundant minerals in the upper crust, but this mineral is also relatively resistant to weathering/hydrothermal alteration and metamorphic overprint. Above all, quartz displays a range of shock

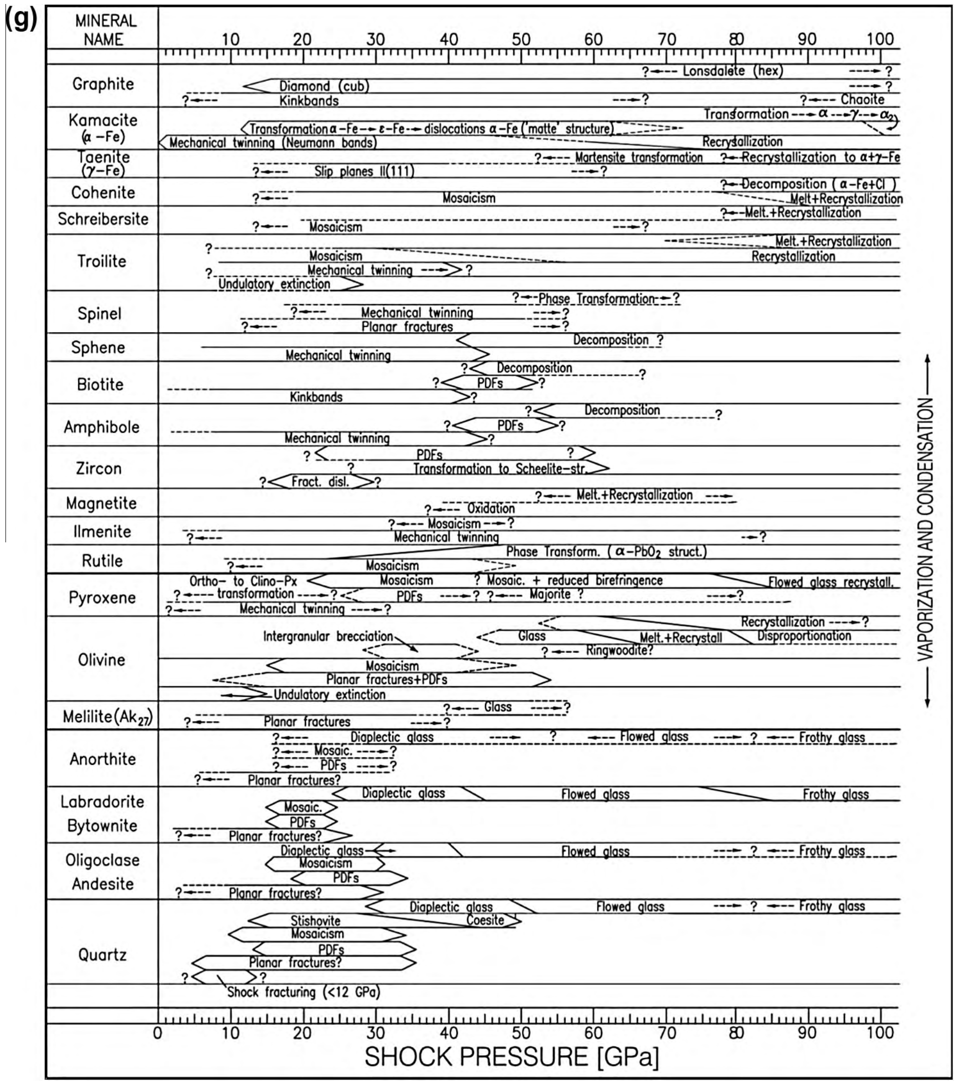


Fig. 11. (g) Shock metamorphism for a large number of rock-forming and accessory minerals, as calibrated through numerous shock recovery experiments.

metamorphic effects that are shock barometers well calibrated by shock experiments: PFs, mosaicism, PDFs, diaplectic quartz glass, high-pressure polymorphs, and finally quartz fusion. PDF analysis is, furthermore, facilitated by the uniaxial character of this mineral that allows straightforward determination of crystallographic orientations by universal-stage or spindle-stage analysis.

At elevated shock conditions, certain minerals may be transformed to otherwise rare or non-existent high-pressure polymorphs. Coesite and stishovite have been found in many impact structures, with coesite observed under crustal conditions as well (in kimberlite and in rocks associated with subduction zones) but stishovite being only known from impact structures. Reidite, the high-pressure polymorph with scheelite-structure after zircon has also been observed widely in impact structures. Its formation requires shock pressures between 20 and 40 GPa (Reimold et al., 2002b). At even higher shock pressure/shock temperature conditions zircon may melt and then crystallize again to the so-called granular or 'strawberry' shock texture. Monazite shows similar behavior (Deutsch and Schärer, 1994; Moser et al., 2011). At even higher impact conditions, zircon will dissociate to silica plus baddeleyite (ZrO_2). Where target rock contains graphite, chances are that this mineral may be converted to shock diamond or lonsdaleite. Particular abundance of these phases is known from the so-called 'impact treasury' of Russia, the Popigai impact structure in northern Siberia. A host of high-pressure polymorphs is also known from shocked meteorites, including high-pressure phases of feldspar, pyroxene, olivine, and other minerals (e.g., Gillet et al., 2007).

It must be expressly emphasized here that shock deformation is a very heterogeneous phenomenon (shock heterogeneity) due to the complex behavior of compression waves in a heterogeneous, multi-particle and – possibly – texturally complex rock. Waves may be reflected, scattered, or refracted, resulting in shock attenuation or amplification. Unshocked and highly shocked grains may occur in shocked target rock right next to each other, with assemblages of grains of all shock stages in microscopic volumes being possible. This holds for shocked and unshocked particles in impact breccias (next section) and shocked rock, as well as clasts in impact melt rock, which normally exhibit a wide range of shock metamorphic conditions, from unshocked to partially or wholly melted.

Different minerals display different affinity for multiple shock effect generation. As mentioned, quartz is considered the most useful shock barometer that has a distinct sequence of shock effects

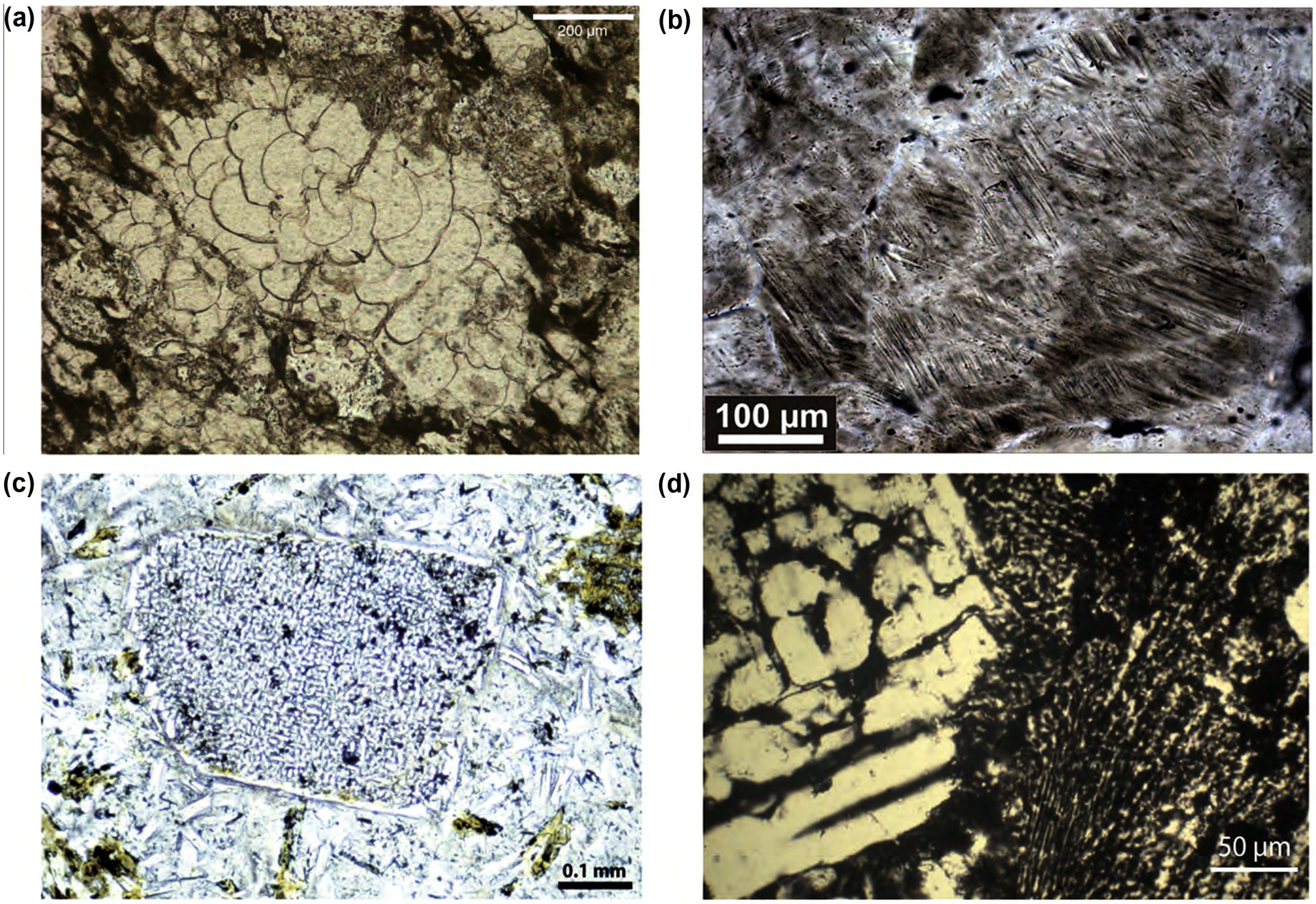


Fig. 12. Shock-induced thermal alteration. (a) Ballen cristobalite in suevite from the Bosumtwi impact structure (Ghana), plane polarized light. Courtesy of Ludovic Ferrière (NHM Vienna). (b) Toasted quartz in an impact breccia from Bosumtwi crater. Plane polarized light. Courtesy of Ludovic Ferrière (NHM Vienna). (c) Checkerboard feldspar, a clast in a well-crystallized sample of impact melt rock from the Lappajärvi impact crater (Finland). Note the typical H-shape plagioclase crystals in the matrix that indicate crystallization under relatively fast cooling. The plagioclase clast has been partially melted by the superheated impact melt, with melt channels developed predominantly along crystallographic orientations. The continuous outer rim of the clast indicates reaction with the surrounding melt. Image courtesy of Dieter Stöffler (MfN Berlin). (d) A second example of a checkerboard feldspar within a largely melted granitic clast in Vredefort impact melt rock (the so-called Vredefort Granophyre). Note the crystallographically controlled narrow zones of feldspathic melt (dark grey) between rectangular remnants of the original crystal. The right part of the image shows garben texture of finest-grained crystals grown from the melt matrix of the Granophyre.

covering the range of shock metamorphic conditions from <10 to >50 GPa (Fig. 11g). In contrast, pyroxene does not allow correlating deformation effects with different levels of shock pressure. Intra-granular fracturing, enhanced cleavage, and some twinning are the only shock effects known for this mineral, and their impact-diagnostic value is, at best, limited. Zircon and monazite are also very useful: due to their strong refractivity, they are the minerals of choice when searching for shock deformation in old, Proterozoic and Archean, rocks that may have been subjected to polymetamorphism and extensive alteration. In addition, these two minerals are highly useful for U–Th–Pb-based geochronology.

In accordance with the different shock behaviors of minerals, different rock types also display different shock behavior. Two examples of the sequences of progressive shock deformation, for a basaltic rock and a granitic rock, are given by Stöffler and Grieve (2007), and progressive shock deformation for mafic, pyroxene- and olivine-rich meteorites are discussed, for example, by Stöffler et al. (1991).

Table 1
Shock metamorphic effects in quartz and feldspar, in relation to increasing shock pressure.

Shock pressure (GPa)	Quartz	Shock pressure (GPa)	Feldspar
<10	Irregular fracturing, planar fractures	<10	Irregular fracturing, weak mosaicism
8–10	Onset of PDF development		–
10–25	Moderate to strong mosaicism	8–20	Moderate to strong mosaicism
10–30	Irregular and planar fractures	8–17	Irregular and planar fractures
15–30	Multiple sets of PDF → beginning isotropisation	15–30	Multiple sets of PDF → beginning isotropisation
25–35	Diaplectic glass	15–30	Diaplectic glass
15–45	Stishovite formation	>40	Feldspar melting
27 → 60	Coesite formation		
>45	Quartz fusion (lechatelierite)		

Note: At shock pressures in excess of 50 GPa bulk rock melting becomes important. The limits of the shock pressure regimes quoted are strongly dependent on the pre-shock temperature of the target rock. The information tabled is based on the findings of Huffman and Reimold (1996) and Grieve et al. (1996), and earlier works referenced therein. A comprehensive scheme of shock pressures vs. deformation effects for granitic and basaltic rocks is found in Stöffler and Grieve (2007).

Where they are available, it is desirable to sample impact breccias (next section) for investigation of possible evidence for shock metamorphism. Where, however, only monomict clastic breccia or unbrecciated rock can be sampled, it may still be possible to detect shocked minerals – albeit in comparatively lower abundance. Such breccias are generally formed at much lower shock pressure in the outer reaches of an impact structure (such as the crater floor and crater rim), and it may well be that shock deformation is only notable where the shock front may have been scattered/reflected at heterogeneities.

It must also be carefully observed that there can be other mineral deformation styles that are not impact-diagnostic, such as non-planar features (including: fractures, fluid inclusion trails, deformation bands) of tectonic origin. Great care must be taken not to confuse these with bona fide shock microdeformation. The literature is full of erroneous reports of shock deformation, which unfortunately is continuously promulgated due to still persisting widespread lack of proper tuition about impact deformation and the requisite methodology for their investigation. A recent example for this is the report of a possible impact structure at Maniitsoq (Greenland) by Garde et al. (2012, 2013). Despite repetitive publication of their alleged PDFs from this site, the impact-relevance of these alleged shock features did not improve. Reimold et al. (2013a) discussed the hypothetical nature of a Maniitsoq structure in some detail.

In the course of recent studies of quartz-rich samples from a newly discovered impact structure in Brazil (Santa Marta: Uchôa et al., 2013), A. Kowitz and W.U. Reimold (Museum für Naturkunde Berlin) recently investigated planar and seemingly non-planar deformation features in quartz with the application of a universal-stage. They noted several – in comparison with PDFs – broad and dense fluid inclusion trails. When tilting the thin section on the U-stage by ca. 20°, these features were revealed as well-defined and optically recognizable PDFs. Clearly, the features were inclined at a very low angle (<10°) to the thin section plane, so that essentially a side-on view of them was generated when looking straight down onto the section. Thus, some features that with normal optical microscopic observation will be – and have been in the past! – discarded as non-impact specific evidence may have been erroneously discredited as shock-diagnostic evidence – but only because the proper analysis by universal stage (or TEM analysis) was not applied. On the contrary, there are cases of previously published supposed PDFs that must be readdressed by proper methodology.

Reliable mega- to macroscopic shock deformation occurs only in the form of *shatter cones* (Fig. 13a and b), centimeter to meter-sized cone-shaped fracture phenomena, where striations diverge in a ridge-and-groove pattern from a small apical area. Such apical areas may measure just a millimeter or two in width, or may involve centimeter- to decimeter-wide planar areas. Striations may also converge towards a fracture plane that acted as generation plane. Shatter cones are known from a large number of impact structures and as they have not been observed in any other settings, are considered diagnostic of impact structures. However, care must be taken in their recognition, as some other fracture phenomena (e.g., plumose fractures, or cone-in-cone structures of sedimentary origin (Fig. 13e), cone fractures – so-called percussion

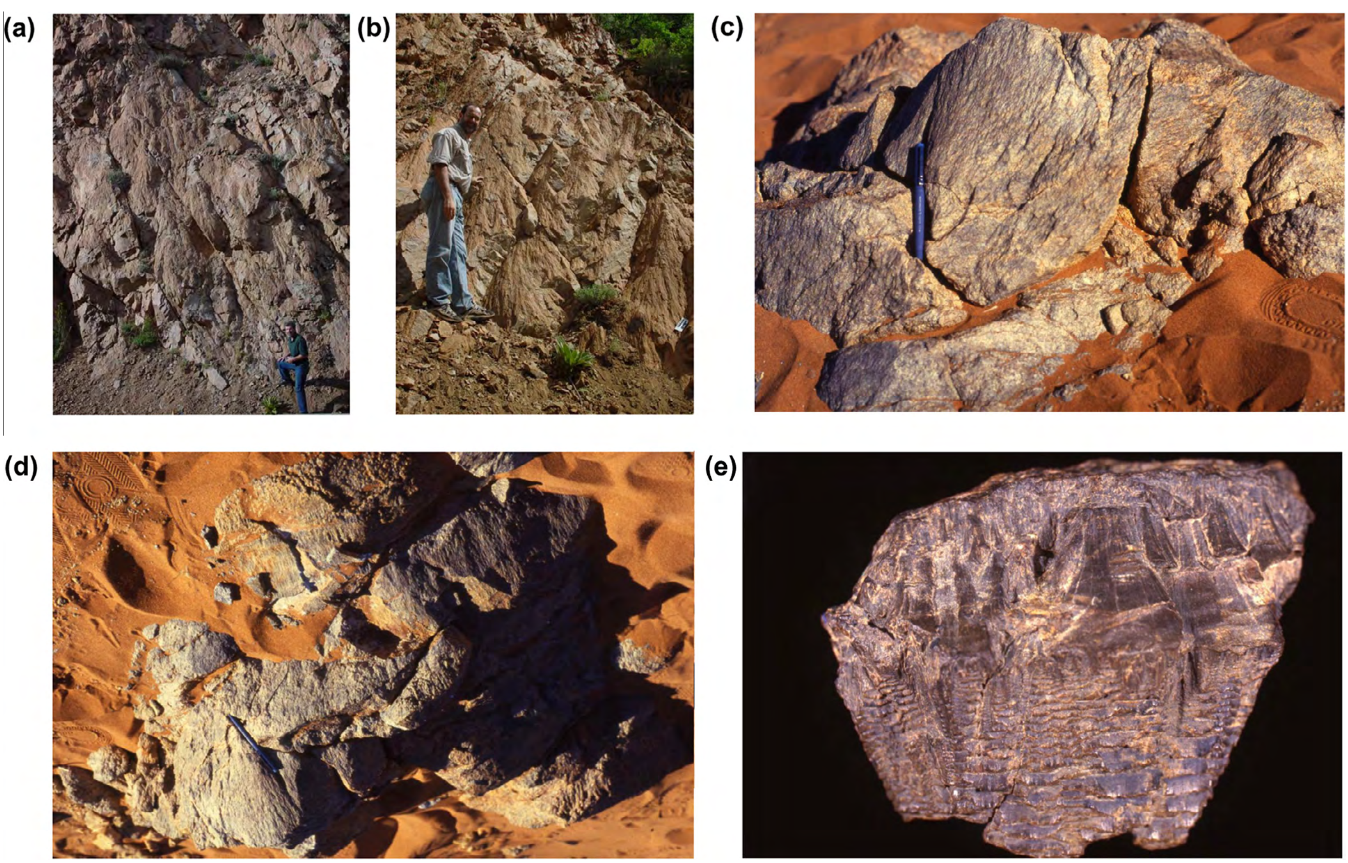


Fig. 13. Comparison of shatter cones and other striated fracture phenomena. (a) An exceptional outcrop with very large shatter cones in the vicinity of Santa Fe (New Mexico, USA). The late impact researcher Jared Morrow for scale. (b) Another location with very well developed shatter cones in Santa Fe, with Christian Koeberl for scale. Note the well displayed relationships between joints and cone structures in (a) and (b). (c, d) Wind from a consistent direction and loaded with sand of the Namib Desert has created these parallel to fanned grooves (ventifacts) on granitic basement on the crater rim of the Roter Kamm impact crater (Namibia). Pen for scale about 14 cm long. (e) Sedimentary cone-in-cone structure (cf. Lugli et al., 2005) in silicified shale from the Hamada region southwest of Erfoud (Morocco). Note the striated cone features in the upper part of the specimen. The specimen is about 8 cm wide.

marks – at the base of fossil water-falls, wind-ablation features in desert environments, such as those shown in Fig. 13c and d) may be mistaken for shatter cones and, at times, have erroneously been applied to suggest the presence of impact structures (for comparison of sedimentary and impact cones, see, for example, Reimold and Minnitt, 1996; Lugli et al., 2005).

A number of hypotheses have been promoted for the origin of shatter cones, including shock wave scattering or refraction on heterogeneities such as pebbles, pores, or fractures; interference of multiple joint sets of distinct curviplanar geometries. These ideas have been discussed most recently by Sagy et al. (2002, 2004), Baratoux and Melosh (2003), and Wieland et al. (2006), but the issue is still far from resolved. Planar deformation features have been found in many shatter cone-bearing rocks, and Nicolaysen and Reimold (1999) reported the presence of microscopic glass development on fractures inside of a shatter cone specimen – thus, suggesting that a shear component was involved in the formation of such striated fractures, resulting in local friction melting. Nicolaysen and Reimold (1999) also drew attention to the apparent relationship between shatter cones and more linear arrangements of striations on fracture surfaces (actually already recorded by Manton, 1965). They mapped out up to a dozen different sets of multipli-repeated curviplanar fractures in specific rock volumes in the Vredefort Dome. Nicolaysen and Reimold (1999) referred to these fracture sets with spacings between subparallel and often curviplanar fractures of millimeters to 1 cm as MSJS (= Multipli-Striated Joint Sets). Their detailed fieldwork resulted in the observation that orientations of shatter cone striations seemingly are related to MSJS trace intersections on stereoplots. To date, the actual relationship between shatter cones and MSJS has, however, not yet been clarified.

Small shatter cones have also been produced experimentally. The shock pressure regime of shatter cone formation has been estimated to range from <5 GPa to some 30 GPa. The range from 15 to <2 GPa is, for example, encompassed by the widespread occurrences of shatter cones in the Vredefort Dome (Wieland et al., 2006). More recently, Zaag et al. (2011) and Zaag (2013) applied computer tomography in order to investigate the 3D distribution of microfractures in a shatter cone (i.e., they attempted to gain further insight into the shatter cone-MSJS relationship) – with limited success due to the limited contrast achieved for their sandstone sample with the available micro-CT instrument.

The largest manifestations of shatter cones are known from the Slate Islands impact structure in Canada (Sharpton et al., 1996) and from a road cut in a suburb of the city of Santa Fe (New Mexico, USA – Fackelman et al., 2008), with both locations having exhibited meter-sized occurrences. The largest cone feature known to the authors from the Vredefort Dome (South Africa) measured some 45 cm in length and about 35 cm in width, but the overwhelming majority of shatter cones at Vredefort are <10–20 cm in size.

While it is desirable to have shock metamorphic indicators that are easily observed, preferably by widely accessible optical microscopy, it has been frequently necessary to revert to not as readily available, sophisticated but time-consuming transmission electron microscopy (TEM) to resolve features not explainable by optical microscopy alone or that have been controversial as shock features or tectonic deformation, or where the origin was obscured due to metamorphic overprint. Recent advances of scanning electron microscopy (SEM) and associated techniques such as color (or color composite) cathodo-luminescence (CL) study or electron backscatter diffraction (EBSD) have lead to an enlarged arsenal for shock metamorphic investigations. Another advantage of SEM applications is that the study of polished thin sections with such an instrument allows immediate correlation of optical and electron optical microscopic observations on the same specimen. Pioneering work in this regard on quartz has been carried out by Hamers (2013; also Hamers and Drury, 2011). They found that PDF are readily distinguished from tectonic deformation lamellae by both limited wavelength grayscale and composite color SEM-CL analysis. PDFs are, thus, revealed as straight, narrow, well-defined features, whereas tectonic features are comparatively thicker and/or slightly curved, and often the boundaries to host quartz are not clearly observable. Two types of CL response were noted for PDFs: red to infrared, or non-luminescent. CL signals of tectonic lamellae range from blue to red. Several causes for the red luminescence of PDFs are discussed by the authors. Hamers (2013) also investigated SEM-CL signatures of amorphous PDFs and healed PDFs, respectively. Amorphous PDFs turned out non-luminescent, whereas healed PDFs or basal Brazil twins yielded red luminescence, with dominant emission of 650 nm. This red luminescence was interpreted as due to damage caused by the electron beam preferentially along dislocations, fluid inclusions, and twin boundaries.

Electron-backscatter diffraction (EBSD) analysis, coupled with electron microscopic techniques, has also been employed for detailed shock metamorphic analysis of zircon by, e.g., Moser et al. (2011), Timms and Reddy (2009), and Timms et al. (2011). SEM techniques now have the potential to bridge the gap between optical and transmission electron microscopy.

4.2. Impactites – a recommended nomenclature

Within and around impact structures, a series of lithologies formed as a direct consequence of the impact event occur (prior to erosion). Stöffler and Grieve (2007) published a detailed nomenclature and definitions for these materials that are generally known as *impactites*. These authors distinguished proximal – occurring within and just around a crater – and distal impact facies. The latter comprises ejecta that can be distributed – according to the magnitude of the impact event – regionally around a crater structure, or on continental and even global scales. This includes the tektite and microtektite occurrences of many hundreds to thousands of kilometers extent. Four tektite strewn fields are currently known on Earth (Fig. 14), i.e., the Central European (moldavites), the Australasian (including microtektite occurrences in Antarctica), the Ivory Coast, and the North American strewn fields (Simonson and Glass, 2004; Glass and Simonson, 2012, 2013). The moldavites are related to the Ries impact, the Ivory Coast tektites to the Bosumtwi impact in Ghana, and the North American tektites to the Chesapeake Bay impact at the eastern seaboard of the USA. The Australasian tektites have not been correlated with a proven impact structure yet but are believed to originate from Southeast Asia; microtektites occur as part of this strewn field at distances of up to 6000 km, in Antarctica, from the inferred place of origin in Indochina. Several sites in Cambodia have so far been investigated in vain.

Secondly, distal impact ejecta comprise regionally or globally occurring ejecta, the best-known global occurrence of which is related to the K–Pg boundary. This also includes the so-called Archean and Proterozoic *spherule layers*, distal impact ejecta related to unknown, and likely completely obliterated, impact structures. Glass and Simonson (2012, 2013) provided a comprehensive account of the currently known distal impact ejecta deposits. Geoscientists should be aware that there could be many more such marker horizons in Earth's stratigraphic record, knowledge of which would improve the terrestrial impact cratering record tremendously.

Amongst the proximal impactites, one distinguishes *shocked rocks* and *impact breccias*. The former are rocks that have generally maintained their texture and fabric but contain shock metamorphosed minerals. And among the impact breccias, one categorizes *monomict breccias* that are nothing else but shattered target rock, equivalent to tectonically produced cataclasite, but perhaps carrying shocked minerals. *Polymict impact breccias* are classified

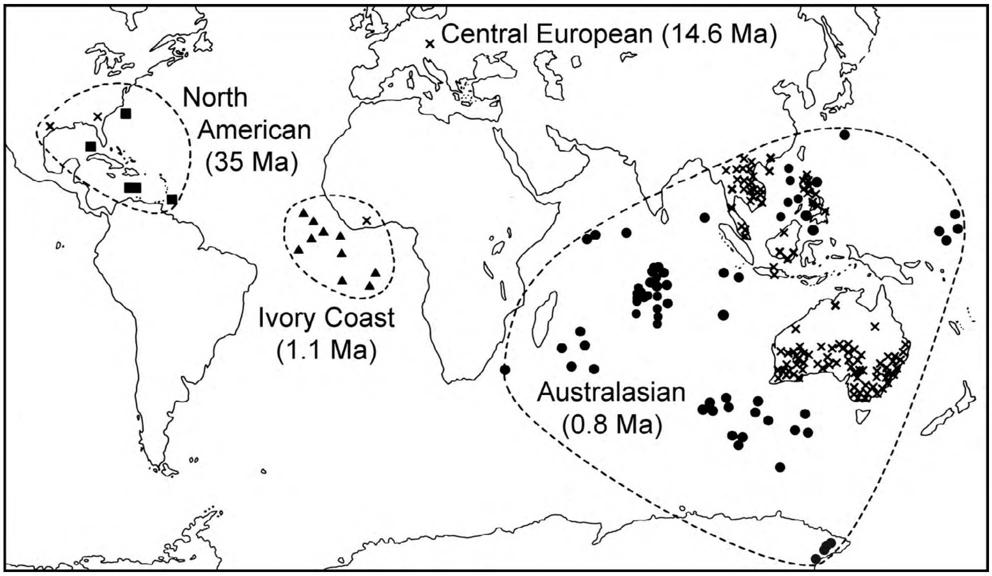


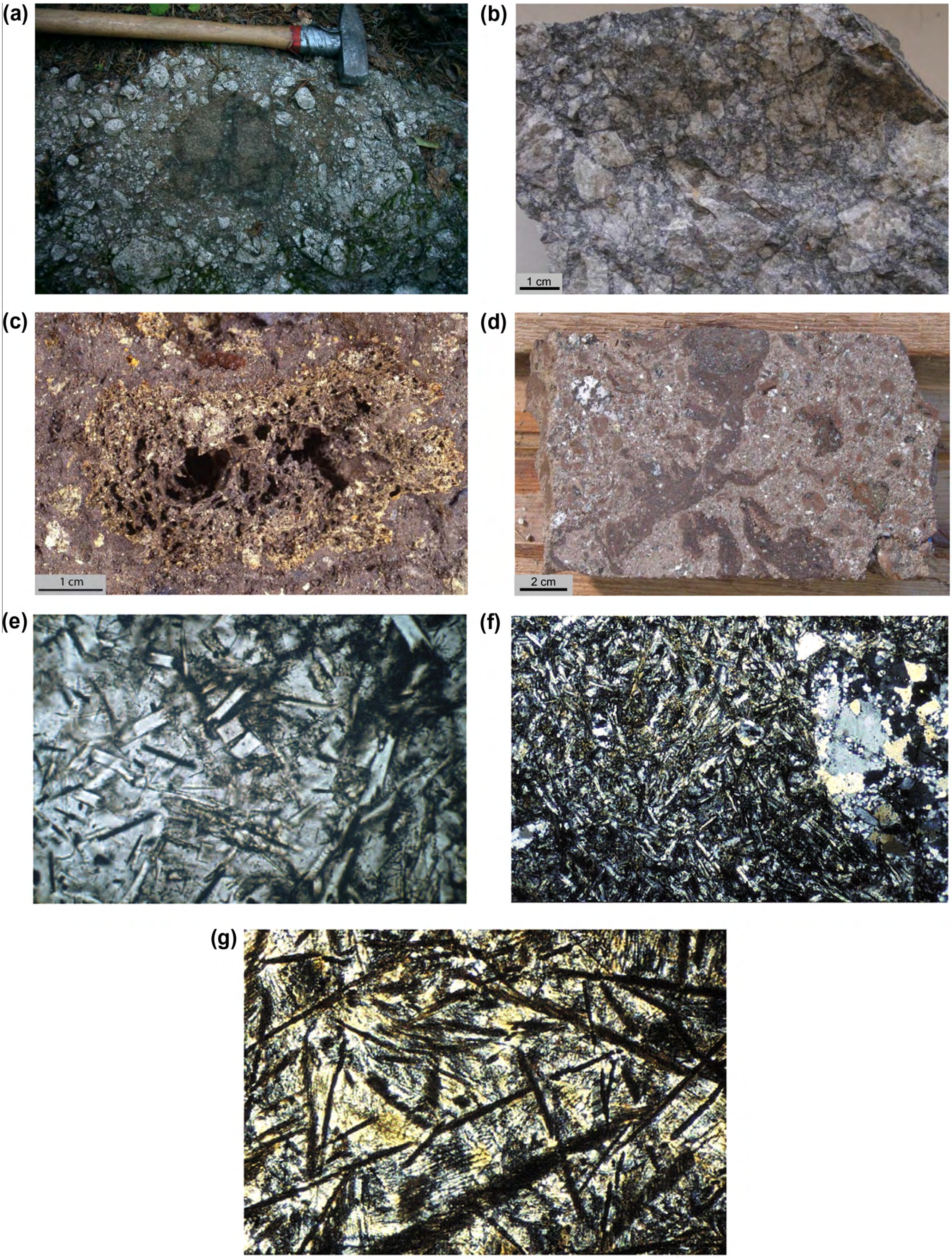
Fig. 14. Geographical distribution of tektites and microtektites in the four known strewn fields.

according to their clast content and matrix composition: *lithic breccias* (in the historical literature also described as "fragmental impact breccia") that are composed of clastic components only (no melt component!); *suevite* is a distinct rock type comprising mineral and lithic clasts of varied shock metamorphic degree, but invariably including a melt component. Both clast content and the relatively finer-grained matrix are composed of individual or variably sutured clasts of melt and lithics. The third major impact breccia type is *impact melt rock* that is composed of a coherent melt matrix, in which there are varied amounts of mineral and lithic clasts. According to the nucleation and cooling history of a melt, the resultant textures can be highly varied – for example, ranging from porphyritic to ophitic or subophitic, and intersertal, inter alia. Considering that impact melt rock may not only resemble volcanic or plutonic rocks with respect to texture but may also have the compositions of basalts or andesites, gabbro or granite, it becomes obvious why in the past so many impact melt rocks have originally been considered products of magmatism. *Impact glass* – in contrast to impact melt rock – is not fully crystallized and either glassy or semihyaline (composed of crystallites in glassy mesostasis).

All these impact breccia types (some examples are shown in Fig. 15) can also occur as local formations or as melt injections in the crater floor, as so-called *dike breccias*. A population of rocks known as dimict breccias has also been described from the Moon. These samples comprise an impactite component and a lithic rock component – the host rock to the breccia injection. Dike breccias also include the enigmatic *pseudotachylitic breccias* (PTB [a term coined by Reimold (1995, 1998)]); for recent discussions of such breccias, see Reimold and Gibson, 2006; Mohr-Westheide et al., 2009) that in the past have been indiscriminately termed "pseudotachylite". The problem is historic, as the first application of this term (historical spelling: pseudotachylyte; Shand, 1916) was used for the enigmatic melt breccias of the Vredefort Dome, i.e., the central part of the world's largest known impact structure. Today, the term "pseudotachylite" is reserved in structural geology for bona fide friction melt only. However, in impact settings pre-, syn-, and post-impact melt breccias of widely different nature and origin can occur: actual friction melt, cataclasite, ultracataclasite that in the field cannot be distinguished from friction melt or impact melt rock, and impact melt rock, and even ultramylonites can resemble the other breccia types closely (an example for multiple types of PTB in the crater floor of an impact structure was reported by Reimold et al. (1999a) from the Morokweng structure of South Africa). Mohr-Westheide and Reimold (2010, 2011) have discussed that a number of genetic processes could be responsible for PTB formation in impact structures, such as genesis under compression and immediate decompression during the early stage of cratering, friction melting of course, and the combination of compression and friction melting, and finally during the modification phase PTB can be produced due to frictional movement on large blocks and also due to rapid decompression. Only a few occurrences of massive pseudotachylite (friction melt rock) are known from tectonic occurrences, in the Italian Alps (Fig. 16a) and in the Musgrave Block of Australia (Camacho et al., 1995).

Besides those at Vredefort, massive PTB occur at the Sudbury impact structure (Canada), where they are also known as Sudbury Breccia. Small manifestations of PTB are known from a host of impact structures. For example, centimeter-wide veinlets of friction

Fig. 15. Microscopic aspects of different impact breccias. (a) Polymict lithic breccia from Lockne impact structure, Sweden. The lithic clast content comprises both granite (light) and dolerite (dark, in center). Hammer for scale some 30 cm long. (b) Monomict lithic breccia (northern rim, Lockne crater). The field of view is about 0.8 m wide. Both images (a) and (b) are of two varieties of so-called Tandsbyn Breccia and were kindly provided by Jens Ormö (Madrid). (c) A ca. 6 cm wide, highly vesicular and partially altered impact melt patch in Bosumtwi suevite. (d) Fourteen cm long drill core of suevite (note the prominent dark melt particles) from 69.22 to 69.36 m depth in borehole Enkingen (SUBO-18) in the southeast of the Ries Crater (Germany), with a significant proportion of dark-colored melt bodies. (e–g) Comparison of microtextures in impact melt rock and pseudotachylitic breccia. (e) Semi-crystalline groundmass of an impact melt breccia from Lappajärvi (Finland), with intersertal texture of isolated plagioclase laths and skeletal pyroxene crystals in abundant mesostasis. Plane polarized light, width of field of view ca. 1 mm. (f) Well crystallized melt groundmass of pseudotachylitic breccia from Leeukop quarry (Vredefort Dome), with feldspar laths and tiny prismatic crystals of amphibole forming the bulk of the groundmass. On texture alone, it would not be possible to classify this sample as a pseudotachylitic breccia – a volcanic/hypabyssal or impact melt origin would be equally possible. The large clast on the right is from a granitoid precursor. It is partially annealed. Cross polarized light, width of field of view ca. 3 mm. (g) A totally different textural expression of pseudotachylitic breccia – this time from a narrow (2 cm wide) veinlet from Lesutaskraal farm (Vredefort Dome). Extremely fine-grained crystals are predominantly plagioclase, and long, dark needles are amphibole. Cross polarized light, field of view ca. 2 mm wide.


(a)
(b)
1 cm
(c)
1 cm
(d)
2 cm
(e)
(f)
(g)

(a)
(b)
(c)
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Fig. 16. Pseudotachylitic breccia occurrences. (a) A meter-scale occurrence of tectonic friction melt rock (pseudotachylite *sensu stricto*) from the Insubrian Line in northern Italy. The breccia occurs in the mafic granulites of Val d'Ossola; the locality is Rio del Ponte near Colloro (Premosello Chiovenda). Image courtesy of Prof. Attilio Boriani, Università degli Studi di Milano. (b) A 3 m wide exposure of pseudotachylitic breccia from Salvamento Quarry, northern core of the Vredefort Dome. PTB with well rounded granitoid clasts occurs with sharp contact (on the left) in the same granitoid host rock. (c) A strand of mm-wide PTB veinlets in granite-gneiss from Salvamento Quarry. The stylized scale bar measures 1 cm wide. (d) A complex occurrence of pseudotachylitic breccia in quartzite of the western collar of the Vredefort Dome, at Steenkampsberg on Thwane Resort. Note the development of thick injection veins off a partially very thin apparent "generation vein" parallel but strongly inclined to the bedding of the quartzite.

melt have been described from Carswell (Canada), e.g., Hilke (1991). Note that this impact structure also exposes other dike breccias such as impact melt rock and clastic-matrix breccias (Hilke, 1991; Harper, 1982). Lambert (1981) reported various dike breccias from Rochechouart (France), where some up to decimeter wide PTB occur, e.g., in the quarry of Champagnac (Reimold et al., 1987). Reimold et al. (1999a) reported locally produced melt breccia from the crystalline basement below the Morokweng impact structure in South Africa.

Significant dike and pod occurrences of melt breccias are known from the granitic core of the central uplift of the Araguainha structure of Brazil (Machado et al., 2009). Contrary to their findings, however, these melt rocks include at least two types of breccias: one with strongly annealed clasts that is thought to represent impact melt rock and another that could be friction melt (Preuss, 2012). The largest manifestation of such melt breccias at this structure observed is at least 5 m wide (compare Fig. 3d). The more than 600 m long drill core BH5-Hättberg from near the center of the core of the Siljan structure revealed two several decameter-wide zones of melt breccia development, with a large number of other centimeter to decimeter wide intersections of melt rock in the other parts of the core as well. Fischer (2013) recently estimated that the combined occurrence of PTB along this drill core amounted to an aggregate of 60 m. Of this, the melt component amounts to some 30% (i.e., 20 m). It has been suggested that formation of this PTB material must have involved friction melting (Kenkmann et al., 2003; Müller, 2013; Fischer, 2013) but it is unlikely that the two large breccia zones are the result of friction melting alone. Instead, involvement of decompression melting is favored by one of us (WUR). Recently, a further significant occurrence of impact-generated PTB and cataclasites has been described from a drill core (MCB-10) into the floor to the Dhala impact structure in India (Pati et al., 2013).

In every case of observation and analysis of such dike breccias, careful investigation in the laboratory, if necessary aided by electron microscopic methods, is required to properly classify PTB according to its matrix type. If this is not possible (for instance, because the groundmass is too far altered), the non-genetic term *pseudotachylitic breccia* ought to be maintained until such time that new conclusive information has become available for final classification. In Fig. 16b, a prominent example of massive pseudotachylitic breccia from the Vredefort Dome (South Africa) is shown.

4.3. Chemical tracers of extraterrestrial projectiles

An important question at any impact structure regards the nature of the impacting body. The detection and verification of an extraterrestrial component in impact-derived melt rocks or breccias can be of diagnostic value to provide confirming evidence for an impact origin of a geological structure (see, e.g., Koeberl, 2007; Koeberl et al., 2012). Generally, a very small amount of meteoritic melt or vapor is mixed with a much larger quantity of target rock vapor and melt, and this mixture is then incorporated into impact melt rocks or melt breccias (= clast-rich melt rock), suevite, or impact glass. In most cases, the amount of extraterrestrial material within these impactite lithologies is much less than one percent by mass. Detecting such small amounts of extraterrestrial matter is very difficult and only elements that have high abundances in meteorites but correspondingly low abundances in terrestrial crustal rocks (e.g., siderophile elements such as Ni and Cr, and the platinum-group elements [PGE]) are used in such studies (cf. Koeberl, 2014; Koeberl et al., 2012). Distinctly higher siderophile element contents in impact melts, compared to target rock abundances, can be indicative of the presence of either a chondritic or an iron meteoritic component (e.g., McDonald et al., 2001).

Complications may arise (a) because meteorites have a range of compositions within each class and some are better constrained than others; (b) if the target rocks have variable siderophile element concentrations; or (c) if siderophile element concentrations in the impactites are very low. Furthermore, the contribution of the target rock (the so-called indigenous component) to the composition of impactites can only be understood if either a well-constrained mixing relationship exists between the impactor and the target rocks that produces a reliable regression line and a lower intercept that reflects the average PGE concentration in the target rocks (e.g., McDonald et al., 2001), or all contributing target rocks have been identified and their relative contributions to the melt mixture are known – something that is very difficult to achieve in practice. This is part of the reason why the nature of the projectiles has not been identified for most of the known impact structures on Earth. Lack of preservation or exposure of melt rocks and breccias suitable for projectile analysis is another reason.

Isotopic compositions can be more specific indicators for the presence of a meteoritic component than elemental concentrations. Of these, the first one to be used with success was the Os isotopic system (see review by Koeberl and Shirey, 1997). Here, the different isotopic compositions and significantly different abundances of the element Os are used to fingerprint a meteoritic component. This system is very sensitive but allows only detecting the presence of a chondritic or iron meteorite component, but without being specific to the meteorite type, and achondritic signatures cannot be identified. In contrast, the Cr isotope system is less sensitive, but offers selectivity between meteorite types, including achondrites (see, e.g., Koeberl, 2007, 2014; Koeberl et al., 2007c,d). Both methods are based on the observation that the isotopic compositions of the elements Os and Cr, respectively, are different between most meteorites and terrestrial rocks and that these differences are sufficiently large to permit detection of relatively small amounts of meteoritic Os or Cr present in the impact rock. The best candidate for detection of a meteoritic material are impactites in the form of melt rocks or breccias, either from a coherent impact melt body or as melt fragments in suevitic breccias, or ejecta such as impact glasses and tektites, all of which can incorporate impactor material (as solid, melt or gas). Caution needs to be applied, though, when comparing geochemical data, as it is important not to view single chemical anomalies in isolation (cf. discussion in French and Koeberl, 2010).

5. The terrestrial impact cratering record

Currently just a few more than 180 confirmed terrestrial impact structures are recorded in the Earth Impact database (http://www.unb.ca/passc/ImpactDatabase). Their locations are shown schematically in Fig. 17. This number increases, as usually a few structures are added each year, but can also go down when previously listed structures (e.g., the paired Arkenu structures, see below) were removed from the record due to subsequent lack of confirmation of original claims. There are other data bases that have collated information regarding a whole lot more structures, most of which have been recognized by more or less circular outlines in remote sensing data sets, including the very popular GoogleEarth software. It must be emphasized once again that only the confirmed structures can be referred as *bona fide* (*confirmed*) impact structures, whereas the others are not recognized as such and should only be referred as *possible* impact structures. There has been a strong hype in the wider community, with a lot of people priding themselves of identifying impact structures – albeit *without* proper verification. A number of such doubtful records have entered even the peer-reviewed literature, which adds to the confusion of non-experts in the field. The rules and criteria for the recognition and confirmation of impact structures, as outlined above, are straightforward; so-called identifications (e.g., the alleged but unproven Chiemgau crater strewn field, southern Germany – Ernstson et al., 2010; the Bajada del Diablo, Argentina, strewn field – Acevedo et al., 2009 – equally lacking bona fide impact evidence; the controversial Maniitsoq structure, Greenland – Garde et al., 2012, 2013) can easily be dismissed as not based on any established criteria (Reimold et al., 2014).

On the other hand, readily available remote sensing data including GoogleEarth can be very useful for obtaining initial hints at the existence of previously unknown, potentially interesting

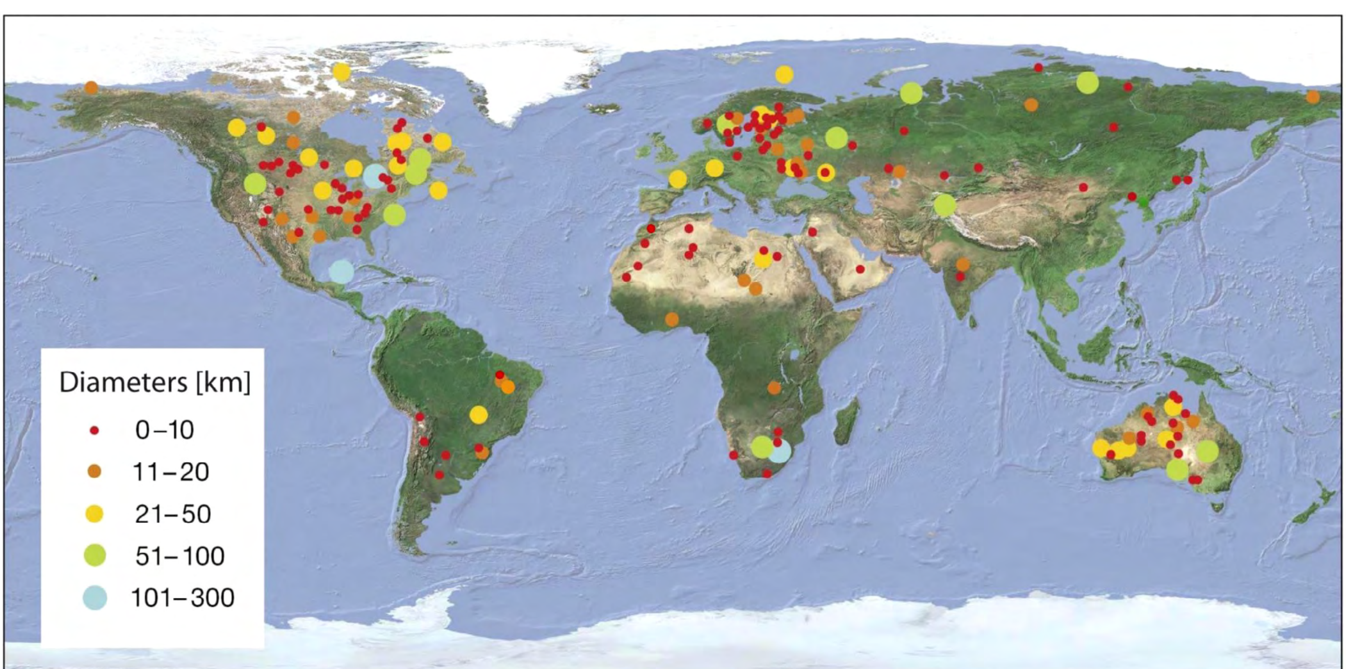


Fig. 17. *Global* distribution of ca. 180 confirmed impact structures, based on a diagram of Reimold and Jourdan (2012). Also compare the Earth Impact Database of confirmed impact structures (http://www.unb.ca/passc/ImpactDatabase).

structures. Such observations should be followed up with ground-truthing, and where the relevant expertise does not exist, interaction with established impact cratering research groups ought to be sought.

The global map of locations of confirmed impact structures (Fig. 17) shows obvious regional variations in crater density. Most known impact structures are located in North America and Eurasia, and there is another cluster in Australia. An obvious reason for the former is that there is a definite link between space exploration and related widespread knowledge about planetary impact structures. Consequently in those countries where active space exploration has been pursued, and their allied nations, impact structures have been searched for and investigated ever since the 1950s. Other small, active groups with a strong interest in impact cratering have made regional progress – in Scandinavia, in southern Africa, and in Australia. European efforts have been strongly supported by continental research foci in impact cratering, such as a European Science Network program in the 1990s and the Nordic Countries' Impact Program of the last decade. Both these programs have been highly successful in educating many geoscientists and, especially, students about this important planetary process. The large number of impact structures identified in Australia must be credited to the single effort of the late Gene Shoemaker and his wife Carolyn, in collaboration with a number of Australian colleagues. In South Africa, the Impact Cratering Research Group at the University of the Witwatersrand has been instrumental in the geo-analysis of "Vredefort and Company". Recent enhanced efforts in South America should be noted as well, especially the detailed multidisciplinary work of the universities of Campinas and Sao Paulo.

Overall, however, one should not forget that differential preservation due to the varied geological histories of individual regions has been a major underlying factor determining the spatial distribution of impact structures. Young oceanic crust, recently exhumed continental crust, and – to the contrary – long exhumed, stable continental platforms should be expected to have very different crater accumulation and obliteration rates.

Nevertheless, there are several regions in the world that do not have their share of the terrestrial impact record at the same level as those regions mentioned above. These are large parts of Asia, especially the eastern regions of Siberia, Mongolia, and China where so far only a few relatively small impact structures are listed, the Middle East, and Southeast Asia. Furthermore, the rain forest regions of the equatorial belt have not lent themselves to detailed crater exploration, as well as the hostile environments of Antarctica and Greenland, with large parts being covered by kilometer-thick ice.

The terrestrial cratering record is clearly incomplete. Just by how much is a complex question to tackle. Stewart (2011) recently estimated that 228 impact craters >2.5 km remained to be identified in the global Phanerozoic (younger than 540 Ma) sediment record alone. Her estimates for Africa indicate that of the full crater population between <1 and >10 km in diameter, only about 15%

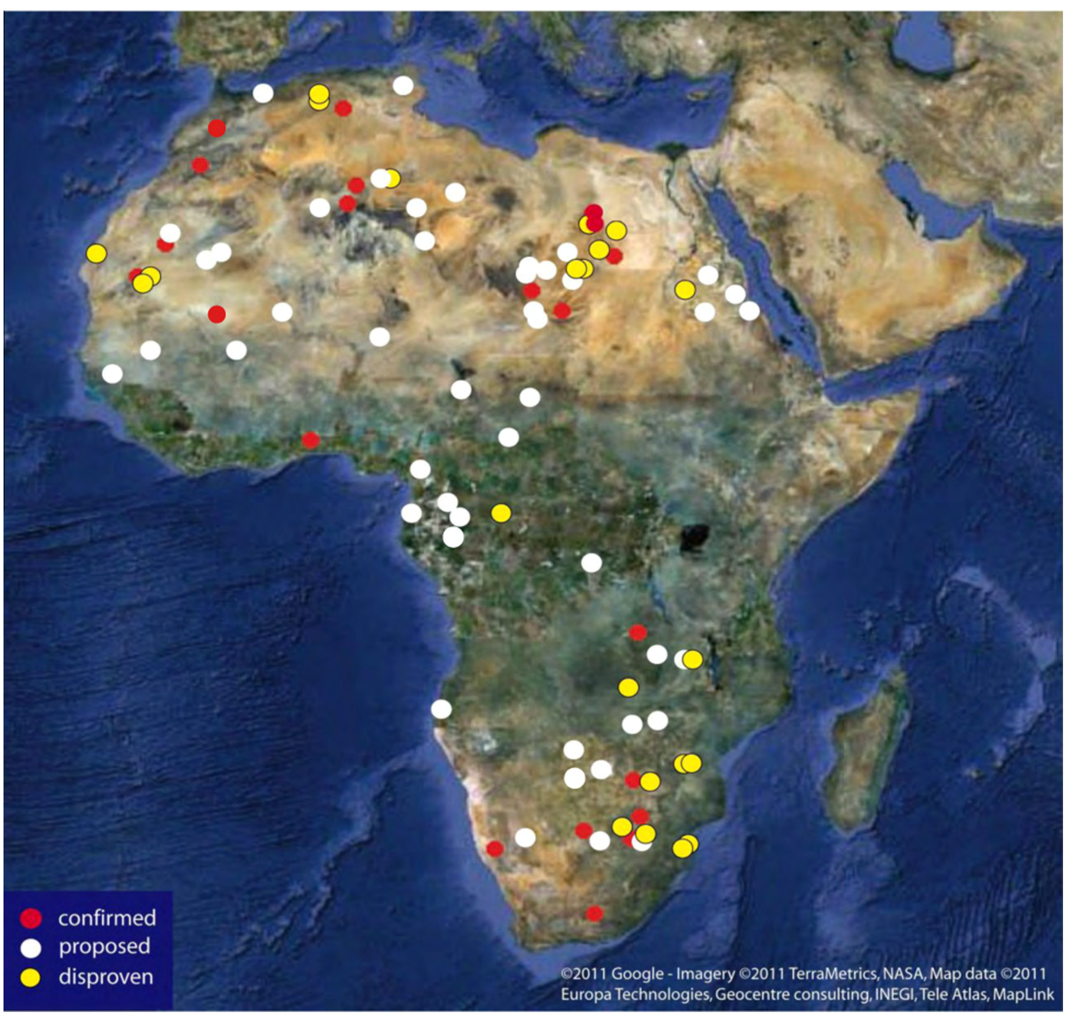


Fig. 18. The record of confirmed and possible impact structures in Africa. Also shown are locations of those structures that have already been disproven (several sites in Morocco have not been plotted in order to avoid clutter). Detail about all these structures is summarized in Tables 2a–c. Background image used with permission from the Earth Impact Database, University of New Brunswick.

had been discovered, whereas for structures <210 km diameter, less than 10% had been accounted for. For craters >10 km in diameter, about 45% (i.e., 5 out of a total of 11) are known. It is obvious that such a computation is not an easy task and that it would depend on a number of generalizations and assumptions. For example, how much of the Phanerozoic sediment record has actually survived plate tectonics and surface degradation, on average? Has the impact flux since the end of the Precambrian remained constant or were there anomalies? And can the global rate of degradation be averaged? How would differential rates of uplift go into the equations? And how can we estimate the differential rates of obliteration particularly for very small (<2 km size) craters? Apart from the record of impact in Phanerozoic sedimentary basins, how much evidence of impact remains to be discovered in the extensive African regions of Precambrian crust (estimated at some 15–20% by area of the entire African continent)? In conclusion, however, it is safe to expect that more impact structures will be discovered in Africa in time to come.

6. The impact record of Africa

Since the time when the last review of African impact structures was published (Koeberl, 1994) and merely 14 confirmed impact structures were listed, plus the Highbury structure in Zimbabwe that, even to date, has not been confirmed (see below), the African cratering record has now increased to 19 confirmed impact structures plus the Agoudal shatter cone occurrence. In addition, there are a considerable number of proposed structures that still require confirmation. The African record is shown in Fig. 18 and summarized in Tables 2a–c. In addition, there have been some alleged discoveries that need to be discussed as already disproven cases. Several impact glass and distal ejecta layers have been added to the record since 1994 as well. Quite a few of the structures already reported as known impact structures in 1994 have been subject of further, partially spectacular, research. Established groups interested in impact research continue to ply their trade in Africa, and newcomers, including a number of young scientists, have joined their efforts. Overall, the extensively expanded record of African impact structures tells a real story of progress.

In the following the current "state of affairs" (Tables 2a–c) is discussed structure by structure, starting with the confirmed impact crater structures and impact formations and then progressing to still uncertain candidates, and eventually listing the already disproven cases.

6.1. Confirmed impact structures of Africa

6.1.1. Agoudal, Morocco

An apparent remnant of an impact structure, with shatter-cone-bearing marl, has recently been indicated by Sadilenko et al. (2013)

Table 2a
Summary of confirmed African impact structures (Lat. = latitude, Long. = longitude; NA = not available).

Crater name	Country	Lat.	Long.	Diameter (km)	Age (Ma)	Exposed	Drilled	Target rock	Identification why	Bolide type
Agoudal	Morocco	31°59′12.7″N	5°30′57.3W	?	<170	(Y)	N	Marl	Shatter cones	–
Aorounga	Chad	19°06′N	19°15′E	16	<345	Y	N	Sedimentary rocks	PDF, shatter cones	–
Aouelloul	Mauretania	20°15′N	12°41′W	0.39	3.1 ± 0.3	Y	N	Sandstone	Lechatelierite, baddeleyite, Ni-rich Fe-spherules, siderophile elements, Re-Os isotopes	Iron (IIIB/IIID?)
Bosumtwi	Ghana	06°30′N	1°25′W	10.5	1.07	Y	N	Metasediments + granite intrusions	Shock metamorphism, incl. PDF	Chondrite? Iron?
BP	Lybia	25°19′N	24°20′E	2	<120	Y	N	Sandstone, siltstone	PDF	–
Gweni Fada	Chad	17°25′N	21°45′E	22–23	<345	Y	N	Sandstone, limestone	PDF	–
Kalkkop	South Africa	32°43′S	24°34′E	0.64	0.25 ± 0.05	Y	Y	Sandstone, shale	PDF, Re-Os isotopic analysis	–
Kamil	Egypt	22°01′N	26°05′E	0.045	NA	Y	N	Sandstone	Meteorite find	Iron
Kgagodi Basin	Botswana	22°29′S	27°35′E	3.4	<180	Y	Y	Granitoid basement + minor Karoo dolerite	PDF, diaplectic quartz glass, high siderophile elements	–
Luizi	DRC	10°10′S	28°00′E	17	<573	Y	N	Sandstone	Shatter cones, PDF	–
Morokweng	South Africa	26°28′S	23°32′E	75	145 ± 2	Barely	Y	Metasediments, volcanics and granitoids	PDF, high siderophile elements, Re-Os isotopes, meteorite relic	L chondritic
Oasis	Libya	24°35′N	24°24′E	25–36	<120	Y	N	Sandstone, siltstone	PDF + glass pockets	–
Ouarkziz	Algeria	29°00′N	07°33′W	3.5	65-345	Y	N	Limestone, shale	PDF	–
Roter Kamm	Namibia	27°46′S	16°18′E	2.5	<5	Y	N	Metasediments + granitic gneiss	PDF, siderophile elements, PGE systematics	chondritic
Talemzane	Algeria	33°19′N	04°02′E	1.75	<3	Y	N	Limestone	PDF	–
Tenoumer	Mauritania	22°55′30″N	10°24′W	1.9	<3	Y	N	Granitic–amphibolitic basement + thin veneer of sediment	PDF, lechatelierite, diaplectic quartz glass, ballen quartz	–
Tin Bider	Algeria	27°36′N	05°07′E	6	70	Y	N	Clay, limestone, sandstone	PDF	–
Tswaing (Pretoria Saltpan)	South Africa	25°24′S	28°05′E	1.13	0.22 ± 0.052	Y	Y	Granite overlain by Karoo grit, minor diatomite	PDF, diaplectic quartz glass, siderophile element enrichment, Re-Os isotopes	Chondrite
Vredefort	South Africa	27°00′S	27°30′E	⩾250	2023 ± 4	Y	Y	Granitic basement overlain by varied supracrustals	Shatter cones, coesite, stishovite, PDF, shocked zircon + monazite, siderophile elements, Re-Os, PGE systematics	Chondrite

? – indicates that the size of the original impact structure is not known; likely only a small portion of the structure, with shatter cone-bearing marly shale is preserved (see below, section 6.1.1).

Table 2b
Summary of proposed and not yet confirmed impact structures in Africa.

Crater name	Country	Latitude	Longitude	Diameter (km)	References
Anefis	Mali	18°N	0.5°W	3	Rossi (2002)
Azenak structures	Niger	16.5°N	8°E	0.5 + 1	Rossi (2002)
Bangui	Central Africa	04°22′N	18°33′E	~600	Girdler et al. (1992)
Bateke Plateau	Gabon	0°38′45″S	14°27′29″E	~7.1	Master et al. (2013)
Bangweulu Basin	Zambia	11°10′S	29°54′E	150	Master (1993)
Chituli Structure	Zambia	11°15′33″S	32°24′00″E	3.8	Master (2001)
El Mrayer	Mauritania	22.5°N	7.2°W	2	Rossi (2002)
Faya Basin	Chad	18°11′N	19°33′E	2	Schmieder and Buchner (2010)
Foum Teguentour	Algeria	26°14′30″N	02°25′E	8	Lambert et al. (1981)
Gogui	Mauretania	15°30′06″N	11°23′50″W	0.5–0.6	Rossi (2002)
Highbury	Zimbabwe	17°05′S	30°09′E	15–25	Jacobsen (1962), Master et al. (1994)
Ibn-Batutah	Libya	21°34′10″N	20°50′15″E	2.5	Ghoneim (2009)
In Ezzane (5 structures)	Algeria	23°29′N	11°14′30″E	4–9	Bonin et al. (2011)
Jaraminah	Libya	26°32′17″N	10°35′28″W	2.2	Dunford and Koeberl (2009)
Jebel Hadid	Libya	20°52′12.43″N	22°42′17.73″E	4.7	Schmieder et al. (2009)
Jwaneng South	Botswana	24°42′S	24°46′E	1.3	Master et al. (2009), Master (2010a,b))
Karas	Namibia	26°S	19°E	300	Corner (2008), Miller, 2008a,b
Kogo	Equatorial Guinea	1°11′N	10°1′E	4.7	Martinez-Torres (1995)
Lac Iro	Chad	10°10′N	19°40′E	13	Garvin (1986)
Mékambo	Gabon	0°55′39″N	13°40′25″E	50	Antoine et al. (2000)
Minkébé	Gabon	1°21′15″N	12°24′29″E	90	Antoine et al. (2000)
Mora Ring	Cameroon	11°N	14°E	7	Garvin (1986), Nickles (1952)
Mousso	Chad	17°58′N	19°53′E	3.8	Buchner and Schmieder (2007)
Ntwetwe	Botswana	20°55′S	24°50′E	7	Master (1993)
Okavango Delta (Khurunxaraga crater)	Botswana	19°50′S	23°20′E	0.022	Henshaw (1997)
Omeonga (Katako-Kombe)	DRC	3°35′11″S	24°29′10″E	38–45	Ferrière (2011), Ferrière et al. (2012)
Oun	Chad	21°44′N	19°20′E	8	González and Alonso (2006)
Ouro Ndia	Mali	14°59′45″N	4°30′00″W	3	Rossi (2002)
Ras Zeidun	Egypt	NA	NA	7	Barakat (2011)
Reitz structure	South Africa	26°30′S	27°59′E	150–500	Antoine and Andreoli (1995)
Rwanda crater	Rwanda	NA	NA	NA	Denaeyer and Gérards (1973)
Saghira	Libya	NA	NA	2.7	El-Baz and Ghoneim (2007)
Setlagole	South Africa	26°22′30″S	25°07′30″E	25–30	Anhaeusser et al. (2008, 2010)
Sinamwenda	Zimbabwe	17°11′42″S	27°47′30″E	0.2	Master et al. (1995)
Temimichat Ghallaman	Mauritania	24°15′N	09°39′W	0.7	Rossi et al. (2003)
Terhazza	Mali	23°N	6°W	1	Rossi (2002)
Tigraou	Algeria	35°02′32″N	01°54′12″W	1	Chabou (2011)
Tongo	Cameroon	4°51′–5°00′N	10°21′-10°31′E	10 × 12.5	Tenemou (2010)
Tmisan	Libya	27°25′24″N	13°24′33″E	3.2	Dunford (2008), Dunford and Koeberl (2009)
Unnamed	Botswana	19°07′40.0″S	23°18′12.7″E	15–20	Cooper et al. (2010)
Unnamed	Libya	NA	NA	2	Ben Musa and Baegi (2009)
Unnamed	Sudan	17°57′N	37°55′E	5.5–6	Di Achille (2004)
Unnamed	Sudan	19°12′47″N	35°59′00″E	3	Sparavigna (2010a)
Unnamed	Sudan	21°17′25″N	33°59′00″E	6–7	Sparavigna (2010a)
Unnamed	Sudan	18°03′25.52″N	33°30.22′E	10	McNally (2010), Sparavigna (2010b)
Unnamed	Angola	15°12′07″S	12°45′08″E	1.1 × 0.9	Roger Swart (Windhoek, pers. commun.)
Unnamed (Jebel al Bukrah)	Tunisia	35°45′N	9°8′E	5	Tomlinson (1999), Youbi et al. (2011)
Uri	Chad	21°17′N	19°20′E	5	González and Alonso (2006)
Velingara	Senegal	13°02′13.2″N	14°7′40″W	48	Master et al. (1999), Wade et al. (2002)

NA = not available.

from the vicinity of the small town of Agoudal. The location has also given its name to the Agoudal iron meteorite strewn field (Chennaoui Aoudjahane et al., 2013). The discovery site of shatter cones (two examples of occurrences are shown in Fig. 54e and f in section 6.5.4) is located at 31°59′12.7″N/5°30′57.3″W against a hill slope. It has been surmised that these shatter cones, which are clear evidence of shock deformation, could be related to the Agoudal meteorite fall, and along this line of thought, N. Artemieva (pers. comm., 2013) reports that an impactor of initial (pre-atmospheric) size of 1–2 m and a mass compatible with the density of the Agoudal meteorite type (IIAB iron) could have generated enough energy to produce shatter cones. The iron meteorite strewn field was created by small fragments, whereas the largest fragment of 30–50 cm size could have reached – in such a scenario – the surface with a velocity of 0.5–1 km/s and generated local shock pressure of 1–2 GPa sufficient to produce shatter cones. These ideas would then be consistent with known fragmentation models of iron meteorites. Nevertheless, such ideas do not agree with other known occurrences of shatter cones that so far have only ever been found in larger impact structures.

Despite the scarce information published to date about this alleged impact site, we decided to place Agoudal into the list of confirmed impact structures. Indeed, when we visited the site in October 2013, we found that the occurrence of shatter cones is much more widespread than initially indicated. Over an area of several hundred square meters of outcrop and suboutcrop marly limestone with abundant shatter coning was mapped, and in an even wider area of at least 1000 × 1000 m extent up to several centimeter sized fragments of shatter coned marl were found on surface and within the upper talus blanket. It is, however, reasonable to assume that some of this extensive distribution is the result of surficial erosion (run-off during the sometimes torrential downpours of the rainy seasons).

The detailed results of our field work will be presented elsewhere (Chennaoui Aoudjehane et al., 2014). At this stage, we can conclude, however, that a first impact event in the territory

Table 2c
Summary of already discarded structures that had been proposed as impact structures. (NA = not available).

Crater Name	Location	Latitude	Longitude	Diameter (km)	References
Aflou	Algeria	34°00′N	02°03′E	3 × 5	Lambert et al. (1980)
Al Mouilah	Algeria	33°51′N	02°03′E	4.5	Lambert et al. (1980)
Arkenu 1	Libya	22.1°N	23.8°E	10.3	Paillou et al. (2003), Di Martino et al. (2008)
Arkenu 2	Libya	22.05°N	23.72°E	6.8	Paillou et al. (2003), Di Martino et al. (2008)
Bir Anzarane	Morocco	23°3′13.59″N	15°23′35.33″W	1.5	Chaabout et al. (2011)
Bushveld Complex	South Africa	24–26°S	26–31°E	500 × 350	Buchanan and Reimold (2002)
Chegutu	Zimbabwe	18°08′13″S	30°08′09″E	NA	Reimold (1994)
Delmas Sinkhole	South Africa	26°09′S	28°46′37″E	~0.08	R. Meyer, Pretoria (pers. commun.)
El Baz	Egypt	24.2°N	26.3°E	4	El-Baz (1981), El-Baz and Issawi (1982), Orti et al. (2008)
Gilf Kebir Crater Field	Egypt	23°14′N–23°32′N	23°17′E–27°27′E	NA	Paillou et al. (2006), Orti et al. (2008)
Kebira	Lybia	24.40°N	24.58°E	31	El-Baz and Ghoneim (2007)
Kwa-Zulu-Natal 1	South Africa	26°56′36″S	32°47′00″E	0.065 × 0.170	Brandt et al. (1999, 2001)
Kwa-Zulu-Natal 2	South Africa	26°53′00″S	32°51′20″E	0.050–0.060	Brandt et al. (1999, 2001)
Lac Télé	DRC	1°20′N	17°10′E	6 × 8	Garvin (1986), Master (2010a,b))
Lukanga Swamp	Zambia	14°24′S	27°54′E	52	Vrána (1985), Katongo et al. (2002)
Mazoula	Algeria	28°24′N	7°49′E	0.8	Lambert et al. (1981)
Nyika Plateau Structure	Malawi	~9°30′S–14°12′S	~32°15′E–34°00′E	0.08	Master and Duane (1998)
Richat	Mauritania	21°07′N	11°23′W	38	Matton et al. (2005)
Save East	Zimbabwe	20°00′09″S	32°21′27″E	0.6	Reimold (1994), Master and Robertson (2009)
Save West	Zimbabwe	19°55′11″S	32°15′59″E	0.8	Reimold (1994), Master and Robertson (2009)
Semsiyat	Mauritania	21°00′55″N	11°50′02″W	5	Dietz et al. (1969)
Thuli	Zimbabwe	21°55′S	29°12′E	1.1	Reimold (1994), Master and Robertson (2009)

of Morocco is indicated by considerably widespread occurrence of shatter cones. An upper age for this impact is only constrained by the mid-Jurassic age of the local country rocks. A link to the Agoudal iron meteorite fall is purely hypothetical and – based on our observations – not supported by field evidence. It appears that the meteorite fall occurred over an already impact-affected terrane.

6.1.2. Amguid, Algeria

This 450 m diameter crater structure (Fig. 19) was first recognized at 26°05′N/04°23′E from an airplane in 1954. Geological information was contributed by Lambert et al. (1980). The crater is exposed in Lower Devonian sandstones. It is characterized by an up to 50 m high elevated rim and filled with coarse-grained alluvial and eolian deposits overlain by fine-grained eolian silts. The sandstone beds of the crater rim have dips that become progressively steeper in the upper part of the rim wall; in the NNW and SSE parts of the rim overturned beds have been described. Lambert et al. (1980) contributed a schematic geological map and cross section (reproduced in Koeberl, 1994) that summarize the geology at the crater. The relatively young age of this structure, estimated by Lambert et al. (1980) at less than 0.1 Ma, is based on the excellent preservation state of the crater structure that includes a continuous ejecta blanket outside of and up to 100 m from the rim. Lambert et al. (1980) and Lambert and Lamali (2009) reported limited evidence of impact: deformation in quartz from the crater wall is restricted to irregular fractures and undulatory extinction. Less than 1% of the quartz grains studied displayed mostly non-decorated PDFs, with one or two orientations per grain.

Fig. 19. Amguid impact crater in Algeria, about 500–530 m in diameter (NASA image).

6.1.3. Aorounga, Chad

The deeply eroded Aorounga structure (Fig. 20) is located at 19°06′N/19°15′E in northern Chad, about 110 km southeast of the prominent Emi Koussi volcano in the Tibesti Massif. According to the older literature, the structure is about 13 km wide. Already from remote sensing imagery it is obvious that the structure must be deeply eroded. The entire region is transected by strongly parallel longitudinal dunes/ridges (of likely eolian origin) that transect the crater structure as well. Aorounga has first been studied using photogeology (Gemini, Apollo, Landsat imagery, and aerial photography) by Roland (1976). This suggested an origin either as a granite diapir or as an impact crater, with Roland favoring the former hypothesis. Grieve et al. (1988) suggested that Aorounga could possibly be of impact origin. In the course of a French expedition

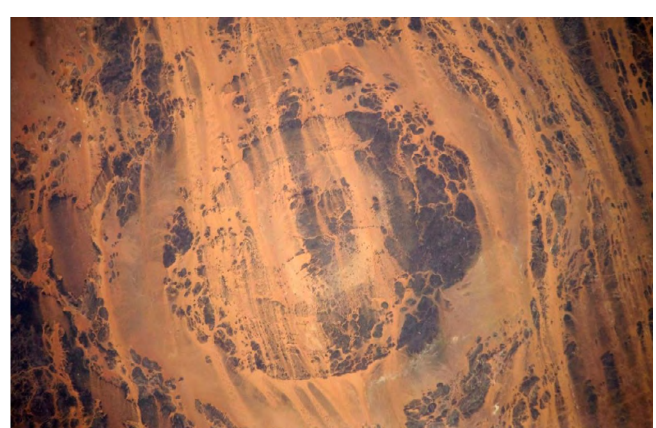


Fig. 20. The deeply eroded Aorounga impact structure in Chad, central Africa, with an original diameter of about 18 km (NASA ISS image).

a few sandstone samples were recovered from the structure; Becq-Giraudon et al. (1992) reported evidence of multiple sets of planar deformation features and field observations of shatter cones as definite evidence for an impact origin. Koeberl (1994) observed that the pictorial evidence given by Becq-Giraudon et al. (1992) was "not necessarily convincing". He also observed that the alleged shatter cones could well be wind-ablation features that are well-known from desert terranes (e.g., features observed at Roter Kamm impact crater – Reimold and Miller; see Fig. 13c,d).

The structure is hosted by fine-grained, well-sorted, sandstone that has a minor carbonate component. Its age is thought to be Upper Devonian (Wacrenier et al., 1958). According to Becq-Giraudon et al. (1992), the Aorounga structure comprises an outer and an inner ring wall at 7 and 11 km from the center, both of which rise by some 100 m above the mean level of the environs. The ring structures are separated by an annular depression of uniform width. The innermost area represents a hilly terrain of ca. 1.5 km width. The ring walls consist of moderately (40–50°) outward dipping strata; dips in the inner central uplift are significantly steeper (80°). Becq-Giraudon et al. reported the finding of a breccia consisting of fine-grained clasts with a fluidal texture on top of the inner ring wall. The age of the structure was estimated by these authors based on ^{14}C dating at 3,500–12,000 years, which appears rather too young in the light of the eroded appearance of the structure. No melt-bearing samples that might have allowed application of more promising methodology (K–Ar chronology, or U–Pb zircon dating) have been found to date.

Cosmogenic nuclide exposure age dating was attempted by Bourles et al. (1995). Miallier et al. (1997) tried thermoluminescence and electron spin resonance dating of quartz from Aorounga samples, and obtained 0.5–1 Ma results that in all likelihood only represent minimum age values for this impact event. This leaves the Upper Devonian age of the target rock – if it is correct – as the only currently available maximum age limit for the impact event.

Koeberl et al. (2005b) reinvestigated the Aorounga area with remote sensing data and contributed new petrographic and chemical results. They concluded that the Aorounga structure comprises a 9 km wide, structurally complex central zone. This, in turn, is surrounded by a 2–3 km wide ring structure of low topography, and then a further, outer ring of 3.5 km width, which they think could represent the crater rim. This would combine to a revised crater diameter of 16 km. Koeberl et al. (2005b) also carried out optical microscopic studies on a number of target rock samples (quartzite and sandstone), some of which they considered as monomict breccias. They found one sample with quartz grains with up to 5 sets of PDFs, and several others with abundant grains with planar fractures and rare grains with single sets of PDFs. These results support the conclusions of Becq-Giraudon and unambiguously assign an impact origin to the Aorounga structure. The chemical and isotopic results of Koeberl et al. (2005b) for Aorounga samples are typical for upper crustal silica-rich sedimentary rocks.

There is much scope for further detailed ground analysis of this impact structure. However, Chad and adjacent territories have been subject to intense civil strife for the past decades, and solutions to these conflicts do not appear to be close. Ground-truthing of newly proposed possible impact structures and more detailed fieldwork on already confirmed sites is therefore rated highly hazardous at this time.

6.1.4. Aouelloul, Mauretania

Aouelloul impact crater (20°15′N/12°41′W; Fig. 21) is – at 390 m rim-to-rim diameter – a small, simple bowl-shaped crater. The structure is located in the Adrar region of Mauritania. It was discovered from the air by A. Pourquié in 1938 and first visited on the ground in 1950 (Monod and Pourquié, 1951). Target rocks are mainly Zli sandstone, besides subordinate Oujeft sandstone, both of Ordovician age. The well-developed rim rises some 15–25 m above the surrounding desert and about 53 m above the crater floor. It shows a distinctly overturned sequence of strata, which is also in support of very limited crater degradation. The crater is filled with poorly-sorted, sandy silt that is overlain by well-sorted eolian sand. A gravity study by Fudali and Cassidy (1972) indicated a maximum thickness of the sedimentary fill of some 23 m, which was underlain by a breccia lens to crater floor at 130 m depth. Grieve et al. (1989) suggested a slightly different gravity model, also with a ca. 100-m-thick breccia lens.

Fig. 21. The 0.4-km diameter Aouelloul impact structure in Mauritania, West Africa (NASA image).

Koeberl et al. (1998a) added some field observations from a 1989 expedition to Aouelloul. The crater rim is covered by eolian deposits that locally expose loose blocks of sandstone. If these blocks represent fallout from the cratering event, this crater has not been significantly eroded and consequently must be quite young. Where proper rim outcrops were studied, strata were found to be upturned, dipping at 35–60° towards the crater center. Rarely on the western slopes were overturned strata (possibly local, rotated blocks) observed. By far most glass fragments were collected on the southern, southeastern and northern outer flanks of the crater rim, and only one tenth of the collection was retrieved on the southeastern inner flank. With this collection, the glass distribution initially recorded by Monod and Pourquié was greatly expanded.

Glass fragments found at the crater were investigated by Campbell-Smith and Hey (1952), who reported that their composition is that of silica glass similar to Darwin and Wabar glasses but concluded that the Aouelloul glass was a mixture of local sandstone and remnants of the meteoritic projectile. Chao et al. (1966a) interpreted Ni-rich Fe spherules with 1.7–9 wt% Ni as remnants of the projectile. The glass is heterogeneous with a well-developed schlieren texture. Individual schlieren have very different chemical compositions. There are also digested quartz and feldspar grains. Chao et al. (1966b) and Koeberl and Auer (1991) found that the glass composition resembled that of Zli sandstone (also Cressy et al., 1972) with some enrichment of siderophile elements. Morgan et al. (1975) concluded on the basis of siderophile interelement ratios that the projectile could have been of pallasite or iron meteorite (group IIIB or IIID) composition. Koeberl et al. (1998a) judged this classification as equivocal, as at the time the ratios for the target rocks were unknown. El Goresy (1965) and El Goresy et al. (1968) showed that the glass contains the high-temperature phases lechatelierite and baddeleyite. Beran and Koeberl (1997) determined a very low water content of the glass, which is further evidence for high-temperature formation. Combined, these analytical observations have been taken as conclusive evidence for an impact origin of this small crater structure. K–Ar and fission track dating of impact glass by Storzer and Wagner (1977) and Fudali and Cressy (1976) yielded an age of 3.1 ± 0.3 Ma for the Aouelloul crater.

Earlier work did not report definite evidence for impact – only healed planar fluid inclusion trails that were interpreted as possible remnants after PDFs. Koeberl et al. (1998a) reported widespread fracturing and shattering of quartz grains in sandstone samples, with subplanar and planar fractures being abundant. In addition, they also described relatively wide and irregularly spaced fluid inclusion trails, with several grains showing two sets of relatively straight and closely spaced trails. However, these authors concluded that this possible evidence of impact was ambiguous: the fact that the fluid inclusion trails had been alleged to be "healed" implied that they were in all likelihood pre-impact features related to the metamorphic history of the target rocks. Koeberl et al. (1998a) discussed in detail the difficulty of finding definite evidence of impact deformation especially in small impact structures formed in sedimentary rocks. The smaller an impact structure, the less energetic the event was, which implies a very limited zone of shock deformation in excess of the Hugoniot Elastic Limits of the major rock-forming minerals. In the case of quartz, that would be a threshold of 8–10 GPa at maximum (Huffman and Reimold, 1996), the onset shock pressure for formation of shock-diagnostic planar deformation features in this mineral.

In the light of petrographic analysis having failed to provide definite evidence for impact in the form of optically identifiable shock-metamorphic indicators – for this very small crater structure, and as the results of previous chemical investigations remained ambiguous, Koeberl et al. (1998a) resorted to a Re–Os isotopic investigation of glass and target rocks. Their results ascertained that the glass contained an unambiguous meteoritic component, confirming the impact origin of Aouelloul crater.

6.1.5. Bosumtwi, Ghana

The Bosumtwi crater (Fig. 22) in Ghana, West Africa, is centered at 06°30′N and 01°25′W. The 1.07 Ma old impact structure (Koeberl et al., 1997a) is situated in the Ashanti Region, about 32 km east of Kumasi, the regional capital. The Bosumtwi impact structure is arguably the youngest and best preserved terrestrial impact structure larger than 6 km in diameter (e.g., Scholz et al., 2002; Koeberl and Reimold, 2005). The crater has a pronounced rim, with a rim-to-rim diameter of about 10.5 km (Fig. 23a). The structure is somewhat asymmetrical, with the southern margin clearly having been affected by the superposition of the crater excavation onto the northern flank of the prominent, NE–SW trending Obuom Mountain Range. The structure forms a hydrologically closed basin (Turner et al., 1996a,b) and is almost completely filled by the 8.5-km-diameter Lake Bosumtwi, which made it an attractive target for a large drilling project in 2004 (Fig. 23b). The lake has a maximum depth of about 80 m and the crater rim rises about 250–300 m above the lake level. The area forms part of a tropical forest environment with warm climate, high rainfall, and high organic activity. Chemical weathering is intense, leading to the formation of locally thick lateritic soils.

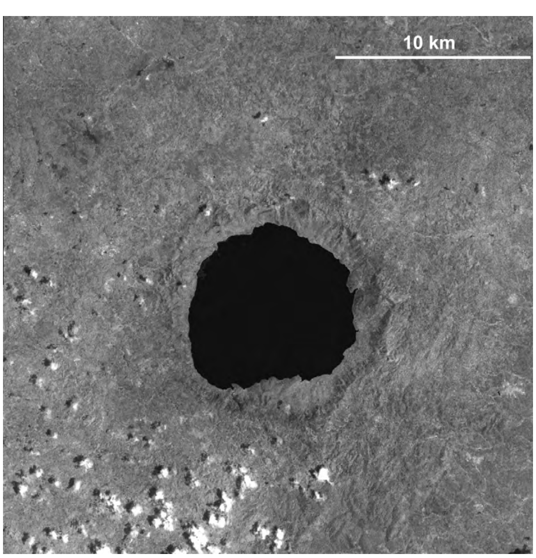


Fig. 22. Satellite view of the 1-million-year-old Bosumtwi impact structure, Ghana, West Africa. The 10.5-km-diameter impact structure is partly filled by an 8.5-km-diameter lake.

Studies over the past 50 years have confirmed that the Bosumtwi crater structure was formed by meteorite impact. This is indicated by outcrops of suevitic breccia around the crater (Chao, 1968; Jones et al., 1981), samples of which have been shown to contain the high-pressure silica polymorph coesite (Littler et al., 1961), as well as Ni-rich iron spherules and baddeleyite in vesicular glass (El Goresy, 1966; El Goresy et al., 1968). In addition, Koeberl et al. (1998b) described shock-characteristic planar deformation features (PDF) in quartz from suevitic breccia (see also Boamah and Koeberl, 2006).

The Bosumtwi impact structure is also the likely source crater for the Ivory Coast tektites (e.g., Gentner et al., 1964; Jones, 1985; Koeberl et al., 1997b, 1998b). This correlation is based on similarities in geochemical and isotopic composition of target rocks and tektites, as well as similarities in the ages of impact melt from Bosumtwi suevite and of the Ivory Coast tektites. Boamah and Koeberl (2003) carried out detailed petrographic and geochemical studies on suevites from shallow drill cores obtained to the north of the crater. Results of structural and geological mapping of the Bosumtwi crater were reported by Reimold et al. (1998a), which included a structural profile through the crater rim. Geochemical signatures of soils from north of the crater and their relationship to airborne radiometric geophysical data were discussed by Boamah and Koeberl (2002).

Various geophysical and remote sensing studies of the Bosumtwi structure have been carried out (Karp et al., 2002; Scholz et al., 2002; Wagner et al., 2002; Pesonen et al., 2003). Koeberl and Reimold (2005) published an updated and revised geological map of the Bosumtwi structure and environs, with explanations containing more detail about the impact structure.

The Bosumtwi impact crater was excavated in lower greenschist facies metasediments of the 2.1–2.2 Ga Birimian Supergroup. These supracrustals comprise interbedded phyllites and meta-tuffs together with meta-graywackes, quartzitic graywackes, shales and slates. Birimian metavolcanic rocks (altered basic intrusives with some intercalated metasediments) reach out to the southeast of the crater. Rocks to the southeast of the crater contain altered basic intrusives (Birimian metavolcanics) in addition to metasediments. Further to the east and southeast occur clastic Tarkwaian sediments, thought to have been formed by erosion of Birimian rocks (for details, see Koeberl and Reimold, 2005).

Several Proterozoic granitic intrusions are found in the structure, and some strongly weathered granitic dikes occur in the crater rim (e.g., Reimold et al., 1998a) and in part have aplitic appearance. Some of these granitoid dikes have granophyric texture (Reimold et al., 1998a). The granitic complexes and dikes probably mainly belong to the ca. 2.0–2.2 Ga Kumasi-type granitoid intrusions (see also Karikari et al., 2007; Losiak et al., 2013). In addition, a few dikes of dolerite, amphibolite, and intermediate rocks (minor intrusives) occur around the crater (e.g., Koeberl and Reimold, 2005). In the immediate environs of the crater, greywacke and sandstone/quartzitic rocks dominate, but especially in the

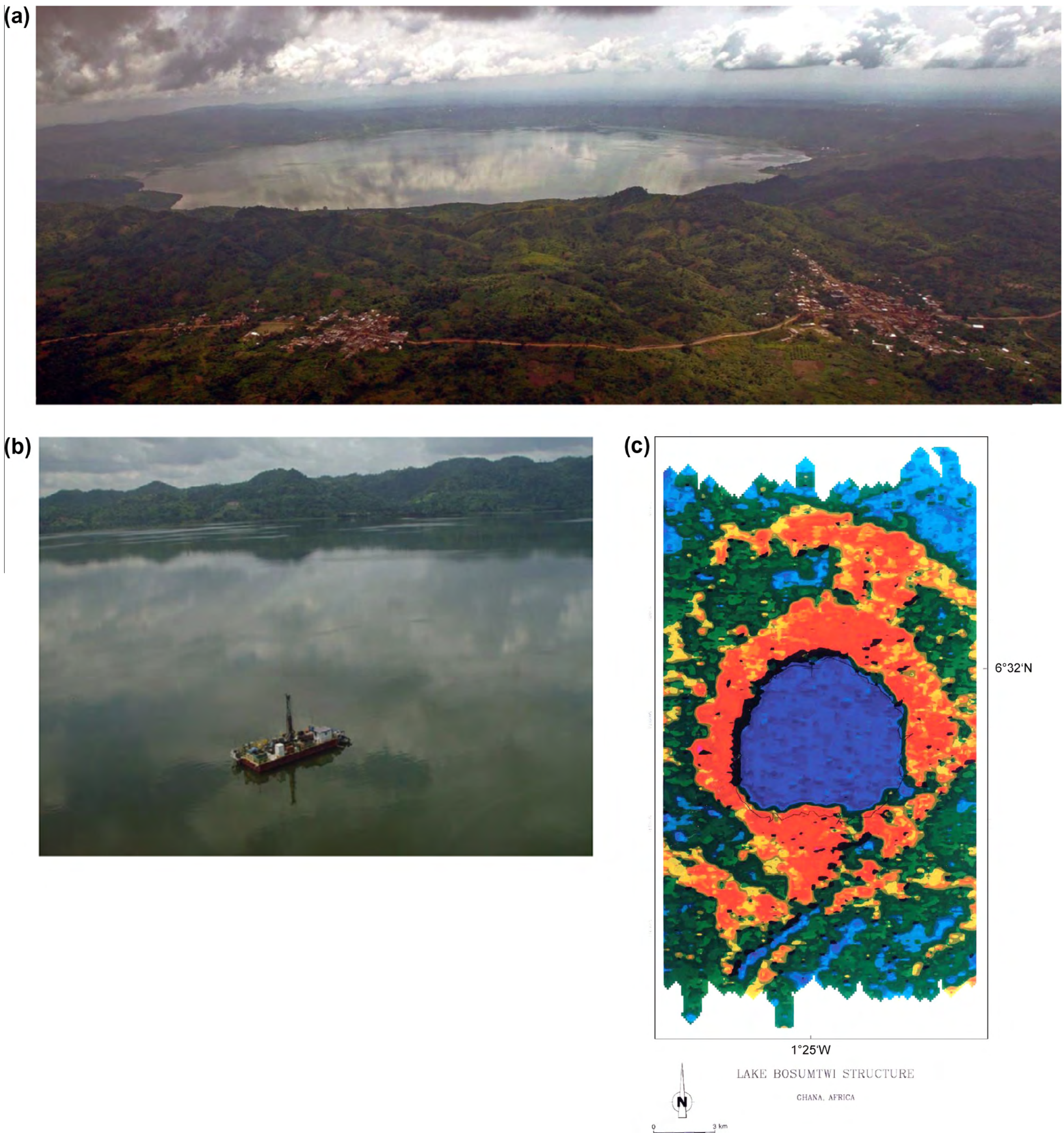


Fig. 23. The Bosumtwi impact structure, Ghana. (a) Aerial view (from the west), showing the lake and the crater rim. (b) The GLAD-800 drill rig during the ICDP scientific drilling project in 2004, on Lake Bosumtwi. (c) Map of the airborne potassium radiometry data, showing the chemical signature of an outer ring, which otherwise is topographically barely noticeable.

northeastern and southern sectors, shale and mica schist are also present (e.g., Reimold et al., 1998a). Quartz veins and stringers of up to 20 cm width cut through all the rock formations in the area, or occur in the form of pods. In addition to information in Koeberl et al. (1998b) and Koeberl and Reimold (2005), petrographic and geochemical details of the country rocks at Bosumtwi are given by Karikari et al. (2007) and Ferrière et al. (2010b,c). Losiak et al. (2013) report ages as well as chemical and isotopic compositions of granites from the Bosumtwi structure and its immediate surroundings and note that the similar composition (strongly peraluminous muscovite granites and granodiorites) and age (between 2092 ± 6 Ma and 2098 ± 6 Ma) of all granitic intrusions in the proximity of the Bosumtwi crater suggest that they are co-genetic.

Recent rock formations include the Bosumtwi lake beds, as well as soils and breccias associated with the formation of the crater (see review by Koeberl and Reimold, 2005). The breccias at Bosumtwi can be grouped into three types, based on composition and texture. These are the apparent monomict breccia (thought to be autochthonous); lithic breccia or polymict breccia (allochthonous); and breccia with glass fragments (suevite) (also allochthonous) (see, e.g., Boamah and Koeberl, 2002, 2003, 2006; Koeberl and Reimold, 2005). Monomict breccias often grade into unbrecciated rock. The rocks are shattered, more or less in situ, without much relative displacement of fragments. Shattered rocks consist of angular fragments of different sizes, irregularly distributed and recemented in a matrix of the same but finer-grained material. This type is found, e.g., on the road from Nyameani to Asisiriwa, and along the crater wall.

Rarer is the Bosumtwi suevite, a glass-bearing breccia similar to the suevite of the Ries crater in Germany. The Bosumtwi suevite is grayish in color, with a lot of glass (melt) and clasts up to about 40 cm in size. It contains target rocks in all stages of shock metamorphism, including vitreous and devitrified impact glasses. The Bosumtwi suevite occurs as large blocks of up to several meters width and as patchy massive deposits (compare Fig. 3b) more or less covered by thick vegetation in a marginal zone (about 1.5 km^2) outside the rim of the crater in the north, about 2.5 km from the lakeshore (location in area 1°23.5′–1°24.5′W and 6°33.5′–6°34.2′N). One such outcrop comprises massive suevite, and its exposure allows to estimate a thickness of about 2 m. It contains melt inclusions and rock fragments (graywacke, phyllite, shale, granite) up to about 40 cm in size, with graywacke dominating. Shallow drill cores (Boamah and Koeberl, 2002, 2003, 2006) were obtained to the N of the crater rim. The suevite cores show that melt inclusions are present throughout the whole length of the cores, in the form of vesicular glasses, with no significant change of abundance with depth. Graywacke-phyllite and granite dikes seem to be important contributors to the compositions of the suevite and samples from various roadcuts (fragmentary matrix), with a minor contribution of Pepiakese Granite. The thickness of the fallout suevite in the northern part of the Bosumtwi structure was determined to be up to 15 m, and occupying an area of about 1.5 km^2 (Boamah and Koeberl, 2002, 2003). The present distribution of the suevite is likely a result of differential erosion and does not reflect the initial areal extent of the continuous Bosumtwi ejecta deposits. An additional suevite outcrop to the south of the crater is discussed by Coney et al. (2010). Remarkably, the clast population in this breccia is distinct from that of the northern suevite, with more sedimentary (including much shale) and less granitic clasts present in the south.

In 1997, a high-resolution aerogeophysical survey was conducted to obtain more detailed information of the subsurface structure below and beyond the lake (cf. Pesonen et al., 2003; Koeberl and Reimold, 2005). From some of these data, Plado et al. (2000) produced a magnetic model for the Bosumtwi structure. The magnetic data show a circumferential magnetic halo outside the crater, at a radial distance from the center of ~6 km. The central-north part of the lake reveals a central negative magnetic anomaly with smaller positive side-anomalies N and S of it, which is typical for magnetized bodies at equatorial latitudes. A few weaker negative magnetic anomalies exist in the areas of the eastern and western parts of the lake. Plado et al. (2000) also reported petrophysical data on Bosumtwi impactites and country rocks, which show that the suevites have comparatively higher magnetization and have lower densities and higher porosities than the target rocks. In suevites, the remanent magnetization dominates over induced magnetization.

A shallow, near-circular, very slight depression at ca. 7–8.5 km from the structural center of the crater, and a shallow outer topographic ring feature at 18–20 km diameter, which was already noted by Jones et al. (1981) and later discussed by Wagner et al. (2002), is evident not only in radar satellite images (e.g., Fig. 3c–e in Koeberl and Reimold, 2005), but also in aero-radiometry data (Fig. 23c; Pesonen et al., 2003), indicating lithological as well as topographic control. Wagner et al. (2002) suggested that preferential removal of ejecta within the area just outside of the crater rim could be the reason for this shallow depression; original depositional patterns as well as impact-induced concentric fracturing could also be involved.

The Bosumtwi crater has also been of special interest as the likely source crater for the Ivory Coast tektites, which occur in one of only four known tektite strewn fields (the other being the North American, Central European, and Australasian tektites). Bosumtwi has been identified as the source crater of this tektite strewn field (see below).

In 2004, an international and multidisciplinary drilling project was undertaken (see Koeberl et al., 2007a,b). The project, which was conducted by DOSECC for the International Continental Scientific Drilling Program (ICDP), involved two main objectives: (1) to determine a 1 million year continuous paleoclimatic record from detailed multidisciplinary investigations of the post-impact crater fill sediments; such an extended record for the equatorial zone has not yet been accessible; and (2) to obtain a complete section through the impact breccia deposits in the central parts of the crater, both on top of the central uplift and in the surrounding crater moat. Previously, impact breccias at Bosumtwi were only known from the crater environs, and this drilling project allowed for detailed comparison of impactites within and outside of the crater.

From June to October 2004 16 drill cores were recovered at six locations within the 8.5-km-diameter Lake Bosumtwi. Fourteen sediment cores, representing a total of 1833 m aggregate length, completely sampled the Quaternary to Recent lake sediments, and two continuous cores with a total length of ca. 360 m were obtained from the impactites and underlying basement. In addition, extensive geophysical studies were carried out, as described by Elbra et al. (2007), Hunze and Wonik (2007), Kontny et al. (2007), L'Heureux and Milkereit (2007), Morris et al. (2007a,b), Schell et al. (2007), Schmitt et al. (2007), Scholz et al. (2007), and Ugalde et al. (2007a,b,c).

The two impactite cores, LB-07A and LB-08A, were drilled into the deepest section of the annular moat (540 m) and the flank of the central uplift (450 m), respectively. Samples from these cores have been studied by more than a dozen different research teams from around the world (cf. Koeberl et al., 2007a). At both impactite holes, drilling progressed through the impact breccia layer into fractured bedrock. LB-07A comprises lithic (in the uppermost part) and suevitic impact breccias with small but appreciable amounts of impact melt fragments. The lithic clast content is dominated by graywacke, besides various metapelites, quartzite, and a carbonate target component. Shock deformation in the form of quartz grains with planar microdeformations is abundant. Details on core 7A are published in Coney et al. (2007a,b) and Morrow (2007). Core LB-08A comprises suevitic breccia in the uppermost part, followed with depth by a thick sequence of graywacke-dominated metasediment with suevite and a few granitoid dike intercalations. It is assumed that the metasediment package represents bedrock intersected in the flank of the central uplift. A detailed lithostratigraphic column of drill core LB-08A (Deutsch et al., 2007; Ferrière et al., 2007a,b) shows that the drill core consists of approximately 25 m of impact breccia above fractured/brecciated metasediment (i.e., basement).

There are some interesting geochemical and petrographic differences between the crater fill breccias and the impact breccias from outside the Bosumtwi impact structure (e.g., Boamah and Koeberl, 2006; Coney et al., 2010). The relative amount of shocked and melted material in suevites from the area of the central uplift is significantly lower than that of suevite from outside the northern crater rim; this difference should represent differences in the ejection and deposition modes. Both suevites display some differences in major element abundances, mainly in the MgO, CaO, and Na_2O contents that could be related to the higher degree of alteration of samples from LB-08A suevite than suevite from outside the crater rim. Some differences in the clast populations of the two suevite facies can also induce some variations of major and trace element abundances (Coney et al., 2010). It is possible to clearly distinguish the different metasediment lithologies, on the basis of their compositions, using major elements abundances and Sc, Cr, Co, Ni, and Zn contents, which are noticeably distinct for these lithologies. No evidence for a meteoritic component has been detected in these breccias (cf. Ferrière et al., 2007a,b; Goderis et al., 2007; McDonald et al., 2007), in agreement with previous work (e.g., Dai et al., 2005).

Using the optical microscope and the transmission electron microscope (TEM) allowed to characterize the microstructure of the planar deformation features (PDFs) in quartz grains from several samples (from both, suevite and meta-graywacke); the observations of decorated PDFs in Bosumtwi samples argue for an impact into a water-bearing metasedimentary target or, possibly, rapid post-impact alteration (cf. Koeberl et al., 2007a).

The universal-stage technique was used for the investigation of the crystallographic orientations of PDFs in 18 meta-greywacke samples from the basement interval intersected by core LB-08A. Orientations of over one thousand sets of PDFs were measured in several hundred quartz grains to derive the distribution of shock metamorphic effects with depth, as specific orientations of PDFs in quartz are formed at different pressures. These investigations, by Ferrière et al. (2008), have shown that there is a variation of shock pressure in the uppermost part of the central uplift; the shock wave attenuation was of the order of about 5 GPa over the cored part of the central uplift. Subsequent numerical modeling allowed to reconstruct specific displacements of rocks during crater formation and to calculate the apparent shock attenuation along the about 200 m of investigated core (Ferrière et al., 2008).

A comparative study of the orientation of sets of PDFs in quartz grains from LB-08A suevites and from suevite from outside the crater rim was also performed and revealed some differences. Ferrière et al. (2009a) also performed a detailed and comparative study on ballen silica from Bosumtwi (occurring only in suevite samples from outside the crater rim) and a number of other impact structures. Using optical microscopy, cathodoluminescence (CL), Raman spectroscopy, and TEM techniques, it was possible to distinguish different types of ballen possibly related to varied formation conditions. During these investigations, coesite was for the first time characterized (using Raman spectrometer) in ballen cristobalite from the Bosumtwi structure (for details, see Ferrière et al., 2009a).

To compare the different target rocks and impact breccias recovered in drill core LB-08A with samples from outside the crater rim and with Ivory Coast tektites, various lithologies were analyzed for their Rb–Sr and Sm–Nd isotopic compositions (Ferrière et al., 2010a). The results are very similar to earlier studies of the Rb–Sr and Sm–Nd isotope systematics of Bosumtwi crater target rocks. Stable carbon isotope investigations on carbon-rich shale and phyllite samples and on selected clasts in impact breccia samples were also performed. Our values in $\delta^{13}C$ obtained for all measured samples strongly suggest an origin from biogenically derived carbon. SHRIMP U–Pb zircon dating of one suevite sample and of two meta-graywacke samples was reported by Ferrière et al. (2010c) and yielded an upper Concordia intercept age of 2145 ± 82 Ma, in very good agreement with previous geochronological data for the West African Craton rocks in Ghana. The Rb–Sr and Sm–Nd data reported by these authors show that the suevites are mixtures of meta-greywacke and phyllite (and possibly a very low amount of granite).

In core LB-05B, one of the cores drilled to study the lake sediments, the zone between the impact breccias and the post-impact sediments was completely recovered – including the basal, fine-grained impact fallback (Koeberl et al., 2007c,d). This about 30 cm thick layer contains in the top 10 cm accretionary lapilli, microtektite-like glass spherules, and shocked quartz grains. On average, the composition of the fallback spherules from core LB-5B is very similar to the composition of Ivory Coast tektites and microtektites, with the exception of CaO contents that are about 1.5–2 times higher in the fallback spherules (Koeberl et al., 2007c,d). This is a rare case in which an immediate post-impact fallback layer has been preserved in an impact structure; its presence indicates that the impactite sequence at Bosumtwi is complete and that Bosumtwi is a well-preserved impact crater.

However, these were more or less the only glassy components recovered in the cores. Prior to drilling, numerical modeling estimated melt and tektite production using different impact angles and projectile velocities. The most suitable conditions for the generation of tektites are high-velocity impacts (>20 km/s) with an impact angle between 30° and 50° from the horizontal (Artemieva, 2002). Also, a remote magnetic survey was interpreted to reveal the presence of a substantial amount of melt underneath the crater lake (Plado et al., 2000). This ascertion had to be re-interpreted based on the new core data (Ugalde et al., 2007c). The observed situation for breccias within and around the crater is very different from the earlier model results, as much more melt was predicted than is actually observed. Clearly, much more melt has been incorporated in the suevite ejected outside of the structure, in comparison with the low amounts observed in the within-crater suevite occurrences (cf. Coney et al., 2010). The lack of a coherent melt sheet, or indeed of any significant amounts of melt rock in the crater fill, is thus in contrast to expectations from modeling and pre-drilling geophysics, and presents an interesting problem for comparative studies and requires re-evaluation of existing data from other terrestrial impact craters, as well as modeling parameters (Artemieva, 2007). Apparent melt deficiency has also been discussed for the Ries impact crater (southern Germany) and has been addressed through recent mineralogical analysis of impact breccias from within the crater (Reimold et al., 2011a) and through numerical modeling (Artemieva et al., 2013; Stöffler et al., 2013).

Other studies resulting from the ICDP drilling project include a search for biological activity in the form of archaeal membrane-lipids, which were detected in impactite core samples and might be related to the soils or rocks predating the impact event, the hydrothermal system generated after the impact, or due to more recent underground water transport (Escala et al., 2008). The cosmogenic radionuclide Be-10 was used to investigate as to whether surface-derived material is present in the suevitic breccia within the drill core of the Bosumtwi crater so that the extent of mixing of target rocks during crater formation in respect to the fallback breccia can be determined; it was found that only a small number of clasts in the suevite was derived from near the target surface, indicating that in-crater breccias were well mixed during the impact cratering process (Losiak et al., 2014).

As Lake Bosumtwi is a hydrologically closed lake, which lies beneath the path of the seasonal migration of the Intertropical Convergence Zone (ITCZ), it can provide a sedimentary record of monsoon variability in West Africa. The continuous 300-m-long drill cores obtained from Lake Bosumtwi, Ghana, represent one of the longest, continuous lacustrine sequences obtained from an extant lake. Most of the lacustrine record represented by the Bosumtwi sediment cores shows fine lamination (varves). There are intervals of non-laminated sediment with increased density, decreased organic content, and a high-coercivity magnetic mineral assemblage. Some of these massive layers contain slump-folding and intraformational clasts. These lithologies are interpreted to represent lake-level lowstands when a diminished West African summer monsoon resulted in decreased moisture balance and lake-level regression. Increased amounts of high-coercivity magnetic minerals corresponding to glacial stages are also present in the Bosumtwi lacustrine sediment cores. These were interpreted to result from elevated aerosol dust export from arid Sahel sources, possibly accompanied by enhanced magnetic-mineral diagenesis during lake-level lowstands based on the magnetic signature of these cores (e.g., Peck et al., 2004). Thus Bosumtwi contains an unprecedented record of late Quaternary climate change in West Africa (e.g., Shanahan et al., 2006, 2008, 2013).

In summary, the drilling project at Bosumtwi has contributed not only a wealth of new information about the crater itself, but also provided important new data and improved our understanding of global change and impact processes.

In the course of the ICDP drilling project it was also evaluated in how far the Bosumtwi structure and lake had geo- and eco-tourism potential (Boamah and Koeberl, 2007). Ambitious plans to develop a museum site have not come to fruition yet, but there is some reasonable hotel infrastructure. With fishing having become a somewhat limited affair as a result of overfishing, some of the local inhabitants have taken to arts-and-crafts production. Subsistence farming in the area around the lake includes widespread cocoa plantations.

6.1.6. B.P., Libya

There are two long-known impact structures in southeast Libya, known as the B.P. (Fig. 24) and *Oasis* (Fig. 30; see below) craters. Both were discovered by petroleum exploration geologists and named after their respective oil companies. The B.P. and Oasis structures have long been thought to be possibly of the same age as the enigmatic Libyan Desert Glass (see below) occurring to the northeast of their locations, in Egypt, which generated strong interest in these structures (e.g., Abate et al., 1999).

The B.P. structure, located at 25°19′N/24°20′E, was first referred by Martin (1969). It is a complex structure with a small central area of strongly deformed, upturned strata, and a prominent ring of up to 50 m high hills at 1 km from the center of the structure. Still further out, at about 1.4 km distance from the center of the structure, another, subdued structural ring feature is noted, of no more than 10 m elevation. Detailed field analysis demonstrated (Koeberl et al., 2005a) that it is a shallowly inward dipping fault structure that nearly encircles the entire crater structure. Already the morphology of this crater structure suggests that it is deeply eroded, with the hills representing the remaining roots of the crater rim not showing overturning, and only a small fragment of the original central uplift having remained. No crater-fill breccia has been found anymore. The rocks in the B.P. area are more or less ferruginous sandstones, with intercalated siltstones and local conglomerate exposures. They are believed to belong to the Nubia Sandstone Formation for which an Early Cretaceous age is given in the literature. It should be observed, however, that the stratigraphy and age of the Nubia Formation is quite controversial, and in Libyan and Egyptian mapping reports discussed strongly variably (see also below, *Oasis structure*). Ferruginous sandstones are capping many of the ranges and inselberge in the environs of BP structure, generally with (sub)horizontal attitudes of the strata (slight warping of the layers has been observed in places). Contrary to this, the rocks at BP are suspiciously deformed at macro- to mesoscopic scales.

The geology of the structure was studied in some detail by Underwood (1975, 1976) and Underwood and Fisk (1980). Koeberl et al. (2005a) reported a remote sensing study and first outcomes of their fieldwork of 2001. They noted that ERS1 and Radarsat imagery was not very useful for assessing the geometry of the structure, although the radar image showed the positions of the ring structures clearly. Their field work indicated that the actual diameter of the BP structure is just about 2 km. The crater rim (their "middle ring") was characterized by a distinct series of hills of up to 30 m elevation above the surrounding desert, with sandstone dipping at 30–50° outward. They observed that some parts of the rim were strongly folded and faulted. The innermost elevation is a complex terrain of strongly folded sandstone hills, quite a few of which show steeply upturned bedding.

The age of B.P. is only constrained by the ill-defined age of the Nubia Sandstone Formation (90–120 Ma) that provides an upper age limit for the formation event. No datable phases (such as melt breccia or authigenic minerals) have been discovered to date in the crater area.

Abate et al. (1999) reported a detailed chemical study of Libyan Desert Glass and of BP and Oasis sandstone samples. They noted a slight similarity between Libyan Desert Glass and their samples from these impact structures, but as summarized by Koeberl et al. (2005a), more recent Rb–Sr and Sm–Nd isotopic results do not support that Nubian rocks are precursors for the Libyan Desert Glass (Schaaf and Müller-Sohnius, 2002). This is discussed in more detail in the section on the Libyan Desert Glass below.

According to French et al. (1974), medium- to coarse-grained orthoquarzite was sampled with quartz grains that displayed multiple sets of "planar elements", which they interpreted as definite evidence for shock metamorphism and, thus, an impact origin of the B.P. structure. (The term "planar elements" at the time encompassed all types of planar microdeformation features related in origin to impact. In particular, PDFs and planar fluid inclusion trails were summarily termed planar elements.) A 5-day mapping visit to B.P. in 2001, by the authors of this review, did not yield any samples that displayed diagnostic planar deformation features, albeit planar fractures and other, non-diagnostic deformation bands, were observed.

In 2010, one of us (WUR) revisited B.P., for comprehensive resampling of the central uplift area. First thin sections revealed shock metamorphosed quartz in just a few samples. One or two sets of PDFs occur in – again rare – quartz grains. PDFs are generally short and occur in very small patches within, or at the edges of, the host grains. In addition, planar fractures occur, in up to three sets per host grain. Feather features have been observed as well, whereby either planar fractures or segments of a grain boundary represent the "quills", and the short "feathers" are variably inclined or perpendicular to the quill. They commonly are not planar but curved. Several samples from the central uplift of B.P. are characterized by severe reduction of porosity, to the effect that the samples have been compressed so strongly that abundant quartz crystals abut against each other and form radial or conchoidal concussion fractures. Examples of shock deformation are shown in Fig. 24d–f.

6.1.7. Gweni Fada, Chad

Gweni Fada at 17°25′N/21°45′E was first noticed by a French team on Landsat images and aerial photography (Vincent and Beauvilain, 1996). The structure is located some 320 km southeast of the Aorounga impact structure and 30 km northeast of the Fada palm grove in the Ennedi district of northern Chad. Vincent and Beauvilain (1996) estimated a diameter of about 14 km. They also visited the crater structure and collected some samples, for which they reported some preliminary petrographic results including shock metamorphic effects in quartz grains from sandstone.

The structure is slightly asymmetric, with a somewhat longer diameter in northwest–southeast direction. Like the Aorounga structure, Gweni Fada appears strongly eroded. A broad depression of about 12 km diameter forms a crescent around two-thirds of the structurally complex innermost zone. This broad synclinal feature has a distinct outer limit formed by apparently steeply dipping limestones (with unknown dip directions). On the north side, the syncline is surrounded by an elevated outer ring of outward-dipping sandstones. According to Vincent and Beauvilain (1996), the external depression does not extend into the south, where tilted or folded sandstones were described. In the innermost rugged

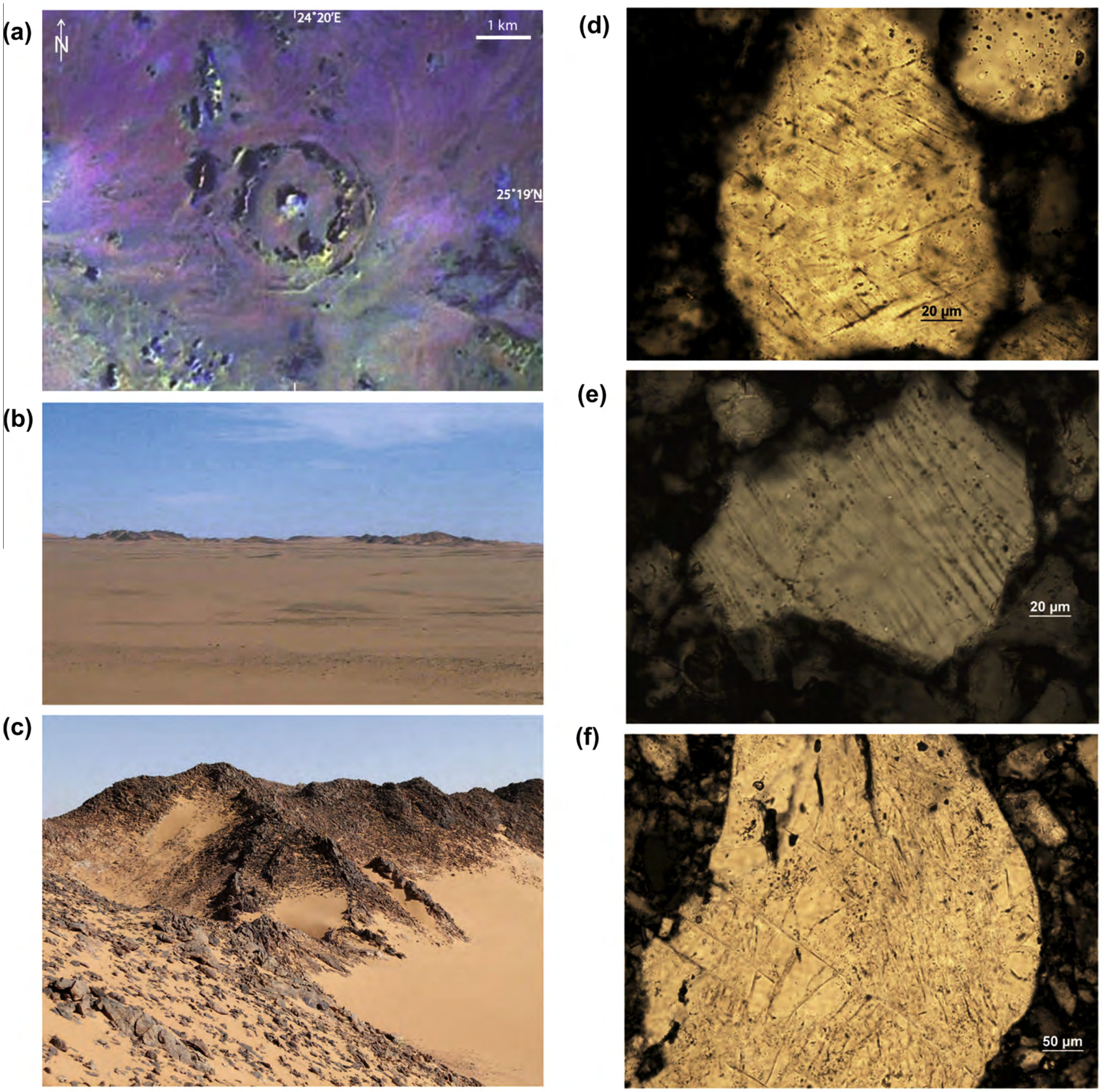


Fig. 24. BP impact structure (Libya). (a) False-color Landsat image of the BP impact structure and surroundings. Note the double ring structure along the outer part of the structure, and the inner central uplift feature (compare text for detail). (b) First impression of BP crater structure upon approaching it from the north. Note the prominent, hilly crater rim anticlinorium formed by strongly blockfaulted sandstone strata. The structure is about 2 km wide. (c) Part of the structurally complex central uplift. Note the fold structure on the highest point, and the repeatedly folded sandstone layer encircling the sand-patch in the center of this image on its right side, and then curving into a trend towards the bottom right again. Width of field of view estimated at about 80–100 m. (d–f) Typical microdeformation features observed but rarely in quartz from sandstone samples of the central uplift. (d) Relatively widely spaced planar fractures in NE–SW and NW–SE direction, with far more densely spaced PDFs in the same orientations. (e) Two sets of PDFs in a quartz grain along NW–SE and north-northwest–south-southeast directions. Note that PDFs are partially decorated with fluid inclusions. (f) Relatively widely spaced planar fractures in northwest–southeast orientation, and a much more densely spaced set of PDFs in north-northwest–south-southeast direction. There are short, somewhat curved fractures coming off some of the planar fractures, sometimes resembling feather feature development, which is also quite abundant in these samples.

terrain, hills of several hundred meters elevation are made up of sandstones of varied dips, generally in outward direction. According to Wacrenier et al. (1958), Gweni Fada – like Aorounga – occurs in Upper Devonian strata. This represents the only constraint on the age of this impact structure.

Koeberl et al. (2005a) investigated Landsat 5 satellite imagery in combination with Shuttle Radar Topography data. They found the near-circular structure slightly offset from its center to the south. They described the structure as "...structurally complex, terrain is surrounded in the western, northern and eastern sectors by an apparently flattish terrain. Radial trends of both ridges and apparent drainage outward from the elevated terrain surrounding the flat ring zone is quite obvious." These authors observed on SRTM data a distinctly different terrain structure inside the crater structure in comparison to its environs. From their image interpretations they considered the actual diameter of the Gweni Fada structure to be much larger than previously assumed, with a diameter of actually 22–23 km. They observed that Therriault et al. (1997) estimated for an impact structure of such size that the central structural uplift area would measure 7–9 km – just about what Koeberl et al. (2005a) determined for Gweni Fada.

A few samples of quartzite, sandstone, and quartz conglomerate, both from the central area and from the outer margin of the ring depression, were analyzed petrographically. Of these, several samples exhibited up to 2 sets of PDFs in quartz grains, although at relatively low overall frequency. Cataclastic bands, in some samples quite prominent, were also recorded and seem to be concentrated in samples where PDF formation is very limited (i.e., these samples seemingly represent a somewhat lower shock pressure regime). Nevertheless, the arenitic samples from Aorounga and Gweni Fada, both moderate-sized impact structures, could provide welcome shock petrographic evidence in comparison with samples from sandstone-hosted smaller impact structures. It might be worthwhile collecting samples along continuous profiles from the center of these structures radially outwards, for a dedicated shock metamorphic study.

6.1.8. Kalkkop, South Africa

Kalkkop (Koeberl et al., 1994a; Reimold et al., 1998b) is a small, 640-m-wide, nearly circular structure in the Eastern Cape Province of South Africa, located at 32°43′S/24°34′E between the towns of Graaff-Reinet and Jansenville. On aerial photographs the structure stands out as a bright disk surrounded by a darker annulus (Fig. 25a). The interior of the crater is filled with light-colored limestone, and the immediate ring surrounding it consists of poorly exposed, upturned sandstone and shale. The best exposure found by the authors is located directly southwest of the road along the northern side of the structure.

Kalkkop was drilled in the 1940s, in the cause of energy-resource evaluation of the Karoo Basin. Further drilling with extensive core recovery was carried out in 1993 (Reimold et al., 1998b). The subsequent detailed petrographic and geochemical studies provided unambiguous proof for an impact origin. Conspicuous though non-impact diagnostic fracturing is widespread in clasts of brecciated bedrock, but the important evidence came in the form of – admittedly rare – PDFs in quartz in suevitic breccia. This material contains a very small component of impact melt that is partially altered by carbonate. Re–Os isotopic analysis of breccia, as well as some Beaufort Group (of the Karoo Supergroup) country rocks (sandstone and mudstone) by Koeberl et al. (1994a) represented one of the first successful applications of this isotope technique for the confirmation of the presence of a meteoritic component (Fig. 25b).

U–Th isotopic analysis of upper and lower crater fill indicated the age of the impact to be in the order of 250 ± 50 ka (Reimold et al., 1998b) – within error limits similar to the age of the Tswaing crater (see below). This has caused some speculation whether Kalkkop and Tswaing could be the results of a dual impact event. To resolve this issue, it would be necessary to determine the compositions of the two projectiles implicated in the formation of Kalkkop and Tswaing.

6.1.9. Kamil, Egypt

The Kamil Crater in southern Egypt (Fig. 26a) is a less than 5000 year old impact crater of 45 m diameter that also features a pristine ejecta ray pattern (Folco et al., 2010, 2011). The crater occurs on exposed pale sandstones (mainly quartz arenites) of the Early Cretaceous Gilf Kebir Formation that are locally overlain by a few centimeters of soil. The sandstones have subhorizontal

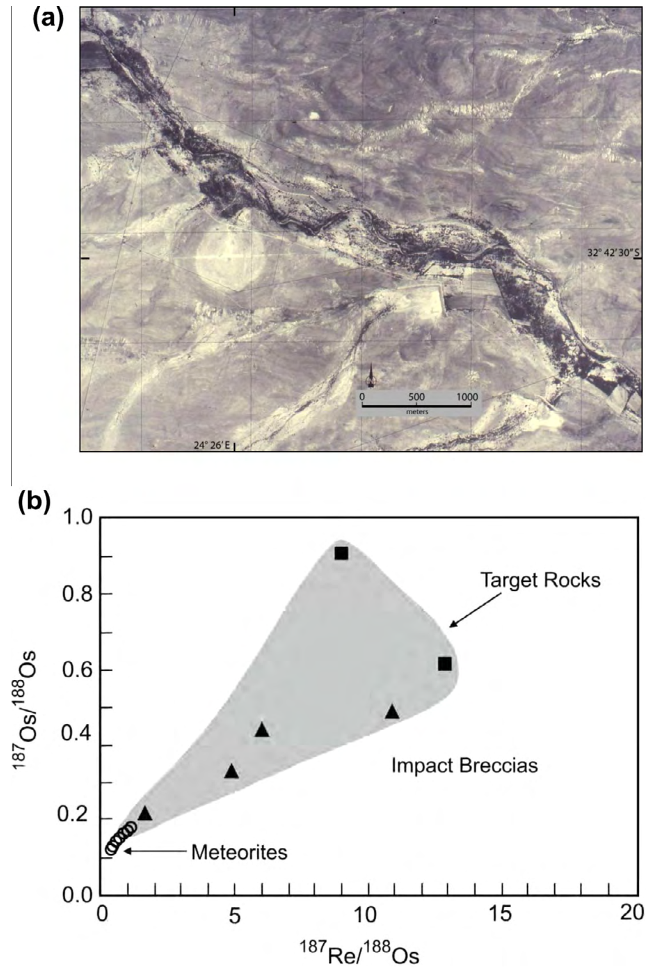


Fig. 25. (a) Aerial photograph of the 640 m wide Kalkkop impact crater in the Eastern Cape Province of South Africa. The Buls River flows tangentially past the crater. Crater fill is made up of limestone, which is clearly visible in this image. (b) Detection of an extraterrestrial component in impact breccias from the Kalkkop impact crater, South Africa, using a Re–Os isotopic diagram. The triangles represent different samples of impact breccia, the squares represent the two main target rock types (sandstone and shale; i.e., the background – or as it is also known, the indigenous – composition), and meteorite data are indicated by open circles. The diagram illustrates the use of this isotopic diagram as a mixing diagram (mixing between the indigenous component and the meteoritic contribution). After Koeberl et al. (1994a).

bedding and constitute part of the sedimentary cover uncomformably overlying the Precambrian crystalline basement. The bowl-shaped crater is circular with an average rim-crest diameter of 45 m and has an upraised rim ~3 m above the presumed pre-impact surface, with part of the northern wall of the crater being covered by a meter-thick eolian sand deposit. Over 5000 fragments (shrapnel) of iron meteorite (ungrouped, ataxite) specimens totaling a mass of about 1.7 tons have been found within the crater and in its vicinity (Fig. 26b and c); the largest specimen had a mass of 34 kg (Folco et al., 2011; D'Orazio et al., 2011). Thorough field studies resulted in a very detailed map of the strewn field of meteoritic debris (Fig. 26c). These fragments are assumed to have been produced by the explosion of the impactor upon hypervelocity collision with the target. Urbini et al. (2012) estimated a minimum original projectile mass of approximately 5 tons. Kamil is a very young and exceptionally well preserved small, simple bowl-shaped impact crater. It has been extensively studied by morphometric, structural and geophysical analysis, and as such it represents an excellent medium for the extension of the recent experimental MEMIN study, in which small impact craters at the scale of 10–20 cm have

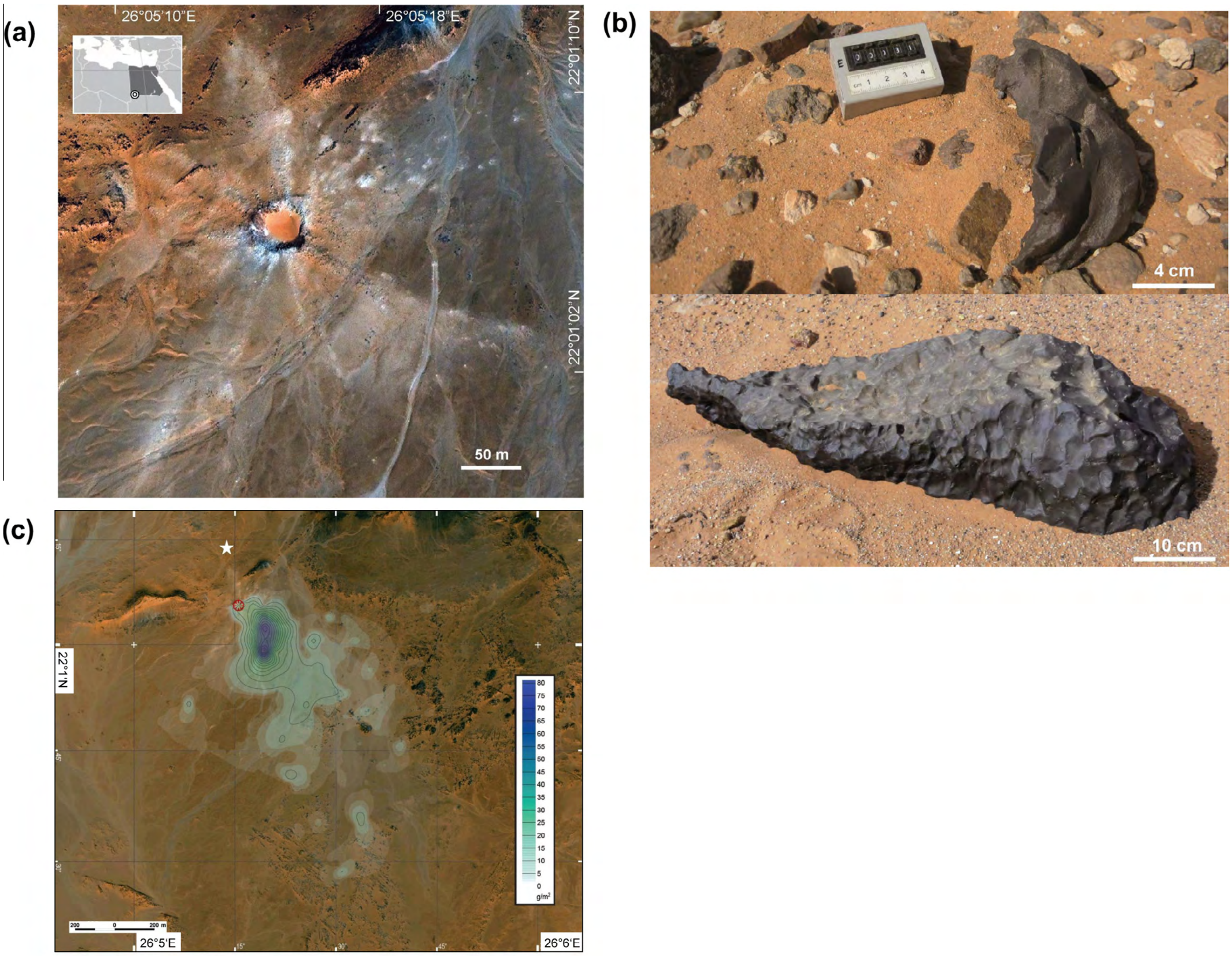


Fig. 26. (a) Quickbird satellite image of Kamil crater (inset indicates approximate location in southwestern Egypt) (after Folco et al., 2011). Note the pattern of ejecta rays, preservation of which testifies to a young impact age. (b) Shrapnel (top) and large, regmaglypted individual (bottom) of the Gebel Kamil meteorite (after Folco et al., 2011). (c) Gebel Kamil meteorite distribution map (g m^2) obtained through linear interpolation of average meteorite density values of 50 m × 50 m cells positioned at their centers. Contour lines are shown at 5 g/m^2 intervals. The white star shows the finding location of the 83 kg individual (after D'Orazio et al., 2011). Figures (a) and (b) courtesy of Luigi Folco, and figure (c) courtesy of Massimo D'Orazio (both at University of Pisa).

been generated in similar material (Kenkmann et al., 2011a, 2013a).

6.1.10. Kgagodi Basin, Botswana

Botswana is a country of ca. 580,000 km^2, of which about four-fifths are covered by the substantial deposits of the Cretaceous to Holocene Kalahari Group (Thomas and Shaw, 1990). Consequently, initial indications for the presence of a possible impact structure would mainly come from remote sensing (circular depression) or geophysical surveys. In addition, extensive drilling for water resources has been carried out in this semi-arid to arid terrane. Botswana's so far only confirmed impact structure, known as the Kgagodi Basin, does not have a significant surface expression (Fig. 27); it was first recognized in 1997/1998 in the course of a water drilling project in southeastern Botswana. The 3.4 km wide structure is located at 22°29′S and 27°35′E, ca. 7 km south of Kgagodi village. First, tentative indications of the presence of a small basin structure were derived from aerial photography and satellite imagery (Thomas, 1971; Paya et al., 1999). In 1997, the structure was drilled by the Geological Survey of Botswana to a depth of 274 m, as it was thought to be located at the intersection of two regional fault lines, which made it a significant hydrological target. On surface, there are only a few small calcrete outcrops in this area, basically limited to the strip along the subsurface crater rim line.

Initial investigation of the drill core by Paya et al. (1999) indicated the presence of breccia, which lead them to suggest that Kgagodi Basin might represent an impact structure. But first petrographic analysis failed to record bona fide shock metamorphic evidence. This only came in the following year (Reimold et al., 2000a), and a detailed account of the impact structure and the confirming evidence was finally published by Brandt et al. (2002). They reported a strong gravity low centered on the structure and a complex, noisy high-frequency magnetic signature. Both techniques delineate the crater area. The magnetic signature was interpreted to represent a number of shallow bodies of variable magnetic susceptibilities, as one would expect from a complex geological terrain composed of gneisses and migmatites and intruded by mafic dikes and pods.

The structure is formed in Archean granitoid basement, which, at the time of impact, was overlain and intruded by Karoo dolerite. This essentially provides the only firm upper age limit for the impact event, at about 180 Ma (e.g., Duncan and Marsh, 2006).

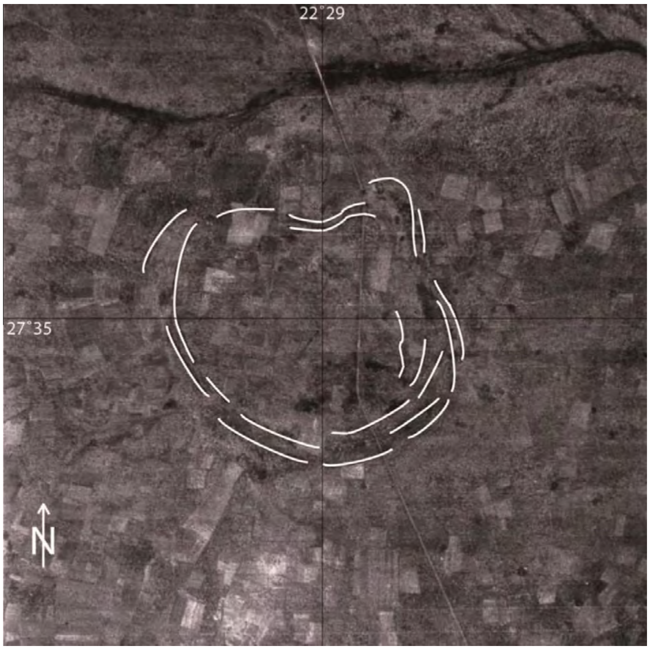


Fig. 27. Aerial photograph of the Kgagodi crater area in southeastern Botswana. The lines indicate the rough location of generally poor suboutcrop of the crater rim. The drill core that provided the proof of the impact origin for this geophysically recognized structure was obtained some 400 m inside of the northeastern rim.

Results of a gravity survey yielded a model consistent with a typical simple bowl-shape crater form of 3.4 km width. The drill core was recovered about 400 m inside of the crater rim. It comprises crater-fill sediment to a depth of 158 m. Impact breccia was then intersected only to the depth of 165 m, after which locally brecciated basement rocks were drilled that gradually change with depth into fractured and then coherent (undeformed) crystalline basement, reached at about 250 m depth.

Only in granitoid clasts of the narrow breccia zone was shock metamorphic deformation detected. Shock effects include multiple sets of planar deformation features in quartz and feldspar, diaplectic quartz glass, and partially to completely isotropized feldspar. In addition, rare melt fragments were observed, on the basis of which this breccia intersection was classified as suevite. Abundances of some siderophile elements, as well as of iridium, are well elevated in breccia samples above the levels of the target rocks, and this has been interpreted by Brandt et al. (2002) to indicate the presence of a small meteoritic component in the suevite.

These authors also provided some palynological evidence that allowed them to assign a tentative impact age to the upper Cretaceous to early Tertiary interval. The small impact crater has, thus, enormous potential to provide a long-term record of paleoenvironmental conditions for this part of the southern hemisphere, if the structure were drilled in its deepest part, where the most expanded paleorecord might be obtained. It is estimated from gravity analysis that it would take about 800 m of drilling to reach basement in this central area. For this reason, and the excellent accessibility (vicinity of a major highway) of this crater structure, Kgagodi Basin was proposed as an important target for paleoenvironmental studies (Reimold et al., 2005b). A further geophysical survey should proceed any drilling though, in order to determine whether there might be a small central uplift.

6.1.11. Luizi, DRC

Not only are large parts of central Africa heavily forested and not very suitable for remote sensing searches for impact structures, but large regions of the African continent have been rather inaccessible in recent decades because of widespread political strife, with the Democratic Republic of Congo (DRC) having been particularly ravished for decades by still continuing conflicts. Thus, it does not surprise that possible impact structures (*Luizi, Omeonga*) in its territory have not been studied until quite recently.

The Luizi structure (Fig. 28) is centered at 10°10′S and 28°00′E in a rather unexplored region on the Kundelungu Plateau of Katanga Province in the southeastern DRC. It was first mentioned in a geological report about a semi-circular basin by Grosse (1919). Preliminary descriptions from satellite imagery were contributed by Dumont (1990), who based his suggestion of an impact origin on the circular appearance of this structure, and by Claeys et al. (2008), who provided further remote sensing information. The impact origin of Luizi was finally confirmed by Ferrière and Osinski (2011) and Ferrière et al. (2011a,b).

Ferrière and Osinski (2011) described Luizi as a 17 km diameter, complex crater, with an intermediate ring at ~5.2 km radius from the center and a central ring at 1 km from the center enclosing a central depression. With this morphology, Luizi resembles the Brazilian Serra da Cangalha impact structure (Reimold et al., 2006; Kenkmann et al., 2011b; Vasconcelos et al., 2013). These authors concluded from ground-geological observation that the interior depression of the central uplift zone was caused by preferential erosion of the relatively poorly resistant lowermost target strata. At Luizi, the outer rim is elevated by some 300–350 m above the crater interior. Ferrière et al. (2011a) used a digital elevation model to indicate that the diameter of the structure is ca. 17 km. A special feature at the eastern side of the Luizi structure is a large regional fault zone.

Ferrière et al. (2011b) presented results of a first ground-based analysis of about 30 outcrops in the structure. According to this, the Luizi structure is formed in massive, tabular sandstone facies with intercalated argillaceous sandstones, all belonging to the Biano Subgroup. This sequence of late Neoproterozoic age represents the uppermost part of the Kundelungu Group (Master et al., 2005). Whereas these strata have subhorizontal attitudes in the outer part of the structure, they are characterized by moderately steep to vertical orientations in the inner parts. Shatter cones are well-developed, up to 40 cm in size, and occur in the inner 3.2 km of the structure. Up to 2 m thick dikes of monomict lithic breccia were observed up to 3 km from the center; samples of this breccia did not reveal shock deformation yet. In contrast, planar deformation features in quartz, and rarely in feldspar, were abundantly observed mainly in thin sections of shatter cones developed in arkosic sandstone (as well as in sandstone samples devoid of shatter cones) from the inner part of the structure (L. Ferrière, pers. comm., 2013). PDFs in quartz occur in up to 5 different orientations per host grain, and the authors concluded that the more heavily shocked target rocks at Luizi experienced shock pressures up to 20 GPa, within radial distance of 2 km from the center. Upon a further visit to the structure in late 2013, L. Ferrière noted a range of other breccias that, at this time, await detailed investigation.

The age of the Luizi impact is poorly constrained – only by the maximum age of sediments of the target region that was reported as ca. 573 Ma by Master et al. (2005). Further information provided by S. Master (U. Witwatersrand, pers. comm. 2013) includes the following: Batumike et al. (2007, 2008) dated the Kundelungu Plateau kimberlite cluster at about 32 Ma. As two of these kimberlite occurrences are located within the Luizi structure, these authors suggested that the structure likely was older than these kimberlites. However, clearly it is also possible that the structure could have formed after kimberlite emplacement – i.e., intrusion of the kimberlite into the structure. Master et al. (2001) argued that the fault zone on the east side of the structure (compare images in Fig. 28) had been overprinted by the Luizi event. Thus, the age of the faults would provide a maximum age for the impact. They considered these faults as part of the Lake Mweru-Luapula graben structure that had been dated on geomorphological grounds as Late Tertiary (Neogene) according to Dixey (1944). Hence, the Luizi structure could be younger than the late Neogene, i.e. <2 Ma old (Master et al., 2001). We feel that this would require a much accelerated rate of erosion for this clearly strongly degraded impact structure. Nevertheless, further work is required regarding the age of this impact event, as well as the high potential that the extensive exposures against the fault zone have to reveal much interesting information about the interior of this large, complex impact structure.

6.1.12. Morokweng, South Africa

A large impact structure is located in the area around Morokweng township in North West Province of South Africa (Fig. 29), centered at 26°28′S/23°32′E. This semi-arid to arid area is part of the southern Kalahari Desert. The crater area is mostly sand-covered and very rare rock exposures are mainly limited to river-beds and to a limited exposure of horizontally disposed quartzite and banded iron formation near Heuningvlei. The area of the structure is characterized by a sizable aeromagnetic anomaly in the area of the Ganyesa Dome made up of Archean granitoids. Andreoli et al. (1995) must be credited for having been the first to draw attention to this feature: they studied drill core from the Ganyesa Dome and reported the existence of impact melt rock, as well as shock deformation in underlying granitoids. Corner et al. (1997) investigated the geophysics of the wider region and found that a ca. 300 km wide, circular structure could be visualized in potential field data. As they also detected shock metamorphic deformation in the form of PDFs in quartz from a surface sample of a Transvaal Supergroup arenite, they could also support the existence of an impact structure.

Hart et al. (1997) investigated melt rock from drill cores and found a significant enrichment of siderophile elements in comparison to the compositions of country rocks. Koeberl et al. (1997a)

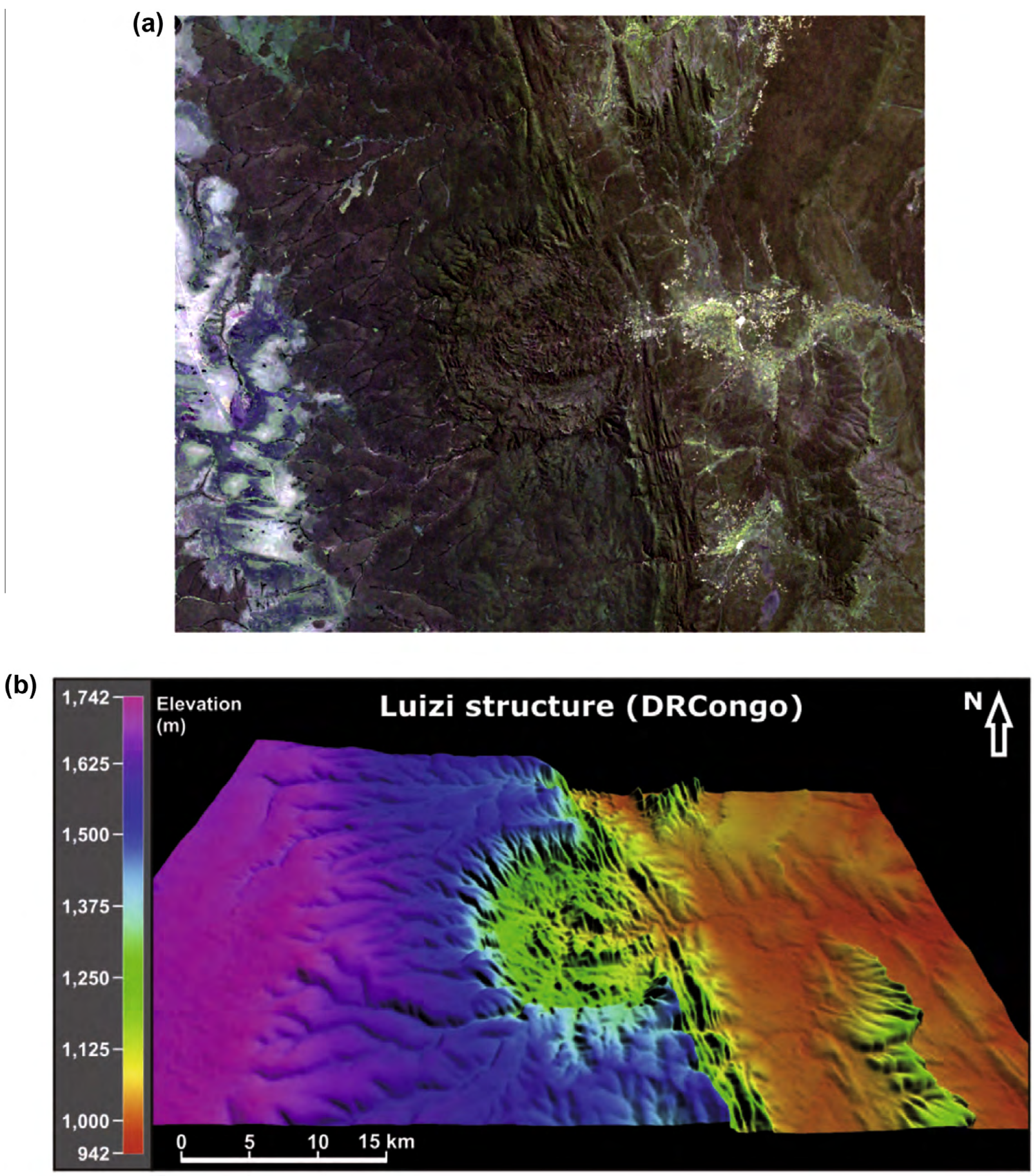


Fig. 28. Luizi impact structure, Democratic Republic of Congo. (a) Landsat view of the 17 km diameter Luizi impact structure. Images courtesy of Ludovic Ferrière of the Naturhistorisches Museum Wien. (b) Digital Elevation Model of the Luizi impact structure, based on the Shuttle Radar Topography Mission data (http://www2.jpl.nasa.gov/srtm/).

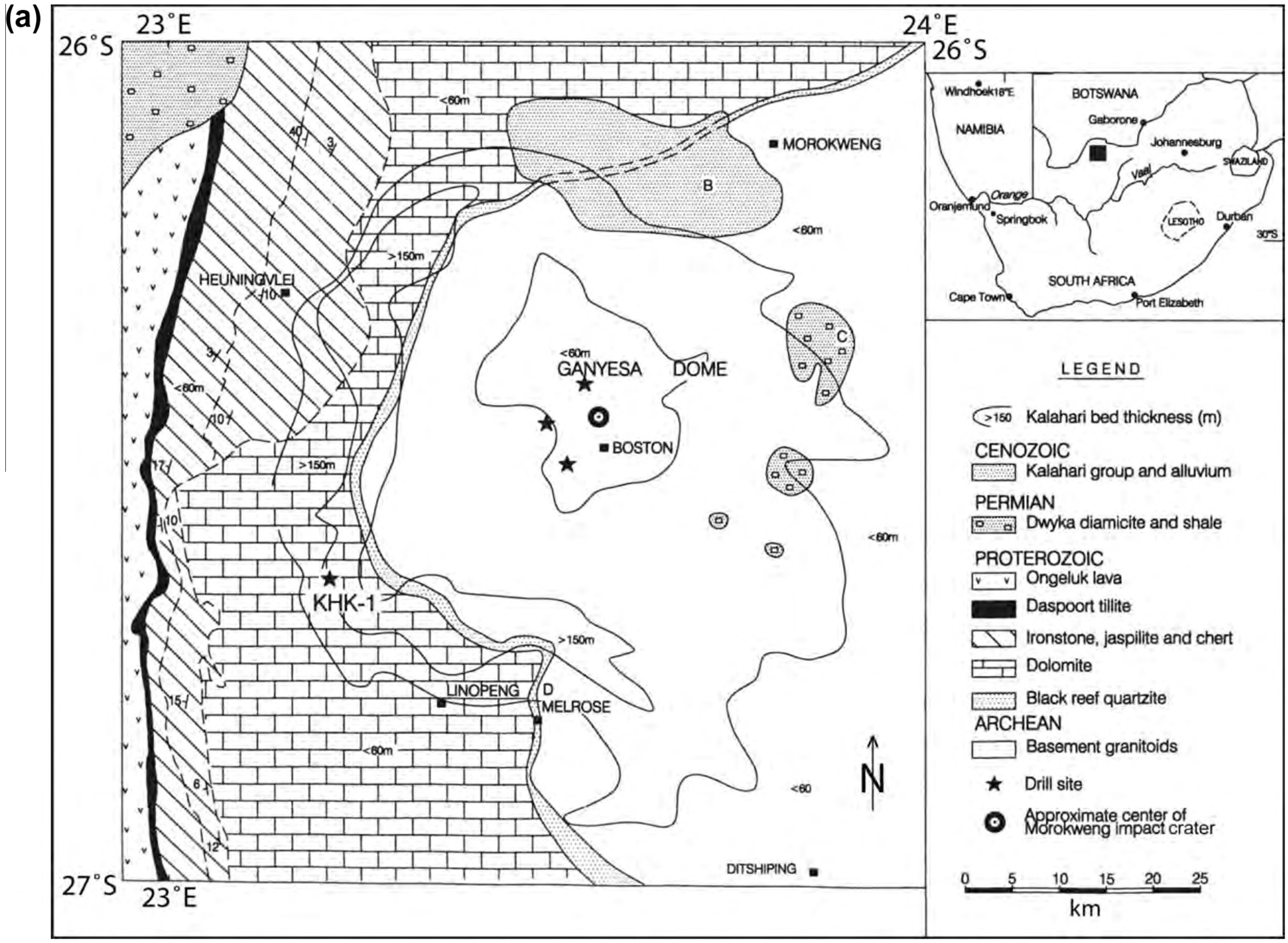


Fig. 29. (a) Simplified geological map of the area of the Morokweng impact structure, modified after a diagram by Reimold and Koeberl (2003). Inset shows the location of the Morokweng structure in North West Province of South Africa. The position of the boreholes that our group worked on are marked, including the deep KHK-1 borehole that was used by Reimold et al. (2002a) and Reimold and Koeberl (2003) to delimit the maximum extent of the Morokweng impact structure. (b) Aeromagnetic anomalies in the region of the Morokweng impact structure (modified after Henkel et al., 2002; data kindly provided for that publication by Geodass (Pty.) Ltd., now Fugro Airborne Surveys of Johannesburg). Note the disruption of the NE–SW trending regional dike swarm over a distance of ca. 80 km in the area of the impact structure. Also shown is the location of borehole KHK-1, with very limited evidence of deformation that could be unequivocally related to the impact event, which also constrains the maximum size of the impact structure (Reimold et al., 2002a). (c and d) Two examples of typical groundmass of Morokweng impact melt rock. The lithology is made up of finest-grained, often graphic intergrowths of quartz and feldspar (both alkali feldspar and plagioclase), between larger plagioclase laths and prismatic to lath-like pyroxene. Both images taken with crossed polarizers. (e) In contrast to the previous two images, here a considerably finer-grained variety of Morokweng Granophyre is shown, with spherulitic growths of sometimes skeletal orthopyroxene crystals emanating from clasts. Plane polarized light. The two textural varieties shown here closely correspond to the textures also observed in Vredefort Granophyre. (f) Plagioclase in basement granite intersected by drill core from the inner part of the structure. It shows planar deformation features (PDFs) in several directions (north-northwest–south-southeast, central part of the image; east–west, middle left; and parallel to the northwest–southeast trending twin plane), as well as subplanar fracturing parallel to the twin plane. Also note that predominantly one of two twin individuals has been annealed (best seen in the lower left corner).

confirmed that the Morokweng melt rock contains a meteoritic component. Much of the melt rock has a granophyric texture and resembles the impact melt rock from the Vredefort structure, the Vredefort Granophyre (see below). Interestingly, the Morokweng Granophyre is also directly comparable with the Vredefort Granophyre in terms of their major element compositions. Both target areas have a similar stratigraphy of crystalline basement and supracrustals – but it must be considered quite fortuitous that the rock mixtures that constituted the respective impact melts turned out to be nearly identical in major element composition.

Both Hart et al. (1997) and Koeberl et al. (1997a) determined the age of the Morokweng Granophyre by U–Pb single zircon dating. Their combined data sets constrain this age to 145 ± 2 Ma. Interestingly, this age is identical, within error limits, to the age of the Jurassic–Cretaceous Boundary that is associated with a significant, though not as prominent as the Cretaceous–Tertiary Boundary, mass extinction. Koeberl and Reimold (2003) reported on detailed petrographic and geochemical studies of the melt body at Morokweng. McDonald et al. (2006) investigated samples from the Jurassic–Cretaceous Boundary in southern England and northern France, but failed to find any evidence for a link between the extinction event and impact – neither in chemical results nor in the form of shock deformation.

An important feature of the Morokweng structure is the very strong chemical signature in the impact melt rock, which indicated to Koeberl et al. (1997a) and Koeberl and Reimold (2003) – from siderophile element abundance and Re–Os isotope data – that between 2% and 5% of meteoritic component could be determined in the impact melt rock. Koeberl et al. (2002) used Os and Cr isotopic data to constrain the impactor type to an ordinary chondrite. The finding of abundant PGEs has been taken by some to indicate that the Morokweng melt rock could be as important for its siderophile

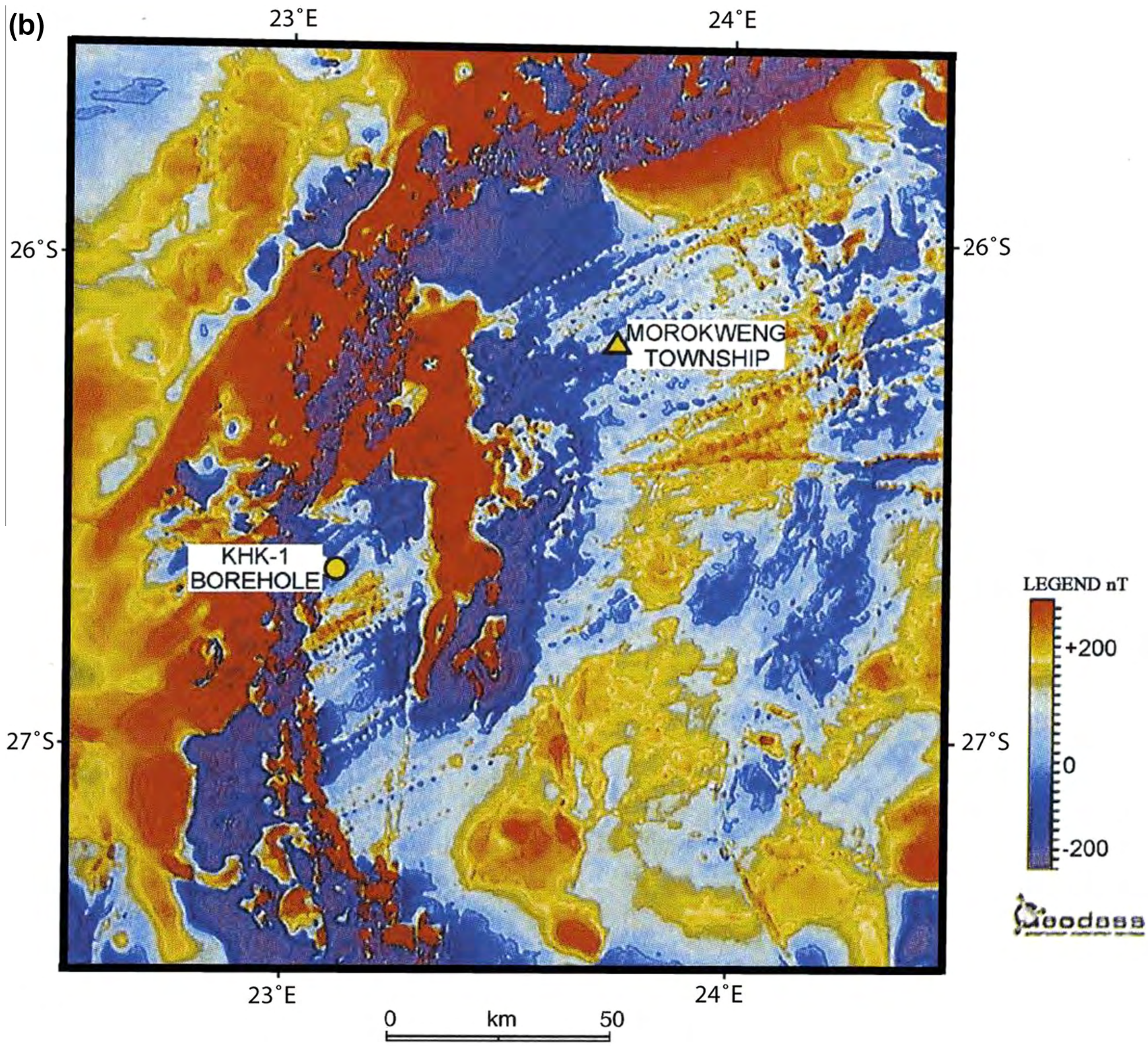


Fig. 29 (*continued*)

and possibly PGE content as the Sudbury impact melt complex (the so-called Sudbury Igneous Complex), but simple mass balance calculations have shown that the entire siderophile element content of the Morokweng melt rock could be accounted for by the – in terms of absolute mass – small contribution from the meteoritic projectile. And yet, the idea of "a second Sudbury" has persisted, especially, after Maier et al. (2006) established that a new borehole (M3) had intersected an 800 m thick section of melt rock. It has, however, been ignored that the older boreholes did only contain melt rock intersections of about 150, 125, and 95 (though this drilling was terminated still in melt rock) m thickness (Reimold et al., 1999a). Notably these three older boreholes were not drilled marginally to the melt complex. These observations indicate that the thickness of the melt rock complex in the inner part of the Morokweng impact structure is seemingly highly varied; an explanation for this is not readily available but one can speculate that this is the result of the collapse of the central uplift that this large impact structure undoubtedly would have had in its interior, and the remnant of which is represented by the Ganyesa Dome.

And yet, the Morokweng melt rock is indeed unique. Maier et al. (2006) discovered a 25 cm meteorite clast in impact melt rock of a drill core, an extraordinarily rare find. This work also established beyond doubt the type of the meteoritic impactor – namely an LL chondritic signature (also McDonald et al., 2001). Further details on the LL chondrite from Morokweng are reported by Jourdan et al. (2010).

Results of geophysical modelling of the Morokweng impact structure by Henkel et al. (2002) are consistent with the conclusions from geological observations on drill core from borehole KHK-1 (compare Fig. 29a) ca. 35 km to the southwest of the center of the Ganyesa Dome by Reimold et al. (2002a) and Reimold and Koeberl (2003). These authors found very limited evidence of impact deformation in this drill core – limited to a single, 10 cm thick injection of impact breccia between horizontally stratified and essentially undeformed country rocks – and no shock metamorphic deformation. According to these findings, the impact structure can hardly have been any wider than 70–80 km. However, several groups (most recently, Andreoli et al., 2008 – ca. 260 km) have nevertheless continued to favor a much larger, up to 380 km diameter for the Morokweng structure, thereby neglecting several lines of evidence that are consistent with the much smaller diameter: (1) The melt body and the magnetic anomaly are no wider than

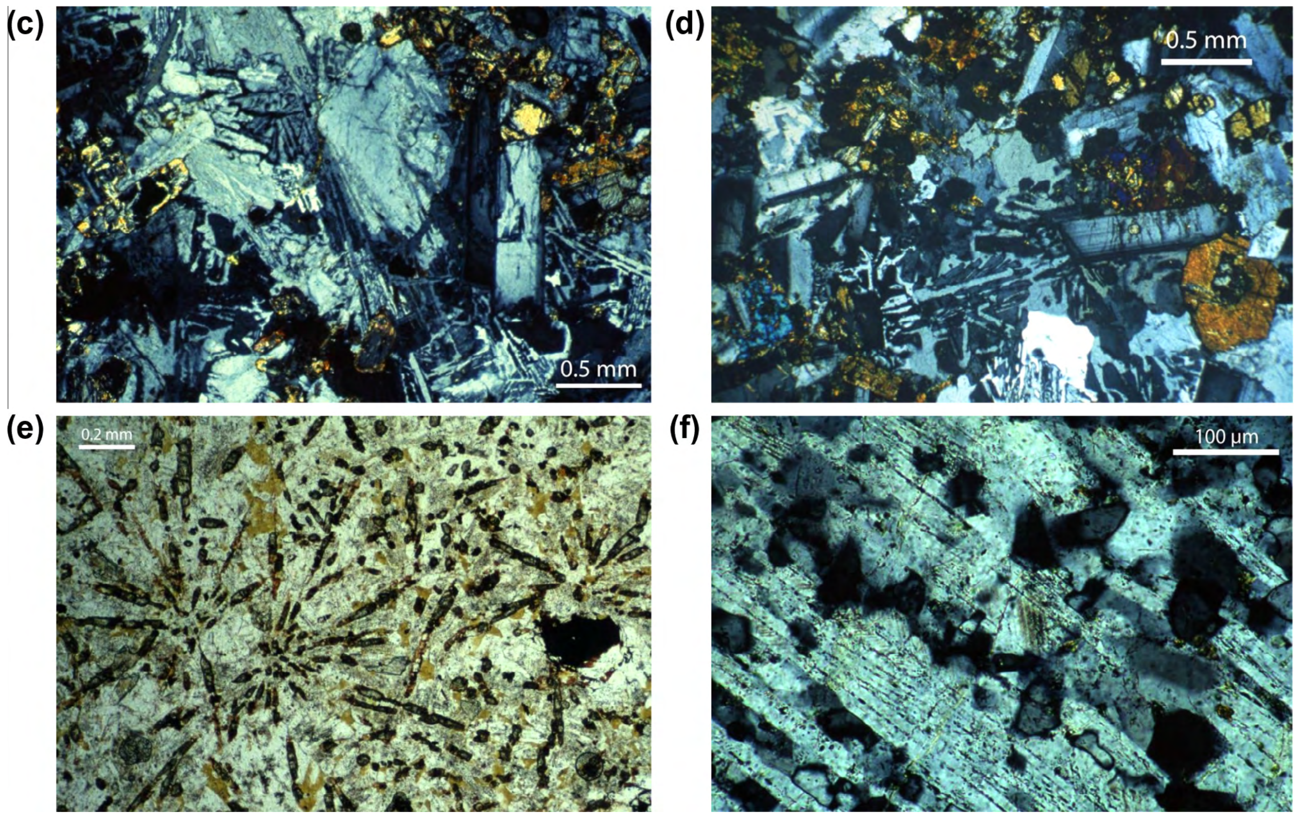


Fig. 29 (*continued*)

30 km, whereas an impact structure of 300–400 km width would have had significantly larger thermal and magnetic anomalies (see, for example, Ivanov, 2005: numerical modelling of the Sudbury and Vredefort impact structures, especially the inherent distribution of impact melt and associated thermal anomalies). (2) A very large Morokweng impact structure would have had a very large central uplift. Such a feature is not in evidence. (3) If one were to compare the 145 Ma old, several hundred kilometers wide Morokweng impact structure with the 250 km wide Vredefort Structure of 2.02 Ga age, one should expect a strong geological-structural expression of the Morokweng structure that would have been morphologically similar or even wider/deeper than the record at Vredefort. Instead, no central uplift with strongly up- or even overturned strata as observed at Vredefort is in evidence, and no crater rim. (4) One should expect to find widespread remnants of impact deformation around the Morokweng region, if the crater structure would have exceeded the current estimate of 70 km. And (5), a several hundred kilometer wide impact structure at Morokweng, that would have been even larger than the Chicxulub structure of K/Pg boundary age, would have had the potential to be responsible for a very strong environmental catastrophe at 145 Ma ago – which should be notable in the biostratigraphic record and should have accumulated chemical and shock deformation evidence at the J/K boundary – comparable to the so far unique findings at the global K/Pg boundary.

Bootsman et al. (1999) investigated the long-time evolution of the drainage system of the Molopo River. These authors compared the pre-impact and post-impact drainage patterns and concluded that after the impact both drainage direction and pattern were changed dramatically. They noted that prior to the impact, at ca. 300 Ma ago, the regional drainage direction was towards the northwest. After the impact event, at ca. 75 Ma ago, there was a generally southward flowing system. Since Tertiary times, i.e., since the filling of the Kalahari Basin, the affinity between the Molopo drainage and the Morokweng impact structure became less obvious. However, the curvature of the present bed of the Molopo River mirrors the northern outline of the Morokweng impact structure.

6.1.13. Oasis, Libya

The second (with *B.P.*, see above) Libyan impact structure that has long been related to the origin of the Libyan Desert Glass (below) is Oasis (Fig. 30). This structure is located at 24°35′N/24°24′E in Nubian sandstone Formation of 90–120 Ma age, some 120 km east-northeast of Al Kufrah oasis. The obvious features of Oasis, as noted on satellite and aerial photographs, are a 5.1 km diameter central ring of up to 100 m elevated hills, with predominantly outward dipping strata (French et al., 1974), which surrounds a more or less flat innermost depression. This central ring of strongly deformed, especially intensely folded, strata is surrounded by an also rather flat annular basin. This gives the general idea that Oasis is quite deeply eroded. Previously, various diameters have been quoted for this structure: 5.1–11.5 by French et al. (1974) and up to 18 km by Koeberl et al. (2005a). Oasis strongly resembles the 24 km wide, and equally strongly eroded, Gosses Bluff impact structure in Australia (Milton et al., 1996a,b). The Oasis structure is located about 85 km due south of the B.P. structure and occupies the same stratigraphic position in Nubia Sandstone Formation. As with B.P, the rather ill-defined age of this formation is the only age constraint for the Oasis impact as well.

French et al. (1974) reported multiple sets of 'planar elements' in quartz grains from orthoquartzite, as well as an alleged glass-bearing microbreccia found in the inner depression. They reported small interstices filled with brownish, partly devitrified glass containing sandstone fragments and shocked quartz.

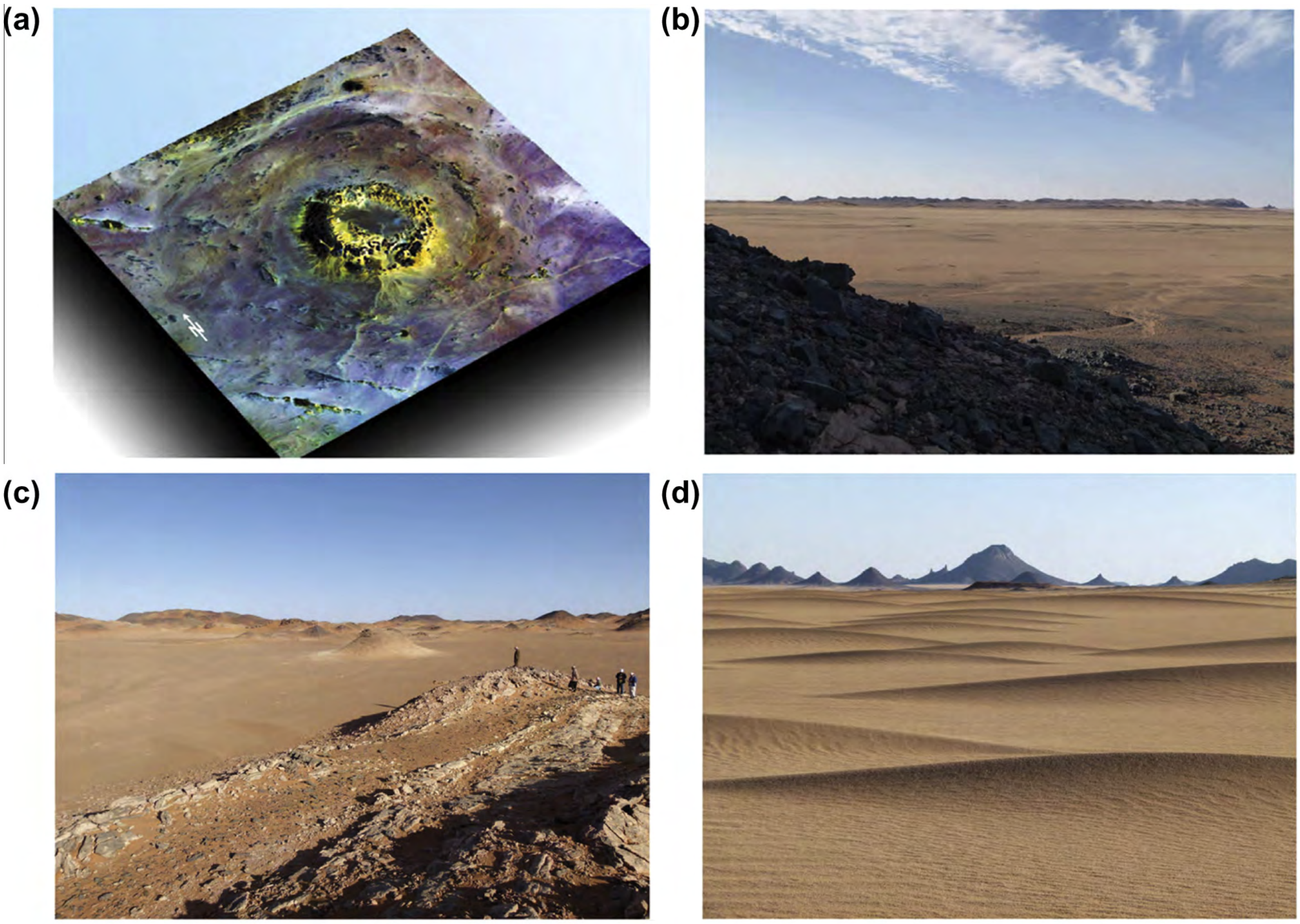


Fig. 30. Oasis impact structure (Libya). (a) ASTER satellite data combined with SRTM topographic data into a 3D impression of Oasis impact structure. The prominent inner ring of 6.5 km diameter and its immediate surroundings – together – are thought to represent the central uplift. Image courtesy of Alvaro Crósta, University of Campinas (Brazil). (b) View onto the ca. 6.5 km wide inner ring from the north-northwest where steeply inward dipping strata occur. The wide expanse in front of the inner ring has patchy outcrop that generally displays strong folding of the sedimentary strata. (c) A shallow exposure of generally subhorizontal sediments within the innermost depression of the inner ring. View towards the southern segment of the inner ring. (d) Coming from BP crater to the north of Oasis, this latter impact structure is first recognized by this ring of chimney-like features that represent the remnants of hematized (now weathered to limonite) sand dikes, the prominence of which above the other strata is due to the durable hardpan-like material that forms or coats this sediment. Field of view about 3 km wide. (e) Extensively folded, shallow outcrop about 18 km from the center of Oasis structure. If this deformation still represents within-crater geology, then the structure would be significantly larger than previously thought. (f–h) Shock deformation in quartz in sandstone and sandstone breccia (angular, well-separated clasts of quartz, cemented by goethite) from the innermost part of the structure. (f) A quartz grain in quartz breccia, with two sets of planar fractures (northwest–southeast, east-northeast–west-southwest, sometimes with feather feature-like arrays of short fractures coming off the principal planar fracture (e.g., just left of the image center). (g) A well-developed set of decorated PDFs in northeast–southwest orientation. Note the abundance of fluid inclusions in this grain. (h) Arrays of planar fractures in northwest–southeast direction, sometimes with shorter, slightly curved fractures coming off such planar fractures. In between planar fractures one finds densely spaced features that, at least in part, could represent PDFs.

In 2001 the authors of this review were permitted a rather short (less than 3 h) visit to Oasis – allegedly because the travel permits were not in order. Any attempt to resolve this issue with the authorities at Al Kufrah oasis remained unsuccessful. What is more, by far most of the samples, all composed of sandstone, collected on this short visit to the structure were confiscated at the Libyan–Egyptian border – as they were wrongly identified as meteorites by customs personnel, and the few specimens of sandstone that could be exported did not show any satisfactory evidence of characteristic shock deformation. A limited remote sensing study and some findings made during this very short visit to the crater structure in 2001 were interpreted by Koeberl et al. (2005a). They considered as the upper size limit of the structure the apparent (on satellite imagery) truncation of arcuate structural features centered on the structure by a series of northwest-trending regional lineaments.

On invitation by the Libyan Centre for Remote Sensing and Space Sciences, R.L. Gibson (University of the Witwatersrand, Johannesburg) and one of us (W.U. R.) were able to spend nine days at Oasis in October/November 2010, which allowed to map parts of the structure in some detail (Gibson et al., 2011a,b). Findings of *Lepidodendron* and trace fossils in some strata of the inner ring revealed that Oasis is not only formed in Nubia Formation sandstone and conglomerate, but also involves sandstones, siltstones and claystones of Upper Carboniferous age. In the inner ring and interior depression, some breccia occurrences were noted. The limited exposures did, however, not allow investigating contact relationships to adjacent rocks. It appears that the concentric topographic variation is a reflection of the variable resistances of the underlying lithologies to the impact-induced deformation: the prominent hills of the inner ring are capped by highly resistant, siliceous sandstones of the Lower Cretaceous, whereas the surrounding flats are underlain by less resistant Carboniferous beds. A distinct quartz-pebble conglomerate horizon that allegedly marks the base of the Carboniferous occurs widespread throughout the structure.

Gibson et al. (2011a,b) also found that the rocks of the central parts of the structure are strongly kaolinitized and rich in iron and manganese nodules, with local iron impregnation and chert

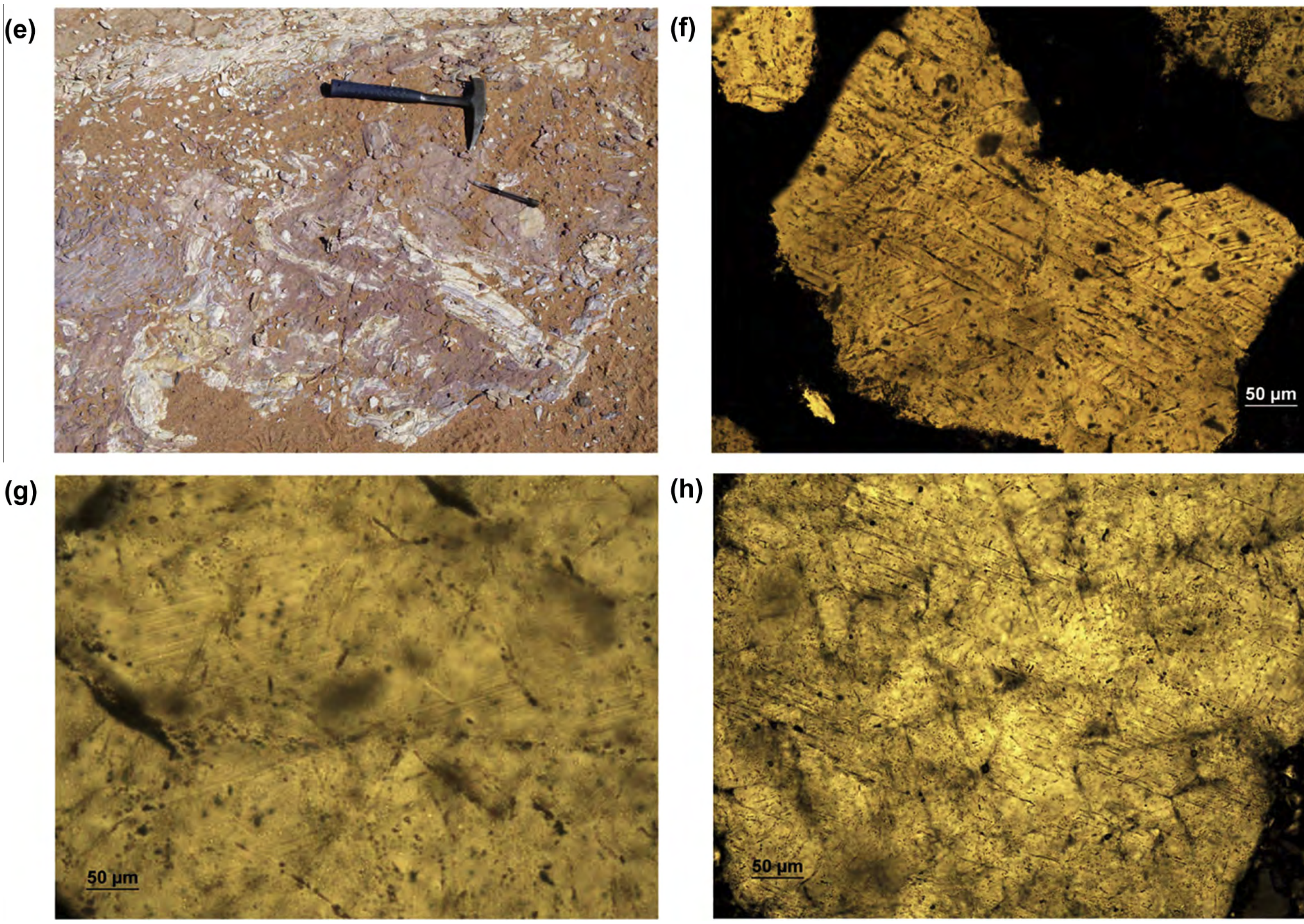


Fig. 30 (*continued*)

formation. They suggested that these effects post-date much of the impact-related features and could represent the overprint from an impact-related hydrothermal system. However, iron and manganese impregnation is also prominent in the outer reaches of the structure and, thus, could be – at least, in part – of pre- or post-impact age. While the regional geology is characterized by horizontal to gently warped strata, Oasis is pervasively and intensely folded and faulted, and strata locally have up- to overturned attitudes. This is particularly evidenced in the innermost ring structure. Tangential folds extend to at least 12.5 km from the center, and small-scale (decimeter to centimer scale) folding could be observed in the subcrop of the northeastern sector of the outer structure as far as 18 km from the center (compare Fig. 30e), which might indicate that the impact structure is significantly larger than previously thought. This macroscopic to mesoscopic deformation of the rocks exposed in the environs of the prominent inner ring does diminish in intensity with radial distance A recent evaluation of Radarsat and ALOS/PALSAR data by A. Crósta (U. Campinas, pers. comm.) suggested that larger fold structures may occur in this area as well. Still further from the structure no more folding has been observed at all, with essentially all exposures showing (sub)horizontal attitudes of basically undeformed rocks.

No impact-diagnostic features such as shatter cones were detected, and it is thought that this could be because the current erosion level represents a relatively deep section through the impact structure. To date, a large number of samples recovered at Oasis in 2010 have been studied petrographically, and up to 4 sets of PDFs, besides abundant planar fractures, have been observed in quartz of several breccia samples from the innermost part of the structure (Fig. 30f–h). This confirms the initial report of shock deformation by French et al. (1974). Shock pressure is estimated for these new samples to range from 5 to >15 GPa.

Based on personal communication by M. Baegi (Libyan Centre for Remote Sensing and Space Sciences, Tripoli, 2011), it is suggested that the impact occurred prior to formation of northwest-trending, normal faults thought to be of Tertiary (ca. 39 Ma) age. This would make the impact event significantly older than the apparently unrelated Libyan Desert Glass (see below) of 29 Ma age. Remote sensing analysis and field evidence were interpreted by Gibson et al. (2011a,b) to suggest a diameter of the Oasis impact structure of possibly as large as 36 km.

The particular style of folding-dominated deformation in the outer parts of the central uplift and its wider environs has stimulated discussion amongst the members of the 2010 field party whether this could be a result of impact into an unconsolidated sedimentary target – or due to the different material responses to shock within a target made up of strata of very different competencies.

6.1.14. Ouarkziz, Algeria

Ouarkziz (Fig. 31), also known as Tindouf, is a 3.5 km wide, severely eroded structure at 29°00′N/07°33′W. The structure is superposed onto a regional, NW–SE trending fault structure. According to Fabre et al. (1970), the crater structure is located in Carboniferous limestones and shales of Upper Viséen and Lower Namurian age, which sets an upper age limit for the impact event. Strata are upturned at the crater rim, where they have also been observed to be faulted and folded. The interior of the crater is largely covered by alluvium. Breccia presence and "planar features" in

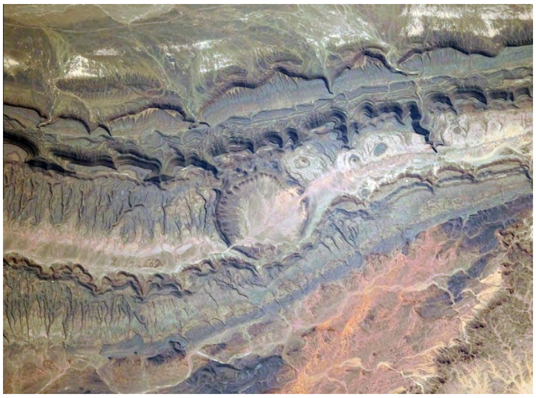


Fig. 31. Ouarkziz impact structure, Algeria, with a diameter of about 3.5 km (NASA ISS image).

quartz in samples from the inner crater rim were reported by these authors.

Lambert and Lamali (2009) remarked that it is necessary to verify whether these "planar features" constitute bona fide evidence of impact. They also distinguished three different, concentric zones in the structure, an outer zone with inward dipping concentric faults, followed towards the center by an inward dipping zone, and in the central part by a circular uplift feature with vertically dipping strata. They concluded that the impact occurred after formation of the Mesozoic regional peneplain and, thus, was of pre-Tertiary age. It is noted that since the last review by Koeberl in 1994, no new findings have been added to this initial account. Ouarkziz would make a useful target for further geological exploration. Currently listed as a confirmed impact structure – ever since the initial report by Fabre et al. (1970), it is nevertheless required to confirm the report of shock metamorphism through further field work, sampling, and petrographic studies.

6.1.15. Roter Kamm, Namibia

Roter Kamm (Fig. 32a) is a nearly circular, 2.5 km diameter, impact crater located at 27°46′S/16°18′E, about 80 km north of Oranjemund, in 1200 Ma granitic gneiss of the Namaqua Metamorphic Complex (NMC) of the southern Namib Desert (Reimold and Miller, 1989; Miller, 2008a, 2010). The interior of the crater is completely covered with eolian sands that presumably lie on top of an impact breccia fill. The first to propose an impact origin for this structure were Dietz (1965) and Fudali (1973). This was confirmed in the late 1980s after first detailed geological investigation of the crater and first findings of shock metamorphic effects in clasts of rare impact melt breccia by Reimold and Miller (1989).

The crater was formed in a three-layer target (Miller, 2008a, 2010). The topmost layer is formed by the two Namib ergs. There is the semi-consolidated Tsondab Formation erg at the base, which is exposed in the wind scoop just to the north of the crater (see Fig. 32b and c) and contains fragments of paleo-ostrich egg shells in these exposures. The uppermost facies is the unconsolidated Sossus Sand Formation on which the ejecta northwest of the crater landed. This ejecta apron is cemented by pedogenic calcrete, which has stabilized the dune surface for several millennia and possibly for as much as one million years. At the edge of the wind scoop, one can dig into unconsolidated Eileen sand below the ejecta-bearing calcrete (R.McG. Miller, Windhoek, pers. comm.). The underlying bedrock is composed of Gariep Group metasediments (marble, schist, minor quartzite and sandstone) overlying the granitoids of the NMC basement. Miller (pers. comm.) also notes that only the Tsondab Formation is fossil-bearing at the crater. Its age extends from 21 Ma to 5 Ma (Miocene) (as constrained, inter alia, by various Memoirs of the Geological Survey of Namibia). The Sossus Formation is not fossiliferous at Roter Kamm, but it is elsewhere; its age has been delimited to range from 5 Ma to the present.

The first geophysical analysis of the Roter Kamm structure dates back to the work by Fudali (1973), who collected gravity data and suggested a model of eolian sands overlying impact breccia to a maximum depth of about 800 m. Miller (2008a, 2010) reported results of an aeromagnetic investigation and concluded "that the aeromagnetic signature is weak but discernible over the structure. The magnetic features outside the rim are covered by eolian sand and their cause is unknown". Brandt et al. (1998) reported ground-magnetic data along traverses across the structure. They did not identify an anomaly over the crater interior but Brandt et al. ascribed a slight positive magnetic anomaly to a possible interface between the crater floor and the crater-fill breccia. Magnetization was also investigated by Rajmon et al. (2005), who could, however, not find any evidence of impact-related remagnetization and linked natural remanent magnetic components of crater rocks to Proterozoic events in the regional geological history. With suevite and impact melt rock finds recorded, this leads to the question how much melt and, thus, thermal energy was produced upon impact to affect the crater fill or crater area to allow pervasive remagnetization. Brandt et al. (1998) also discussed new gravity data and interpreted a small negative and symmetrical anomaly over the crater as due to the presence of low-density impact breccia and sediments in the bowl-shaped crater.

Grant et al. (1997) applied ground-penetrating radar to investigate the possibility to use this technique to map out ejecta distribution beneath eolian sand cover. They were, thus, capable of obtaining local signals that could be related to presence of ejecta below a relatively thin sand cover. They also concluded that the crater had been degraded by several tens of meters. Miller (2008a, 2010) employed airborne radiometrics to investigate the possible distribution of ejecta around the crater, and then followed up on this with detailed mapping on the ground, both on the crater rim and in its environs (Fig. 32b–d).

Major conclusions based on the earlier, but particularly on Miller's, mapping results include: (1) Along the rim, the foliation patterns are quite varied – which forces the conclusion that block rotation has affected the walls of the crater. (2) Reimold and Miller (1989) and Miller (2010) showed that the chemical composition of the pre-impact target was highly heterogeneous. (3) Cataclasites are invariably K-enriched compared to the host gneisses. (4) Significant remnants of the ejecta apron could be mapped by Miller (2008a, 2010) outside of the crater. Ejecta were deposited on top of the earliest, fossil-bearing, eolian sands of the Pliocene to Holocene Sossus Sand Formation. This placed a new age constraint on this impact event. Outside the crater, ejecta remnants are most abundant in an "outward-fanning apron" in the north-northwest to west sector around the crater, where concentric and radial swaths of ejected blocks were also detected. (5) Miller (2010) concluded from his ejecta mapping and considering a slight asymmetry of the crater that the projectile for this impact event could have come from a southeasterly direction.

Much discussion has centered on the nature of breccias at Roter Kamm (Reimold and Miller, 1989; Reimold et al., 1994a; Degenhardt et al., 1994), with the conclusion that the majority of breccias found in situ along the rim crest are cataclasites without shock deformation. Early in the investigation of the geology of Roter Kamm, the presence of pseudotachylitic breccias had been suggested, but this was subsequently rejected. The radial trends of the major breccia zones in evidence on the crater rim indicate clearly that these cataclasite formations are related to the impact event. Reimold et al. (1997) detected a patch of suevite on the northwestern crater rim, upon a chance visit to the crater. They observed that this deposit was likely exhumed because of a temporary change in the prevailing wind direction. What is more, in

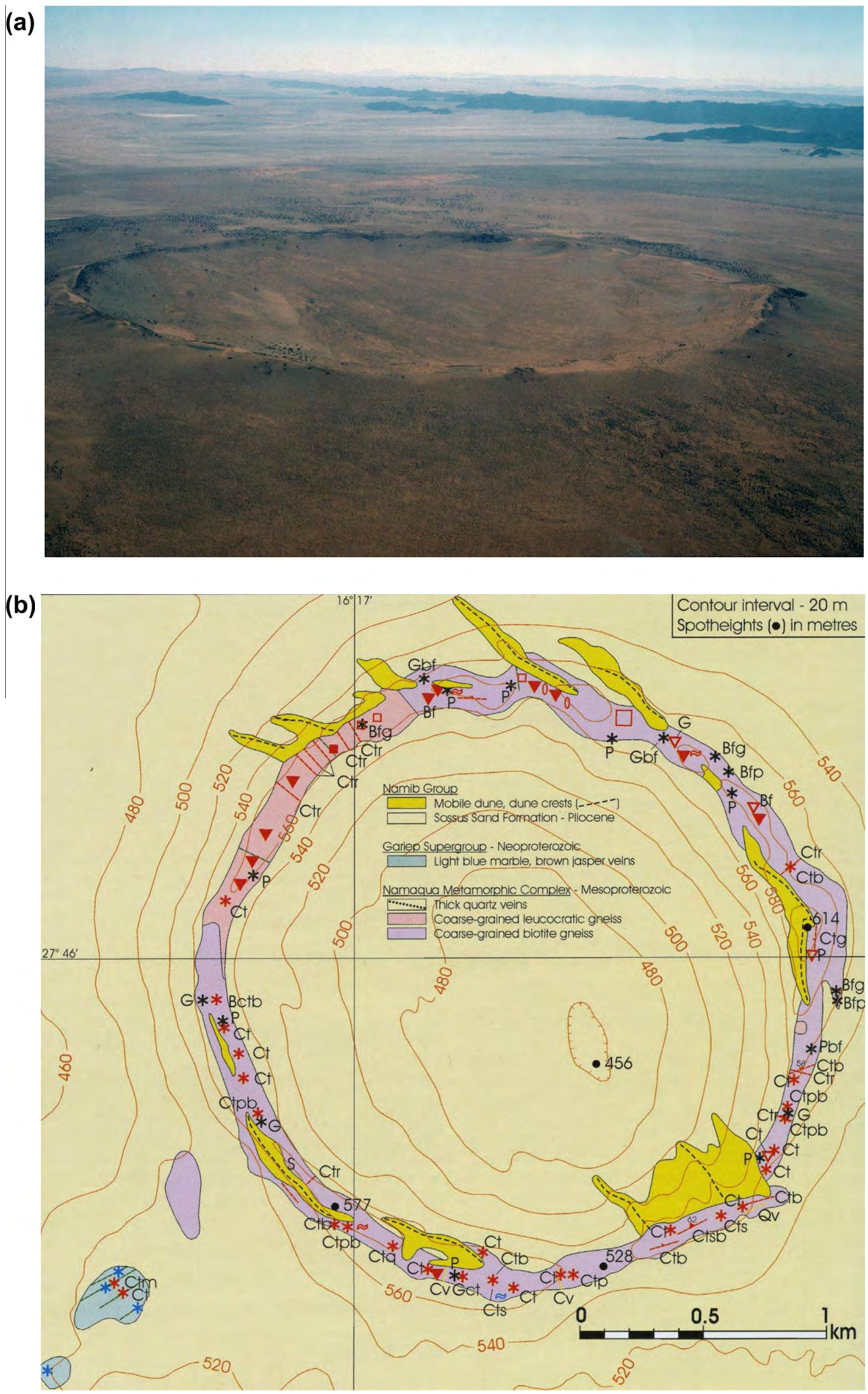


Fig. 32. Roter Kamm impact crater, Namiba. (a) Aerial photograph of the Roter Kamm impact crater, Namibia, with a diameter of about 2.5 km. The structure is almost totally covered by sand dunes (Image: C. Koeberl). (b) Geological map of the Roter Kamm crater and detailed legend (c). (d) Geological map of the area around Roter Kamm crater with ejecta remnants as mapped by Miller (2008a,b). Figures (b)–(d) are only slightly modified versions of original graphics by Miller (2008a,b), reproduced with permission by R.McG. Miller and after copyright release from the Geological Survey of Namibia (Director Dr. G. Schneider).

(c)

Fracturing, no cataclasite in fractures

Radial joint

Qv Quartz vein

Cv Carbonate vein, 2-5 cm wide

S Shear zone with slicken-side surfaces

P * Fractured/brecciated pegmatite

G * Fractured/brecciated gneiss

Gbf * Fractured/brecciated gneiss with blackened feldspar

Pbf * Fractured/brecciated pegmatite with blackened feldspar

M * Brecciated marble in vertical fracture zone

No fracturing or cataclasite veins

Bfg * Gneiss with blackened feldspar

Bfp * Pegmatite with blackened feldspar

Black cataclasite veins without associated brecciation of country rock

Ct * Black cataclasite veins, single or zones of multiple veins, single veins <1 mm wide

Cts * Black, green or reddish cataclasite veins with slicken-side surfaces, veins 5 mm to 5 cm wide

Ctp * Thin black cataclasite veins in pegmatite

Gct * Green cataclasite veins, 1-8 cm wide

Ctq * Black cataclasite veins in 40 x 40 m quartzite xenolith

Cataclasite veins associated with moderate brecciation of country rock

Ctb * Black cataclasite veins, variable penetration of cataclasite into adjoining brecciated gneiss

Ctsb * As for Pts

Ctpb * Cataclasite veins in pegmatite, zones up to 15 cm wide, variable penetration of a fine network of cataclasite veins into the adjoining brecciated pegmatite

Bctb * Light brown cataclasite veins in brecciated gneiss

Ctm * Grey cataclasite veins, 1-15 mm wide, in radially orientated fracture zones, some of which contain brecciated marble

Cataclasite veins associated with brecciated country rock

Ctr Radially orientated cataclasite veins, ±1 mm wide, zones of radially orientated cataclasite veins up to 20 cm wide, zones of fractured/brecciated gneiss, some thoroughly pervaded by black cataclasite, zones between 1 m and 50 m wide

Ctg * Intensely brecciated gneiss thoroughly pervaded by black cataclasite

Concentric cataclasite veins

Dip not indicated, ( ) shallow dip, ( ) steep dip, (58) measured dip, ( ) strike and dip of veins parallel to foliation in gneiss

Ejecta fragments and blocks on crater rim

▼ Blocky breccia of angular quartz fragments welded together by a light reddish cement of very fine-grained devitrified(?), crystalline quartz, blocks up to 1.50 m long

□ Large blocks of vein quartz up to 1.5 m Ø

□ Blocks of fragmented vein quartz cemented by thick veins of black, very fine-grained, devitrified(?), cataclasite, blocks up to 25 cm Ø, veins up to 4 cm wide

■ Strewn field of above quartz blocks with black, fine-grained, devitrified(?), cataclasite as well as rounded fragments of "flädle"-like, carbonaceous schist breccia

0 Subrounded fragments of pink, brecciated(?), fine-grained quartz-feldspar rock that appears to have suffered partial melting

G ▽ Subrounded pieces of brecciated gneiss pervaded by cataclasite veinlets, up to 50 cm Ø

P ▽ Subrounded pieces of brecciated pegmatite intensely pervaded by cataclasite veinlets, up to 20 cm Ø

Bf ▽ Subangular to subrounded pieces of gneiss or pegmatite with black feldspars

Ct ▽ Subrounded pieces of cataclasite

≈ Small fragments of graphitic schist, <5 cm Ø

≈ Blue, mustard-coloured marble, brown jasper

Fig. 32 (*continued*)

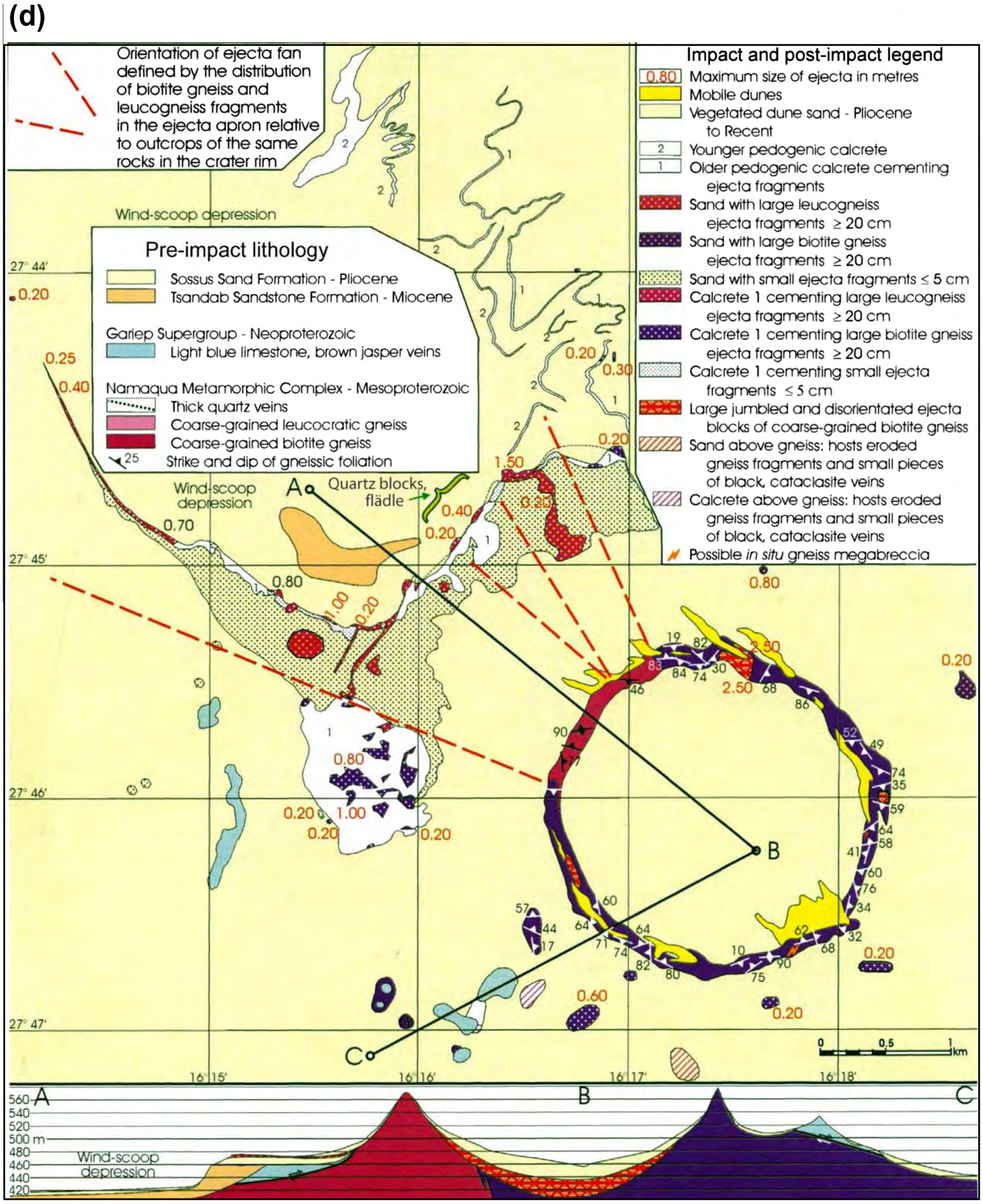


Fig. 32 (*continued*)

2004 a previously unknown patch of impact melt rock was found on the southwestern rim (Hecht et al., 2008). This shows that the prevailing wind direction is subject to periodic change, thereby uncovering previously inaccessible sections of the crater structure. Future visits to the crater may well have potential to obtain more information on this impact structure. Note that Roter Kamm is located in the restricted area of the southern Namib-Naukluft Reserve, and that special permits are required to visit this environmentally protected area.

Koeberl et al. (1989) suggested that unusual quartz pebbles occurring around the crater contained evidence for impact-related hydrothermal activity. The age of this impact event was determined by ^{40}Ar–^{39}Ar step-heating analysis of a melt breccia pebble by Koeberl et al. (1993) to 3.7 ± 0.3 Ma, in good agreement with the aforementioned biostratigraphic constraints. Hecht et al. (2008) reported new ^{40}Ar–^{39}Ar step-heating results for samples from a new locality of impact melt rock but obtained rather poorly constrained ages of 4.69 ± 0.31 Ma and 5.06 ± 0.44 Ma, so that currently the Roter Kamm impact age is only constrained to <5 Ma. Clearly the Roter Kamm impact melt rocks are characterized by compositional heterogeneity and disequilibrium on a centimeter scale, which represents problems frequently encountered with fast cooled melt rock. Hecht et al. (2008) also reported platinum group element abundance patterns and ratios for several melt rock samples that they interpreted to indicate an ordinary chondrite composition of the projectile for this impact.

6.1.16. Talemzane, Algeria

Talemzane, also known as Daïet El Maädna, or simply Maâdna, is a simple, bowl-shaped crater of 1.75 km diameter (Fig. 33) that is

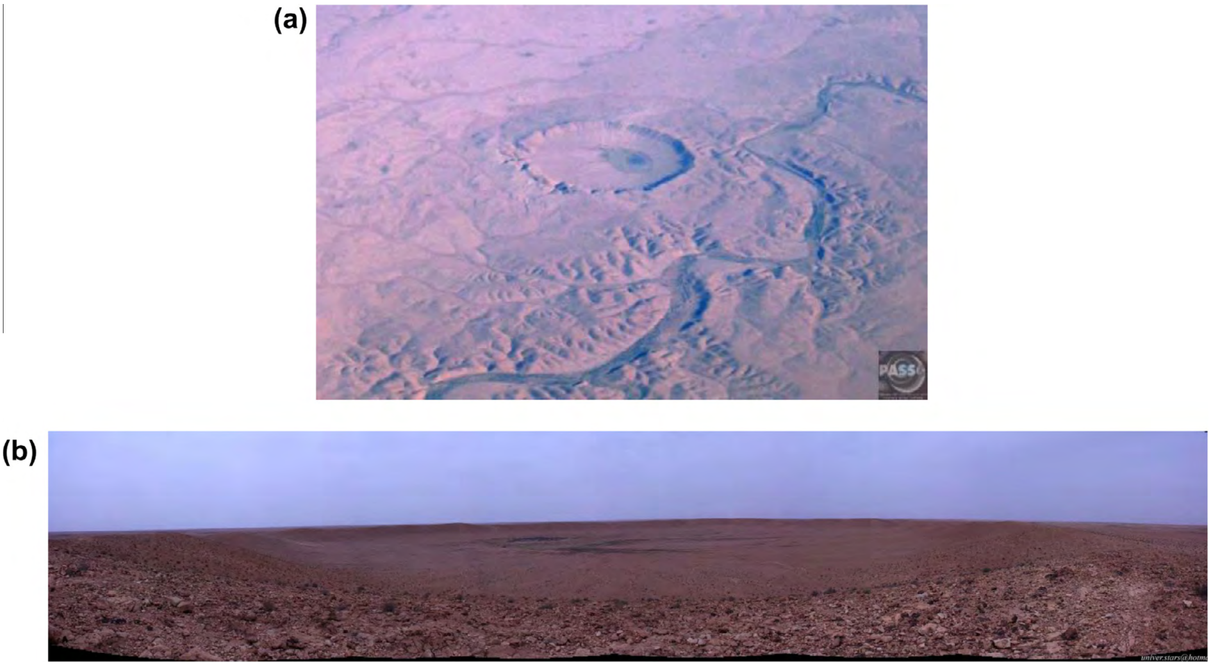


Fig. 33. Talemzane impact structure, Algeria. (a) Aerial photograph; image courtesy of the Earth Impact Database, Planetary and Space Science Centre, University of New Brunswick, Canada, from a photograph by Ross Richards. (b) Panorama of the crater. Image courtesy of the Earth Impact Database, Planetary and Space Science Centre, University of New Brunswick, Canada. Image Credit: Aissa Moussa Mohammed.

located at 33°19′N/04°02′E in Algeria. This structure was first visited in 1951 and then investigated from the air in the following year. Karpoff (1953) was the first to propose a meteorite impact origin for Talemzane. Lambert et al. (1980) obtained limited but definite evidence for impact origin in the form of a few quartz clasts found in the ejecta blanket surrounding the crater, which exhibited poorly developed planar deformation features. A mapping report by Sahoui and Belhai (2011) indicated different types of likely impact-generated breccias, including monomict and polymict lithic breccias, also occurring as breccia dikes. Sahoui et al. (2013) referred to a variety of impact melt-bearing breccias.

The crater occurs in Senonian or Eocene limestones and features an up to 70 m high rim. Limestones are strongly fractured, upturned, and – in the upper rim – overturned. Large, ejected blocks of limestone are scattered around the outside of the crater. Breccia dikes are intersected by the crater wall, and detrital or reworked monomict breccia is found at the crater floor near the rim. Quartz is a rare constituent in the limestones, but Lambert et al. (1980) reported diagnostic PDFs in some quartz grains. These authors estimated the age of the structure at <3 Ma because of the limited degree of erosion observed.

Lamali et al. (2009) reported results of a first detailed ground magnetic and susceptibility study along radial profiles across Maâdna. They obtained a complex pattern of positive magnetic anomalies over the interior of the crater structure, which they related to local accumulations of magnetite, possibly because of the presence of isolated bodies of impact melt (which may be a bold conclusion, as it is unlikely that the limestone target would have been amenable to large-scale melt formation). A distinct positive soil magnetic anomaly seems to superpose the most pronounced magnetic anomaly area. The cause of this susceptibility anomaly is not known but the coincidence with the area of a deep-seated magnetic anomaly suggested to the authors a common link to the impact (they may have called upon presence of impact melt once again).

Sahoui et al. (2013) reported evidence for carbonate melting (i.e., presence of immiscible impact-generated melts involving a $CaCO_3$ rich and a silicate-rich melt phase) as a consequence of the impact. It appears that further detailed studies of the breccia lithologies could be rewarding.

6.1.17. Tenoumer, Mauritania

The 1.9 km wide, almost circular crater Tenoumer (Fig. 34a) is located in the Western Sahara of Mauretania, at 22°55′30″N and 10°24′W, about 400 km northwest of the *Aouelloul* crater (above). The structure was excavated from a peneplained surface of Precambrian gneiss and granite that is covered by a thin veneer of young (possibly Pliocene) sediment. The present depth of the crater, measured from the top of the rim to the apparent crater floor, is about 100 m. The depression is filled with unconsolidated sediment. Based on geophysical data by Fudali and Cassidy (1972; also Grieve et al., 1989), the base of the post-impact sediments above the crater fill is estimated to occur at a depth of 200–300 m. The inner slopes are quite steep (Pratesi et al., 2005, refer to it being "locally abrupt"). In contrast, the outer rim slopes are only steep in their upper reaches, whereas below that they flatten out into the surrounding plain.

Earliest workers on Tenoumer crater favored a volcanic origin: Richard-Molard (1948a) referred to the presence of basalt lava and pumice and suggested a volcanic explosion. Allix (1951) was the first to propose an origin by impact but failed to present conclusive evidence for this. The occurrence of a swarm of small "dikes" supposedly comprising "rhyodacitic lava", intrusive into concentric fractures around the crater or outcropping just outside the crater rim, was interpreted by Monod and Pomerol (1966) as evidence for a possible volcanic origin of the structure.

Finally, however, French et al. (1970) reported up to 8 sets of PDFs in quartz grains from obviously strongly shocked inclusions of granite in the "lava", thus demonstrating that the crater structure is of impact origin and that the "lava" represents impact melt rock. It also contains lechatelierite and diaplectic quartz glass inclusions, as well as ballen quartz. French et al. (1970) reported some Rb–Sr isotope data that showed that the melt was indeed derived from the crystalline basement. Fudali (1974) remarked that

(a)
(b)
(c)

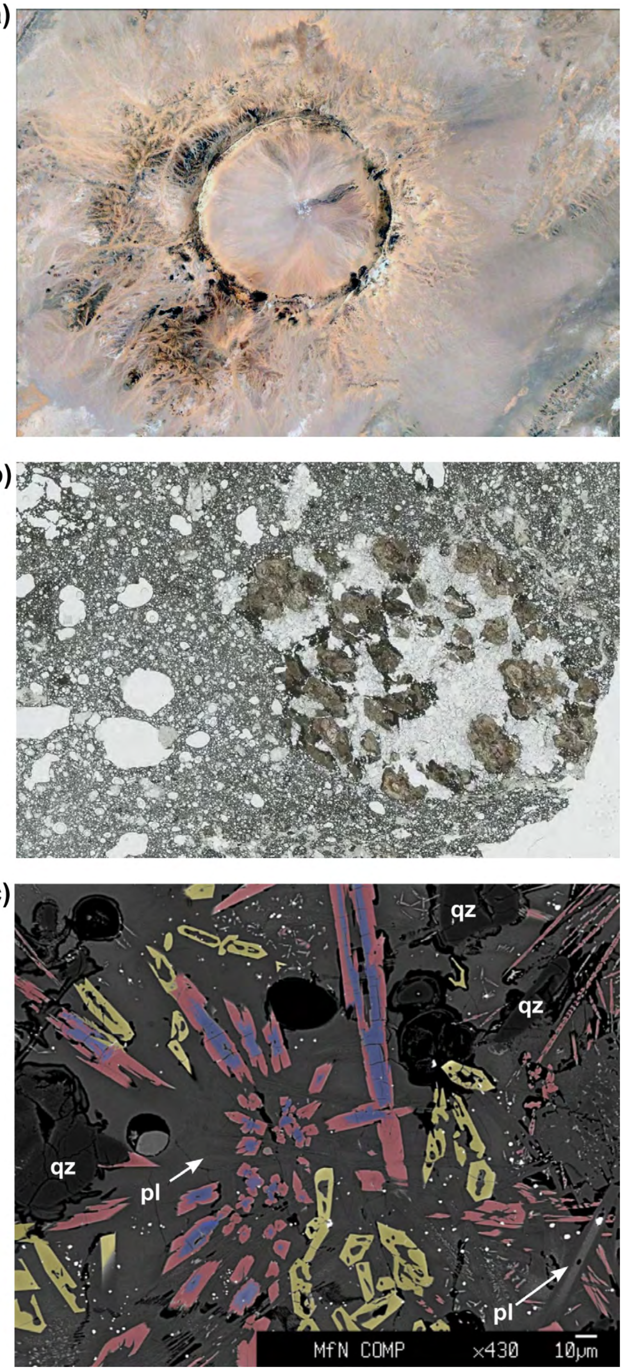


Fig. 34. Tenoumer impact crater, Mauritania. (a) Satellite image of the Tenoumer crater. (b) Thin section scan of impact melt rock (MfN Berlin sample Ten6) with a sizable clast of olivine gabbro in glassy groundmass. Width of field of view: 3 cm. (c) Backscattered-electron image (partially element color coded) of melt rock groundmass (MfN Berlin, sample Ten8). The glass matrix has a dacitic to rhyolitic composition; it contains olivine (yellow), pyroxene (blue = opx, purple = cpx), and plagioclase (pl) microphenocrysts. Some crystals show spinifex-like textures. One relatively larger plagioclase lath is significantly zoned (bottom right). White spots are Fe-oxide grains of <5 µm) size; qz = quartz clasts derived from the target. Scale bar = 10 µm.

the composition of the melt rock does not correspond to that of the gneisses and granites occurring in the regional basement but requires a component derived from amphibolite veins and enclaves found in the gneissic terrain.

In 2002, an Italian team of researchers visited the crater and collected extensive material for a comprehensive geochemical investigation, with particular emphasis on investigation of possible mixing relationships between the impact melt rock and various target lithologies, as well as a search for a possible meteoritic component. Pratesi et al. (2005) summarized these results.

Besides some detailed petrography of the impact melt rocks, these authors reported extensive chemical analyses, including major element, REE, and platinum group element (PGE) data for melt rock samples and various regionally occurring lithologies. Major element systematics for impact melt rocks and target lithologies show that besides felsic granites and gneisses, a mafic component must have been part of the target volume. Mixing calculations applying the HMX mixing calculation program of Stoeckelmann and Reimold (1989) reveal that a definite contribution from mafic sources is required. The impact melt rock composition is best reproduced by a mixture of 50% granitoids, 17–19% mica schist, 15% amphibolite, 10% cherty limestone, and 6% ultrabasite. No contribution from a meteoritic component could be detected. PGE abundances in melt rock are very low and can be accounted for by the ultrabasite contribution. Pratesi et al. (2005) also provided some petrological detail, including evidence for liquid immiscibility between silicate melt and spherules and globules of calcite.

Recent analyses of several Tenoumer melt rock samples (bomb and lapilli sized specimens) derived from the outer, northeastern crater rim were reported by Schultze et al. (2012) and Hecht et al. (2013). The range of whole rock chemical compositions is similar to that found in previous studies (Pratesi et al., 2005) and is clearly due to considerable variation in relative proportions of more siliceous rocks (mainly granitoids) and mafic rocks that constituted the target lithology. Some mafic clasts suggest that more or less metamorphosed olivine gabbro is also part of the target lithological composition (Fig. 34b). The impact melt samples are mainly of intermediate composition (andesite to basaltic andesite), but do show significant amounts of olivine microphenocrysts (Fo_{64-75}, Fa_{24-35}), ranging from 5% to 25% of the groundmass, exclusively in Mg- and Fe-rich samples.

Microtextures suggest that clinopyroxene formed after olivine and orthopyroxene (Fig. 34c). Furthermore, clinopyroxene formed contemporaneously with or prior to plagioclase. All textures are typical for fast cooling, as exemplified by atoll-shaped olivine, or acicular pyroxene and plagioclase (Fig. 34d) – in fact indicating extreme disequilibrium conditions upon crystallization. The heterogeneity of the Tenoumer melt rock samples has two main reasons. First, impact melting of target lithologies resulted in mixing of different target rock proportions, on a local scale. There was probably no coherent melt pool that would represent a homogeneous mixture of all target rocks. Second, melt rock heterogeneity occurs at the thin section scale and is due to fast cooling with disequilibrium crystallization conditions, during and after the dynamic ejection and emplacement of the melt bombs. Two chemically distinct melt phases, a Ca,Fe-rich one and a Si,K-rich one, were analyzed in the interstitial glass matrix. Similar observations were made by Hecht et al. (2013) and Hamann et al. (2013) on Wabar impact melt samples. Hecht et al. (2013) concluded, with regard to Tenoumer, that this phase separation into co-existing liquids could occur due to mixing of different target rock melts, or – as at Tenoumer – due to rapid crystallization.

The age of the structure was initially determined to 2.5 ± 0.5 Ma from K–Ar dating of melt rock (French et al., 1970). This age, however, was questioned by Storzer et al. (2003) as possibly being too high, as there might have been inherited Ar in the samples analyzed by French et al. (1970). By the fission-track method, Storzer et al. (2003) obtained a very young age of 21.4 ± 9.7 ka. Obviously such a young age for Tenoumer would preclude the previous suggestion (Dietz et al., 1969; Fudali and Cressy, 1976) that the

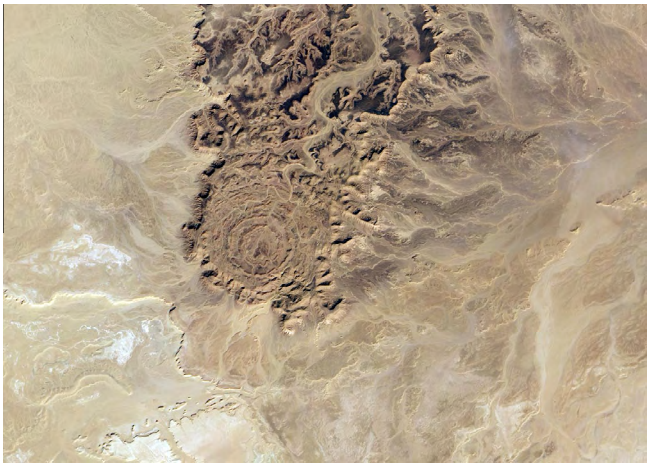


Fig. 35. Satellite image of the Tin Bider impact structure, Algeria (NASA EO-1 satellite image).

Aouelloul crater dated at 3.1 ± 0.3 Ma and Tenoumer could be of coeval origin and possibly linked by impact of two parts of a single bolide.

The age controversy about Tenoumer is not finished yet: An age of 1.52 ± 0.14 Ma was obtained recently by ^{40}Ar–^{39}Ar step-heating analysis in the Argon Chronology Facility at Curtin University. It is based on a weighted mean of three concordant inverse isochron ages obtained on splits from two melt rock samples (F. Jourdan [Curtin University, Perth] and W.U. Reimold, unpublished data).

6.1.18. Tin Bider, Algeria

Tin Bider (Fig. 35), also known as Tademaït, is an about 6 km wide, conspicuous structure located at 27°36′N/05°07′E. It is made up of a series of annular ridges spaced at 2, 3.5 and 6 km from the center. Tin Bider occurs in Lower to Upper Cretaceous clay and limestone formations, with the different susceptibilities of these strata to erosion being the cause of the prominent ridge-and-valley structure. Lower Cretaceous sandstones are exposed in the central part of the structure and seemingly have been uplifted by some 500 m from their original stratigraphic position. The concentric ridges exhibit highly deformed strata with complex and intense folding, likely facilitated by the different rheologies of the target rocks.

No macroscopic evidence (shatter cones) of impact has been reported to date (Lambert et al., 1981), and breccia occurrences have not been reported yet either. However, quartz grains with up to 7 sets of decorated planar deformation features (PDFs) have been reported by these authors from samples of the uplifted, massive sandstone of the central "eye" of Tin Bider, providing definite evidence of impact. The structure seems to be quite deeply eroded, as the entire crater fill has apparently been removed. The age of this impact structure is only constrained by the minimum age of the target rocks, i.e., the impact must have occurred at Lower Cretaceous time or later. Tin Bider is an obvious target for further detailed geological analysis that might provide new insights into the processes related to impact into a target composed of sedimentary rocks of very different behavior under shock compression.

6.1.19. Tswaing, South Africa

A small, only 1.13 km diameter meteorite impact crater is located at the northeastern edge of Gauteng Province, at 25°24′S/28°05′E – just about 40 km north–northeast of the city of Pretoria (Tswane) in north-central South Africa (Fig. 36). The crater structure has in the past been known variably as Zoutpan, Soutpan (Afrikaans), Salt Pan, or Pretoria Saltpan, but its official name is now Tswaing (Sotho, for "Place of Salt") – with the earlier names still in evidence on some regional road signage. The crater is very well defined, with a prominent rim that stands high (about 60 m) above the surrounding terrain, and even higher (90 m) above the crater interior. Tswaing is located in Nebo Granite, part of the felsic phase of the Bushveld Complex (Cawthorn et al., 2006).

The origin of this crater structure has been controversial ever since the beginning of the previous century. Wagner (1922) proposed a volcanic origin on the basis of observation of volcanic intrusives in the crater wall (Fig. 36b and c), and being a well-renowned geologist of his day, his opinion carried weight for decades. In contrast, Rohleder in 1933 was the first to suggest a meteorite impact origin, mainly on the grounds of crater morphological observations. Although this publication appeared in a prominent scientific journal, it was largely ignored by the South African geological community. Several short drill cores were retrieved from the crater interior in the 1950s and 1970s, mostly for economic reasons (to delineate the trona resources below the lake level but also to obtain possibly evidence to resolve the controversy about the crater origin). These efforts did remain inconclusive with regard to the latter problem.

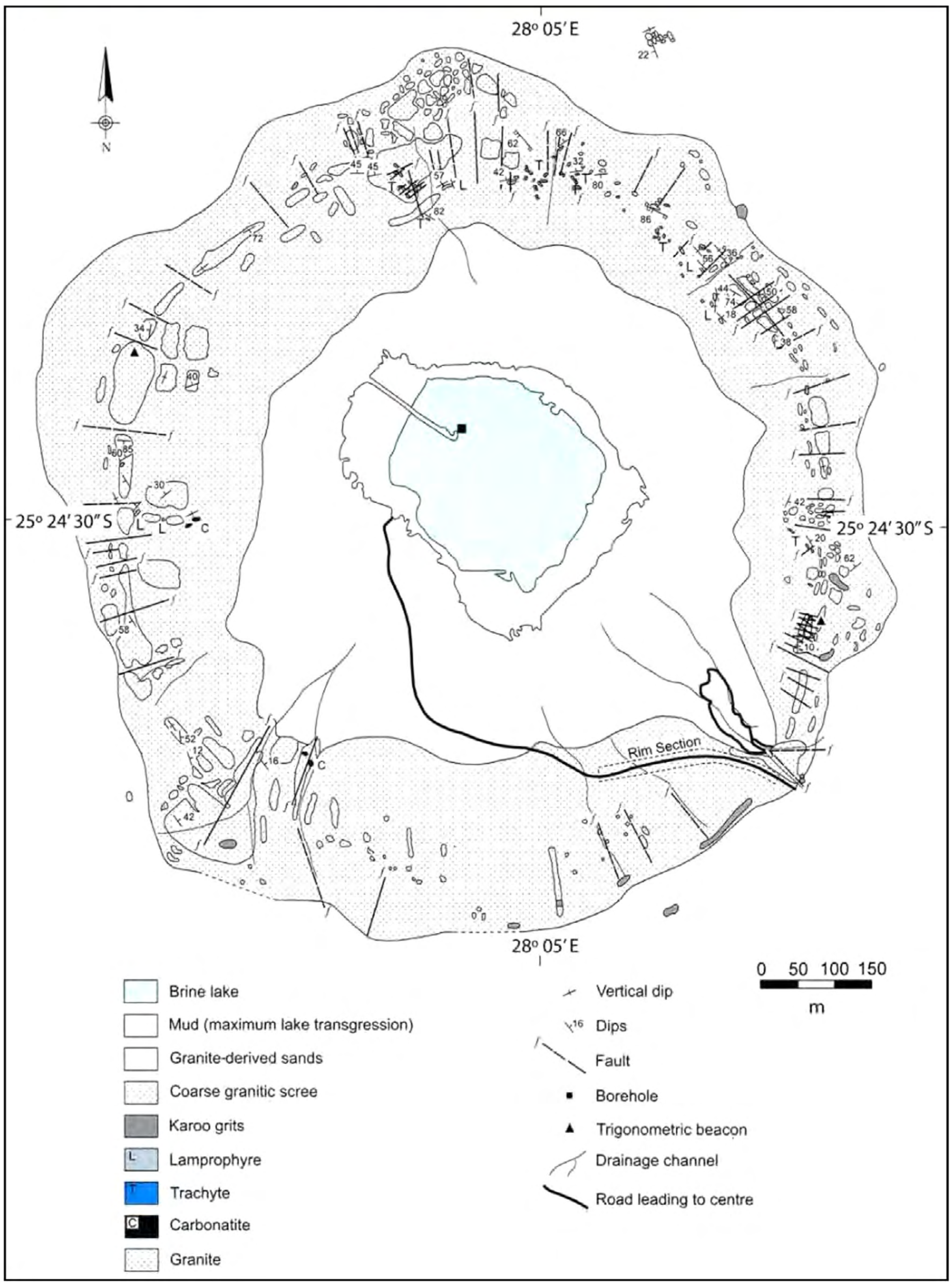
A detailed geological study of the crater and its environs was carried out by Brandt (1994; see also Brandt and Reimold, 1995a,b). Along a dirt-track leading from the shoreline of the crater lake to the top of the southeastern crater rim, a number of outcrops elucidate the lithologies and structure of the rim section (Brandt and Reimold, 1995a; Reimold et al., 1999a,b). Regarding the structural section through the rim (Fig. 36d), significant are – from the bottom upward – steep, inward-dipping faults, followed by prominent anticlinal structures in the midsection, in turn followed by a rim section characterized by low-angle, inward-dipping faults, and then by the overturned stratigraphy of the upper section (Nebo granite folded over younger Karoo grits). Finally, all this is overlain by ejecta comprising less than a decimeter to meter sized, frequently angular blocks of granite. A similar sequence of structural elements was described by Reimold et al. (1998a,b) from the northern crater rim of the complex Bosumtwi impact structure of Ghana.

The Tswaing crater contains a sizable crater lake, which has been the subject of considerable environmental analytical work (see below). The crater itself was a closed basin ever since its formation and, thus, it was thought that the crater fill could have preserved a long paleoclimatic record. For this purpose and hoping that it might also procure some new insight into the origin of the

Fig. 36. (a) Panoramic view across 1.13 km wide Tswaing crater, from the southern view-point towards the north-northwest. Photograph by Hans Knöfler (MfN Berlin). (b) Geological map for the Tswaing crater area, modified after Brandt and Reimold (1999). (c) Contact between a 25 cm wide, vertically oriented carbonatite vein and Nebo granite just west of the northern view-point at Tswaing Crater. (d) Schematic section through the southeastern crater rim along the dirt road into the interior of the crater. The three sections refer: (A) showing the upper crater wall with ejecta breccia overlying Karoo grits, which, in turn, overlies steeply-outward dipping granite; (B) mid-section of the crater rim with the anticlinal structures also shown in the photograph of figure (e); and (C) schematic representation of the lower crater wall with generally shallow, inward-dipping granite (as indicated by the orientation of the traces of prominent pre-impact fractures that had an original subhorizontal attitude, in the granite). Note the varied orientations of fault structures in this rim section. (e) A 20 m wide section of the inner rim profile in the south-east part of Tswaing crater, detailed in (d). Image courtesy of Hans Knöfler (MfN Berlin). (f) Schematic cross section through the Tswaing Crater, as constructed from surface geological, borehole, and gravity information. Diagram modified after Partridge (1999) and Brandt and Reimold (1999). C = colluvium; CS = carbonate-rich sediments; FG = fractured granite; FGB = lithic (fragmental) granite breccia; GB = granite breccia; KG = Karoo grits; PIP = inferred post-impact profile; PP = present profile; SL = saline lake; SM = saline muds; T = talus. Figures (b), (d), and (f) are reproduced with permission of the Council for Geoscience, Pretoria (D. Barnardo, 2012, pers. commun.), with slight modifications.

(a)

(b)


28° 05′ E
N
25° 24′ 30″ S
25° 24′ 30″ S
Rim Section
28° 05′ E
0 50 100 150
m
Brine lake
Mud (maximum lake transgression)
Granite-derived sands
Coarse granitic scree
Karoo grits
Lamprophyre
Trachyte
Carbonatite
Granite
Vertical dip
Dips
Fault
Borehole
Trigonometric beacon
Drainage channel
Road leading to centre

(c)

(d)

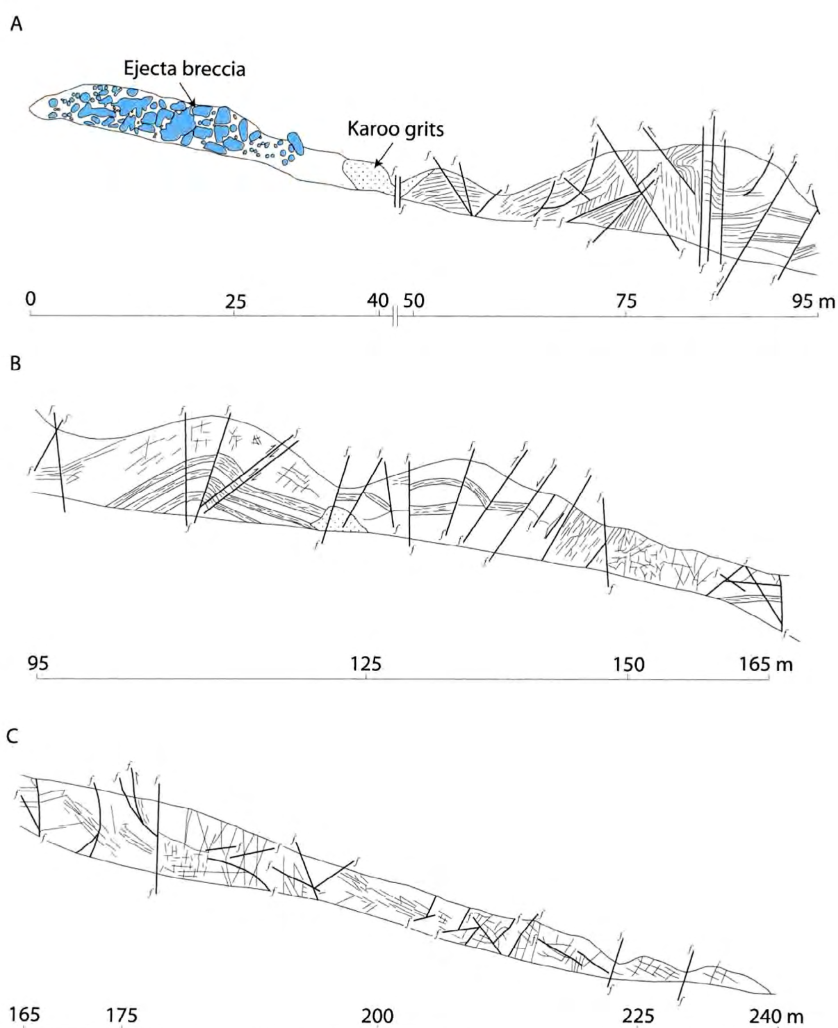


Fig. 36 (*continued*)

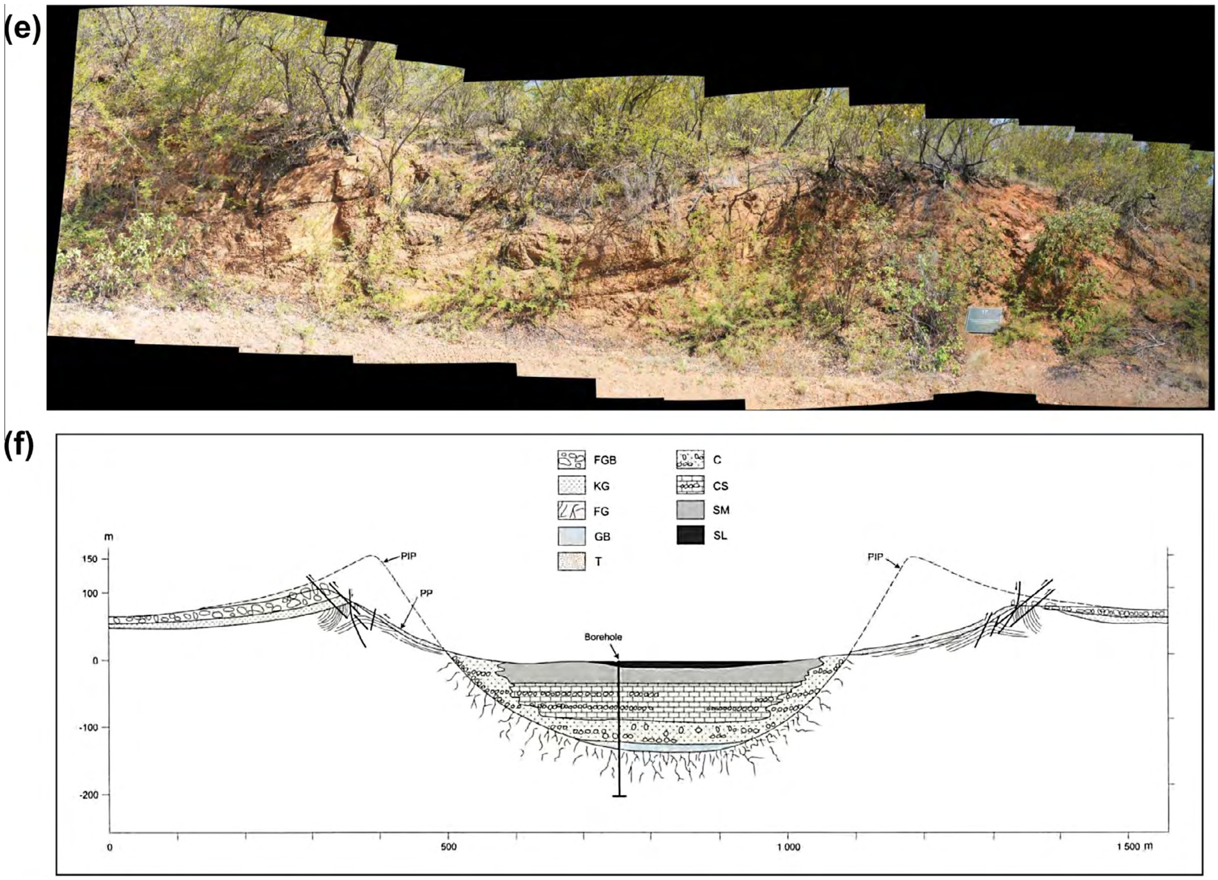


Fig. 36 (*continued*)

crater, a drilling project was initiated by the late T.C. Partridge, and a 200 m long drill core was extracted from the crater in 1988. The core spans, from bottom up, the entire range between crater floor in solid Nebo granite, fractured and brecciated basement, unconsolidated suevitic breccia, and finally an uppermost 90 m sequence of sedimentary crater-fill. A cross section through the crater, based on this drill core, is shown in Fig. 36f. The post-drilling paleoenvironmental and geological-mineralogical investigations were compiled by Partridge (1999a). Diagnostic evidence of shock metamorphism, in the form of PDF in quartz and feldspar particles, diaplectic quartz glass, and numerous particles of glass and melt fragments were detected in the suevitic breccia (Reimold et al., 1999b), so that the genetic controversy could be laid to rest in favor of impact. In addition, the impact melt phases were shown by Koeberl et al. (1994b, 1999) to contain a meteoritic component.

Detailed geochemical studies (Koeberl et al., 1994b) of the glasses and target rocks at Saltpan showed that the target granites have only limited compositional variability. The major and trace element composition of the bulk breccia is very similar to that of average basement granite. Impact glass fragments recovered from the unconsolidated suevitic breccia have a composition similar to that of the basement granites. No evidence for significant admixture of material from any of the minor intrusions (carbonatite, phonolite, trachyte, and lamprophyre) occurring in the crater area was found. The similarity of trace element abundances and ratios, and REE patterns between impact glasses and granites favors derivation of the glasses from the granites. The impact glass fragments show considerable enrichments of Mg, Cr, Fe, Co, Ni, and Ir, compared to the basement granites. The abundances of these elements in the glasses (after correction for indigenous concentrations) can be explained by admixture of $\sim$10% of a chondritic component. High Ir concentrations (up to 100 ppb) have been found in sulfide spherule samples, which may complement the (lower) Ir abundances in the glasses and could indicate some fractionation during impact.

Results of a Re–Os isotopic study on samples of the suevitic breccia were also reported by Koeberl et al. (1994b). They found that the target granites have very low osmium abundances of about 7 ppt and high $^{187}Os/^{188}Os$ ratios of about 0.72 that are typical for old continental crust. In contrast, the breccia samples were found to have much higher osmium abundances (about 80 ppt; which is a very low value) and lower $^{187}Os/^{188}Os$ ratios of about 0.205. These values can be explained by mixing of target rocks with a chondritic component.

The drill core also established the likely composition of the pre-impact target, namely a thin veneer of diamictite (a lacrustine deposit, traces of which occur throughout the unconsolidated suevitic breccia), above Karoo grits overlying, in turn, the Nebo granite.

Fission-track dating of impact glass separated from suevitic breccia (Storzer et al., 1999) yielded an age of 220 ± 52 ka for the impact event. In contrast, the volcanic lithologies occurring at and around the crater (Brandt, 1994) were shown by Brandt and Reimold (1999) to be part of the ca. 1300 Ma regional Pienaars River Alkali Granitic Suite (Harmer, 1985; Verwoerd, 2006). The fission-track age for the impact event overlaps, within error limits, the U–Th age of the Kalkkop crater (see above), suggesting that these two impact events could be related.

A further dating effort by the $^{40}Ar–^{39}Ar$ step-heating technique applied to Tswaing impact glass was made by Jourdan et al. (2007). This undertaking was, unfortunately, foiled by the presence in the samples of considerable amounts of argon inherited from the target rock. The Nebo granite of the Bushveld complex is at 2.06 Ga

Fig. 37. (a) The partially burnt-out Tswaing Crater Museum believed to have been the victim of arson. (b) The still incomplete Visitors Centre in the Vredefort World Heritage Area. The center was inaugurated in August 2008 with a function that was part of the Large Meteorite Impacts and Planetary Evolution IV Conference, but in March 2012 had not progressed from the status observed then. Instead, the building showed (c) dramatic indications of decay.

very much older than the rather young, 200 ± 50 ka, Tswaing crater. Based on these results for Tswaing and several other dating efforts, Jourdan et al. (2007) established a function according to which the discrepancy between impact and target rock ages is critical for the degree to which argon inheritance from target rock components in impact melt rock may be detrimental to the outcome of an argon chronological experiment with impactite specimens.

A limnological investigation of the crater lake (Ashton and Schoeman, 1983, 1988; Ashton, 1999; Schoeman and Ashton, 1982) showed that the lake is unusually shallow (<3 m depth), is characterized by lack of mixing of bottom and surface waters, and that it is hypersaline. Water analysis showed high levels of trace elements and dense populations of bacteria and cyanobacteria.

A series of publications in Partridge (1999a) review the results of the first comprehensive limnological and paleoclimatological investigation of the crater lake and the crater sediments intersected in the 1988 drilling. The upper 90 m of core provide a comprehensive record of sedimentary deposition under changing paleoenvironmental conditions. Ashton (1999) summarized that the ionic proportions of the lake water are characteristic for waters from alkaline igneous rocks such as the Nebo granite of the target. The dense phytoplankton population in the Tswaing crater lake is similar to that of other alkaline, saline lakes in Africa. Low wind-driven horizontal mixing patterns caused by low wind-speeds at the crater floor are responsible for the significant lateral heterogeneity of the nutrient content of the lake. Complete lack of mixing between surface and bottom waters is indicated by a continuously increasing salinity trend towards the bottom and which led to a progressive anoxia in bottom water.

Partridge (1999b) applied facies analysis on the lake sediment cores to reconstruct a paleoclimatic record. Variations in the amount of carbonates and other evaporites precipitated from the lake water reflect changes in the ratio between evaporation and rainfall since crater formation some 200,000 years ago. During wetter periods, the amount of clay and silt particles eroded from the crater walls was increased as well. Since then, further light has been shed on the environmental changes over the past about 200,000 years as deduced from analysis of the crater sediment column. Kristen et al. (2007) used high-resolution XRF scanning, basic geochemistry, organic petrology, and rock–eval pyrolysis for an identification of intervals of decreased carbonate precipitation, increased detrital input, decreased salinity, and decreased algal and organic matter content as proxies for environmental change.

Palynological analysis of the Tswaing crater sediment record provided new insights on the long-term pattern of climate and vegetation change for the interior of South Africa (Scott, 1999) during the periods 1–79 ka and 160–200 ka. McLean and Scott (1999) extended this work to the phytolith record, which provided some additional paleoclimatic evidence of cool periods during Middle to late Pleistocene times. Finally, Metcalfe (1999) reported the results of a detailed diatom study over the entire sediment core interval, which yielded information relating to long-term evaporitic changes and shorter-term climatic variations that affected the hydrological balance of the catchment (Metcalfe, 1999).

Putting all these lines of paleoenvironmental information together, Partridge et al. (1999) developed a model for paleoclimatic change over the last two glacial cycles in the interior of South Africa. The periodic lithologic and geochemical changes are recorded as sedimentary cycles recognizable in changes of the sediment facies. Between 200 and 80 kyr BP, the dominant periodicity of paleo-rainfall change is 23 ka reflecting a precession control on African subtropical climate (Partridge, 1999b; Partridge et al., 1999). More humid periods during the 80–10 ka interval seem to be out of phase with insolation changes. Kristen et al. (2007) and Schmidt et al. (2014) discuss whether the humid intervals (73–68 kyr, 54–50 kyr, 37–35 kyr and 15–10 kyr BP) could be related to southward displacement of the ITCZ (Inter Tropical Convergence Zone) and to changes in ocean circulation.

Kristen et al. (2010) compared the organic matter composition and carbon isotope systematics of modern and Holocene lake sediments and modern plant and lakewater samples from Tswaing. This revealed short-lived changes in the terrestrial and aquatic bioproductivity and provided insight into the change of the carbon cycle under the influence of changing climatic conditions over the interval from the last glacial to the late Holocene (14,000–2000 years BP).

At Tswaing, an effort has been made since the late 1980s to establish a crater museum that would not only present information on the solar system and its constituent bodies, introduce impact cratering in general, the nature of the Tswaing crater in all its aspects geological and environmental, but also regional geography and geology (i.e., South Africa's economic powerhouse, the Witwatersrand, Vaal Triangle, and Pretoria region) including that of the Bushveld Complex with its gigantic mineral resources – the proverbial treasury of South Africa. What is more, this museum was to educate about the history and socio-political evolution of this central part of South Africa. Remember that in the immediate environs of this crater structure several million people have found their homes in recent decades.

A well-designed museum building was constructed several years ago, but the responsible authorities never got around to develop exhibits. In 2010, the authors of this review visited the Tswaing crater and noted with great regret that part of the museum had burnt down (Fig. 37a). We want to emphasize strongly that the South African people cannot afford to neglect this wonderful, multidisciplinary education opportunity. One of us (WUR) has been involved with this crater museum project since its beginning in the late 1980s. For several decades the proposal of a national heritage site at Tswaing has been handled by the South African Heritage Resource Authority – to absolutely no avail. It has not even been possible to declare the unique crater site a Natural National Monument, and consequently its protection from vandalism is in no way assured. This is shameful in the face of the local people, who since 1988 have had high hopes of proper development at and around this museum site.

6.1.20. Vredefort, South Africa

The Vredefort impact structure, centered at 27°00′S/27°30′E, straddling the borders of the Northwest, Gauteng, and Free State provinces of South Africa, is widely known as the world's largest and oldest confirmed impact structure. But this has not always been the case. The origin of the conspicuous Vredefort Dome (Fig. 2a), the central part of the impact structure, has been controversial for nearly 100 years. Endogenic genetic processes were still considered as late as the 1990s (e.g., Colliston, 1990; Coward et al., 1995). Daly (1947) was the first to consider an origin of the Vredefort "Ring", as it was known then, by large-meteorite impact, and he was later followed by Dietz (1961) and Hargraves (1961), who emphasized that shatter cones represented definitive evidence of impact. This interpretation rapidly became quite widely accepted overseas by the early 1970s – but not in South Africa. There, controversy about the Vredefort origin was only concluded in the mid-1990s (see below).

A comprehensive historical and geological review was published by Gibson and Reimold (2008), and a natural history of the Vredefort Structure was produced as Volume 1 of the Springer "Geoparks of the World Series" by Reimold and Gibson (2010). The Geology of South Africa textbook of the Geological Society of South Africa (Johnson et al., 2006a,b) also contains a chapter on South African impact structures, also with a focus on the Vredefort structure (Reimold, 2006).

In July 2005, a portion of the Vredefort Dome with some of the most spectacular exposures of regional geology and impact-related deformation was inscribed by UNESCO as a World Heritage Site. So far, Vredefort is the only impact structure that has been honored, and that has had its importance emphasized, in this way. However, the South African authorities still have to declare this site a National Natural Heritage Site before this World Heritage status will become fully recognized. A modern visitors' center at the town of Vredefort has been under construction for more than 5 years. Its completion and proper staffing with geoscientifically trained managerial personnel would be mandatory for the Vredefort World Heritage Site to become functional as a unique educational and touristic landmark.

Gibson and Reimold (2008) and Reimold and Gibson (2010) have also described many of the most important exposures of mid- and upper crustal rocks that are so well exposed in the Vredefort Dome, as well as those exposures that show excellent displays of impact-generated deformation phenomena. Here, it is important to note that it is not permitted to visit these sites without permission of the respective landowners. And, furthermore, many of the most spectacular locations are within the designated World Heri-

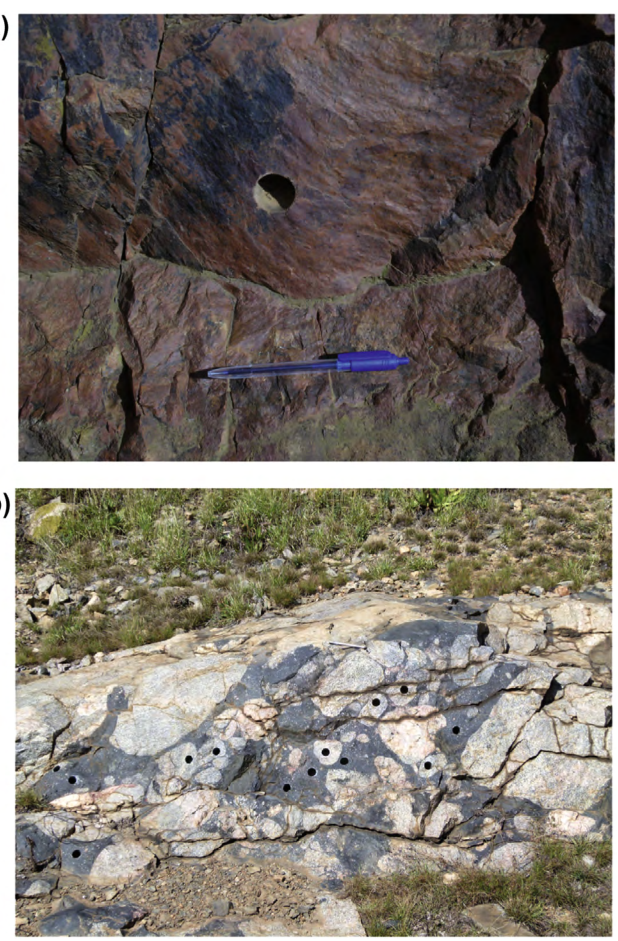
(a)
(b)


Fig. 38. Geovandalism: (a) A Shatter cone in an outcrop along the Potchefstroom-Schoemansdrif road, western collar of the Vredefort Dome, destroyed by indiscriminate drilling, presumably for geoscientific purpose. Pen for scale 13 cm long. (b) Part of the exceptional exposure of pseudotachylitic breccia in the famous Otavi Quarry within the northeastern core of the Vredefort Dome. There are numerous less obvious sites where this impact-generated breccia could have been drilled for paleomagnetic analysis.

tage Area where they are comprehensively protected from rock sampling or, in general, removal of rocks and vegetation. Despite this legal protection, a significant number of exposures has already been vandalized (Fig. 38) by obviously ignorant members of the public, but also by professional geologists, who should be considered educated about the intrinsic value of geosites. A case in point are those geophysicists who have over the last decade and a half indiscriminately drilled (for paleomagnetic sampling) into exceptional outcrops, and – senselessly – even into breccias where the sample cylinders removed could not relate a useful paleomagnetic signature in any case. Readers may wonder why the shatter cone in Fig. 38a was destroyed by such a geo-vandal!

The Vredefort structure is centered some 130 km to the southwest of Johannesburg, near the geographic center of the structural remnant of the Witwatersrand basin (Fig. 39a). The Vredefort structure comprises the Vredefort Dome (Fig. 39b) and the surrounding basin. The central part of the Vredefort structure, the Vredefort Dome, has been the focus of geological interest since 1868 (Stow, 1879), when it was first noted in the course of exploration activity that the geology of the Dome is unique in the regional context. In particular, the Vredefort Dome is important as it represents (1) a unique window into the geology of the upper and middle crust of the Kaapvaal craton; (2) it has regionally unique, spectacular rock deformation phenomena, which have long indicated an enigmatic, cataclysmic origin of the Dome; (3) the origin of the Vredefort Dome has long been controversial, with both endogenic (such as gas explosion or tectonic processes) and exogenic processes having been invoked (see review by Gibson and Reimold, 2008); (4) in the mid-1990s it was recognized that the entire Witwatersrand Basin represents the actual erosional remnant of the Vredefort impact structure; and (5) the Vredefort impact has lead to the preservation of the Witwatersrand ore resources, and also caused widespread hydrothermal overprint onto them (see also economic significance of impact, above). Figure 39a illustrates the concentric fold pattern around the dome that affects the entire basin extent.

Since the last review of African impact structures (Koeberl, 1994) progress has been made with regard to just about every aspect of the Vredefort impact structure – regional geological understanding, evidence of impact, the metamorphic history of the region, structural mapping, detailed investigations of shatter cones and pseudotachylitic breccias, as well as microdeformations in quartz and zircon, in the realm of the Vredefort dome but also far beyond downstream in the Vaal River bed, determination of the age of the impact, the effects of the impact on the ores of the Witwatersrand Basin, and numerical modelling of the impact event. These results have not only furthered the understanding of the Kaapvaal craton and the Vredefort impact structure but have contributed greatly to the general knowledge about impact processes and the Witwatersrand ore deposits.

6.1.20.1. Geology of the Vredefort Dome and surrounding basin. The Witwatersrand Basin (Fig. 39a) is completely underlain by crystalline basement of the Kaapvaal craton. This is followed stratigraphically upwards by Dominion Group strata in the western part of the basin around Klerksdorp and in the area of the Vredefort Dome (Marsh, 2006), and by subsequent deposits of the Witwatersrand (McCarthy, 2006), Ventersdorp (van der Westhuizen et al., 2006), and Transvaal (Eriksson et al., 2006) supergroups (Fig. 39c). Basement exposures occur in the environs of the Basin (e.g., the Johannesburg Dome between the cities of Johannesburg and Pretoria), but none of these exposures shows any of the deformation phenomena for which the Vredefort Dome is famous. In the southern part of the basin, including the southern parts of the Vredefort Dome, the Archean and Proterozoic basement and supracrustal strata are obscured by a thick blanket of sediments and dolerites of the Mesozoic Karoo Supergroup (Duncan and Marsh, 2006; Johnson et al., 2006a,b).

Fig. 2a shows a satellite image of the Vredefort Dome that emphasizes the terrain dichotomy comprising the Vredefort Mountain Land in the northern and northwestern environs (the so-called "collar") surrounding the comparatively rather flat "core" of the Dome. The latter is mostly agriculturally used land where Karoo-derived soils prevail, but there are also many excellent exposures including quite a few abandoned dimension stone quarries, some of which offer three-dimensional views into the upper and mid-crust of a craton. The dome's collar and northwestern/northern sectors of the core have numerous good exposures and have provided extensive structural geological information.

The schematic geology of the Vredefort Dome is shown in Fig. 39b. And in Fig. 39c the general stratigraphy of the region is recalled, while Fig. 39d shows an interpretation of regional reflection seismic data along a profile across the Witwatersrand basin, including the Vredefort Dome. The southeastern sector of the dome is covered by Jurassic Karoo strata. The 45–50 km wide core region is largely composed of Archean granite-gneiss. The 20–25 km wide collar comprises up- and overturned supracrustal strata. The core gneisses were traditionally subdivided into an inner zone of granulite-facies Inlandsee Leucogranofels (ILG) and an outer, amphibolite-facies annulus of the so-called Outer Granite Gneiss (OGG) that involves a series of heterogeneous, largely migmatized granitic and granodioritic gneisses. The area straddling the OGG-ILG contact was termed the Transition Zone by Hart et al. (1990), who claimed that it was the locus of a semicircular Vredefort Discontinuity interpreted by them as either a major intracrustal "discontinuity" or a major shear zone or décollement. Strong brecciation characterizes this zone that displays a lot of pseudotachylitic breccia and charnockitic and enderbitic gneisses. Charnockite is also a major lithology in the third traditionally recognized terrain, the Steynskraal Metamorphic Zone (SMZ, Stepto, 1990) of the innermost dome sector, which was interpreted to represent a large block of granulitic and amphibolitic gneisses and metapelites. This block was thought to be perhaps as old as 3.5 Ga, based on early Rb–Sr isotope geology.

In contrast, U–Pb single-zircon dating by Kamo et al. (1996) and ^{40}Ar–^{39}Ar data for an OGG amphibolite by Reimold et al. (1992) suggested that the granitoid basement of the Dome was 3.1–3.2 Ga old and underwent metamorphism around 3.08 Ga ago. This was confirmed by a comprehensive SHRIMP U–Pb zircon dating study by Armstrong et al. (2006), who dated a range of granitic and granodioritic lithologies that C. Lana had identified as major lithologies of the Archean basement (Lana, 2004; Lana et al., 2003a,b, 2004). A few zircons indicated that the basement in the core of the Vredefort Dome contains a component of granitoid as old as 3.3–3.4 Ga (compare Fig. 39e).

Lana et al. (2003a,b), based on comprehensive lithological and structural mapping, have shown that the traditional zonation (OGG-SMZ-ILG) of the basement in the Vredefort Dome is no longer tenable (although it is still referred to even in recent publications, such as Harris et al., 2013). Instead the entire core represents a terrain made up of different types of granitoids that are distributed in a complex pattern. There is no evidence for a discontinuity that juxtaposes two terranes; instead, the transition from amphibolite to granulite facies is gradational as one would expect along a traverse from upper crust to mid-crust. The structural mapping has shown that fabrics across the alleged transition zone are continuous. Lana et al. (2003b) also found that the outer part of the dome was rotated into vertical attitude – as a result of the impact event (also Lieger et al., 2009).

The collar stratigraphy encompasses (from the contact with the core outward) a succession of locally occurring Dominion Group rocks followed by the sequences of the Witwatersrand, Ventersdorp and Transvaal supergroups, with increasing amounts of

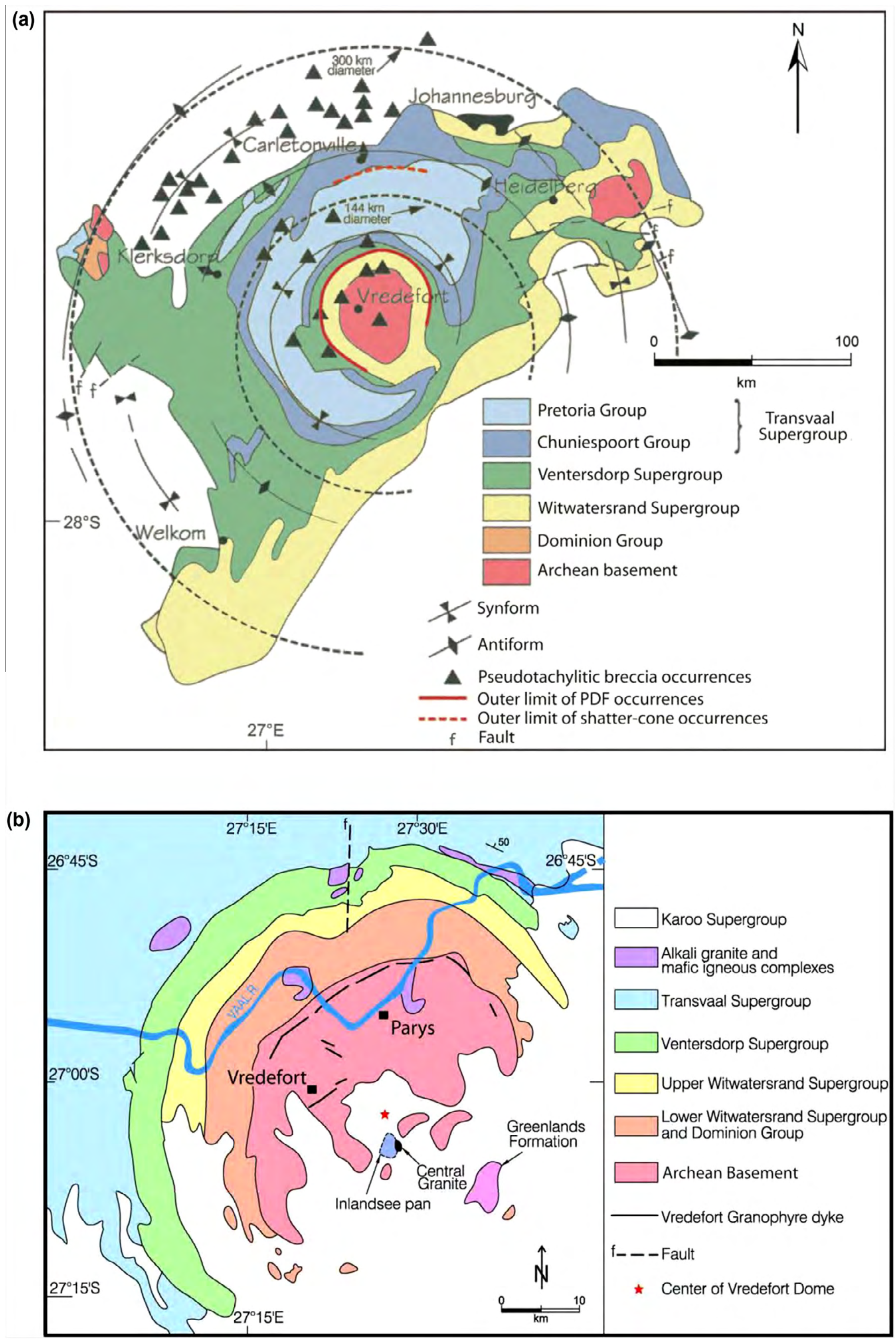


Fig. 39. (a) Generalized geology of the Witwatersrand Basin (South Africa), with the Vredefort Dome in its central area. Also shown are the extents of pseudotachylitic breccia occurrences, shatter cone observations, and the outer limit of occurrence of planar deformation features. Image modified after a diagram by Mohr-Westheide (2011). (b) Schematic representation of the geology of the Vredefort Dome (see text for detail). Modified after a diagram by Mohr-Westheide (2011). (c) Schematic stratigraphic column for the region of the Witwatersrand basin/Vredefort impact structure, from the Archean granitoid basement upward towards the top of the Transvaal Supergroup that would have formed the uppermost part of the target stratigraphy for the Vredefort impact event. Modified after a diagram by Reimold and Gibson (2005). (d) Structural interpretation of reflection seismic profiles across the Witwatersrand Basin, including the Vredefort Dome. Diagram after work by Friese et al. (1995), modified by and courtesy of A. Jahn (formerly Museum für Naturkunde Berlin). Note the extensive block faulting in the rock of the collar of the Vredefort Dome, and the asymmetry of the current structural state of the impact structure along this northwest–southeast profile.

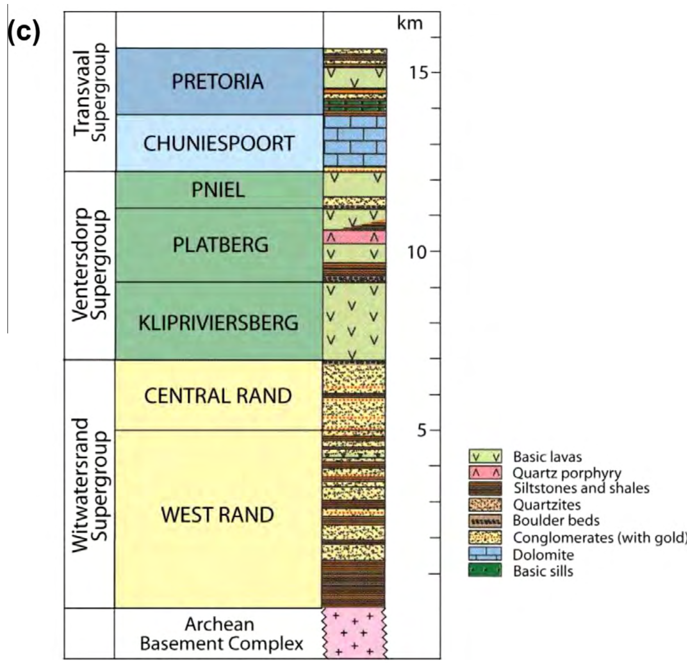


Fig. 39 (*continued*)

overturning the further outward the traverse reaches in the northern-to-western sectors. In contrast, the supracrustals in the southern parts of the dome have not been up- or overturned, as determined by drill cores (e.g., Friese et al., 1995). Various ideas have been proposed to explain this structural asymmetry of the dome – suggesting that either the attitude of beds was asymmetrical already before the impact, or that tilting occurred after the event (e.g., Henkel and Reimold, 1998; Lana et al., 2003a; Wieland et al., 2005). An attempt to reconstruct the pre- to post-impact structural evolution of part of the collar was contributed by Jahn and Riller (2009).

The collar strata were intruded by numerous sills of dioritic composition, known as epidiorite due to their metamorphic overprint. A Bushveld magmatism (2.06 Ga, Cawthorn et al., 2006) association of some of these intrusive rocks has been discussed for a long time, most recently by Coetzee et al. (2006), although the chemistry and especially Sm–Nd data strongly suggest that most of these metamorphosed dioritic rocks correspond to the Ventersdorp (2.7 Ga, Armstrong et al., 1991) magmatic phase and likely form the feeder dikes for this regional extrusive event (Pybus et al., 1995; Reimold et al., 1995). It can, however, not be excluded that some of these intrusive bodies could be related to the Bushveld emplacement event.

Indeed, a number of intrusive alkali granitic and associated mafic to ultramafic rocks occur in the form of discrete complexes in and around the collar of the dome: the Roodekraal, Rietfontein, Lindeques Drift and Schurwedraai/Baviaanskrans complexes (Fig. 40b). All these bodies are related to the Bushveld emplacement event at 2.06 Ga – as partially confirmed by radiometric dating (Gibson and Reimold, 2008, and references cited therein). And to the north of the collar, near the town of Fochville occurs a layered mafic body, the Losberg Complex, which has also been related to the Bushveld Complex by its distinct litho-stratigraphic sequence and as indicated by some radiometric dating results (Coetzee and Kruger, 1989).

A small exposure of apparently undeformed biotite granite occurs close to and to the east of the centrally located Inlandsee pan in the core (Fig. 40b). Hart et al. (1981) termed this exposure "Central Intrusive Granite". In 1991 Hart et al. proposed that it represented a local occurrence of anatectic granite and changed the name to "Central Anatectic Granite". Gibson et al. (1997) obtained a SHRIMP single-zircon U–Pb age of 2017 ± 5 Ma for this phase. They also reported both shock metamorphosed zircon grains – of pre-impact age – as well as an unshocked zircon phase in this granite. While the evidence for impact at Vredefort has not been introduced here yet (see below), it should be noted at this time that a 2023 ± 4 Ma SHRIMP single-zircon U–Pb age for authigenic, unshocked zircon grains from pseudotachylitic breccias (Fig. 41a–c) and impact melt rock (Vredefort Granophyre, Fig. 40a–c) by Kamo et al. (1996) does constrain the age of the Vredefort impact event (Fig. 41e). Concerning the biotite granite, the age obtained by Gibson et al. (1997) implies that this granite represents a phase that was generated (melted) at the time of impact – and likely as a consequence of impact due to very high shock pressure/shock temperature overprint on already hot mid-crustal rock (Gibson and Reimold, 2005; Ivanov, 2005). Currently, the term Vredefort Central Granite is preferred.

A post-impact monzodiorite intrusion occurs with exposures across the core; it is known as the Anna's Rust Sheet with its main exposure north of Parys (compare Fig. 39b) and other exposures or intersections near the town of Vredefort, in a borehole (designated "Beta") near the centrally located Inlandsee pan, and on several farms southeast of Vredefort town. Pybus (1995) investigated this body and determined a ca. 1050 Ma Rb–Sr age for it. Reimold et al. (2000b) related this intrusive event to other widespread magmatic activity on the Kaapvaal craton around 1200–1000 Ma ago.

A few kilometers northeast of the Inlandsee pan occurs a poorly exposed, apparently unshocked wherlite body of limited extent and still unknown age. In the southeast part of the dome is a large inlier of Archean greenstones located, the so-called Greenlands Formation (Fig. 39b; Minnitt and Reimold, 2000; Lana et al., 2003c). Komatiitic basalts and associated lithologies are strongly reminiscent of lithologies occurring in the Barberton Greenstone Belt. Limited dating of Greenlands rocks (Gibson and Reimold, 2008) suggests an age in excess of 3.4 Ga for this formation. There is an excellent exposure of pillow lavas and widespread occurrence of shatter coning. Within the area of this inlier occur two large

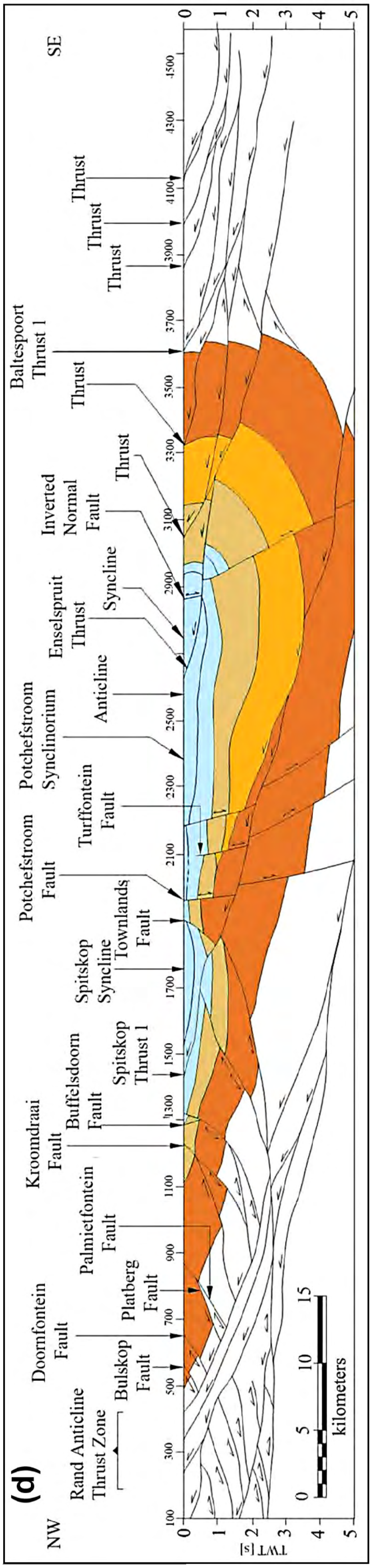


Fig. 39 (*continued*)

mines producing bentonite from Karoo age voelcanic deposits (Gibson and Reimold, 2008). On farm Avondale, east of the Greenlands Greenstone Area, is a prominent exposure of a Li-pegmatite body.

6.1.20.2. The Vredefort Granophyre: impact melt rock. A series of 9 dikes is marked on the geology map of Fig. 39b. The so-called Vredefort Granophyre (Fig. 40a–c) dikes occur either within the core granitoids as far as 9 km from the collar contact (4 known dikes), or straddling the core-collar boundary. Those dikes occurring within the core are oriented northwest-southeast or northeast–southwest, the directions that represent the two main fabric orientations in the gneisses of the core. The Granophyre received its name from the distinct granophyric (also termed micropegmatitic) groundmass texture that is characteristic for much of the melt rock. It is mainly composed of plagioclase laths and hypersthene needles set into a microphyric groundmass of various feldspar minerals and quartz, besides minor biotite and opaques. The Granophyre is regionally of homogeneous chemical composition that is rather unique in combining the silica content of a granodiorite with comparatively high abundances of Ca, Mg, and Fe (e.g., Reimold et al., 1990; Therriault et al., 1996; Reimold and Gibson, 2006). Only one other site is known where such a composition has ever been detected as well, namely, the *Morokweng* impact melt rock (see above). Only locally do slight compositional deviations from the general composition of the Vredefort Granophyre occur, markedly where a Granophyre dike crosscuts epidiorite along the core-collar contact and has assimilated some locally derived host material (Therriault et al., 1997). This is particularly important – and well exposed – on farm Kopjeskraal in the northwestern sector where a Granophyre dike has assimilated material from an epidiorite dike (an intrusion thought to be related to Ventersdorp magmatism). This situation has recently become the topic of a publication by Lieger and Riller (2012) that ought to be considered very carefully in the light of this epidiorite-Granophyre melt rock mixing that remained undetected by these authors. Reimold et al. (2013b) report results of detailed geochemical analysis of a profile across the Granophyre dike and into epidiorite from this location and found clear evidence for "normal" Granophyre composition along the contact of the dike to Archean granite, mixed (hybrid) compositions in the interior of the dike, and proper epidiorite composition on the other side. Similar findings have since been made on farm Rensburgsdrift at a location where a Granophyre dike cuts across epidiorite as well.

All dikes are more or less clast-laden, carrying besides a major clast component derived from the granitoid basement a prominent proportion of Witwatersrand strata-derived clasts (shale $\ll$ quartzite). There are two basic textural variations of this rock type, specifically with spherulitic growths of hypersthene needles and a more granoblastic variety with stubby, prismatic orthopyroxene crystals. Granophyre has long been known to be related to the catastrophe that caused the doming event, as it is one of the rare lithologies on the dome that is not significantly affected by rock deformation (only two locations with very thin veinlets of pseudotachylitic breccia have been referred in the literature (e.g., Reimold et al., 1990; Therriault, 1992).

Kamo et al. (1996) reported the first occurrence of shock metamorphism in clasts of Granophyre, in the form of planar-fractured zircon and granular shock texture in this mineral. Earlier workers had searched in vain for shock metamorphic textures in quartz and feldspar. Only in 2002 did Buchanan and Reimold finally detect shock metamorphic deformation in quartz within Granophyre, including planar deformation features and what they termed a vermicular texture of partially shock-melted quartz reminiscent in appearance of checkerboard feldspar, which has also been

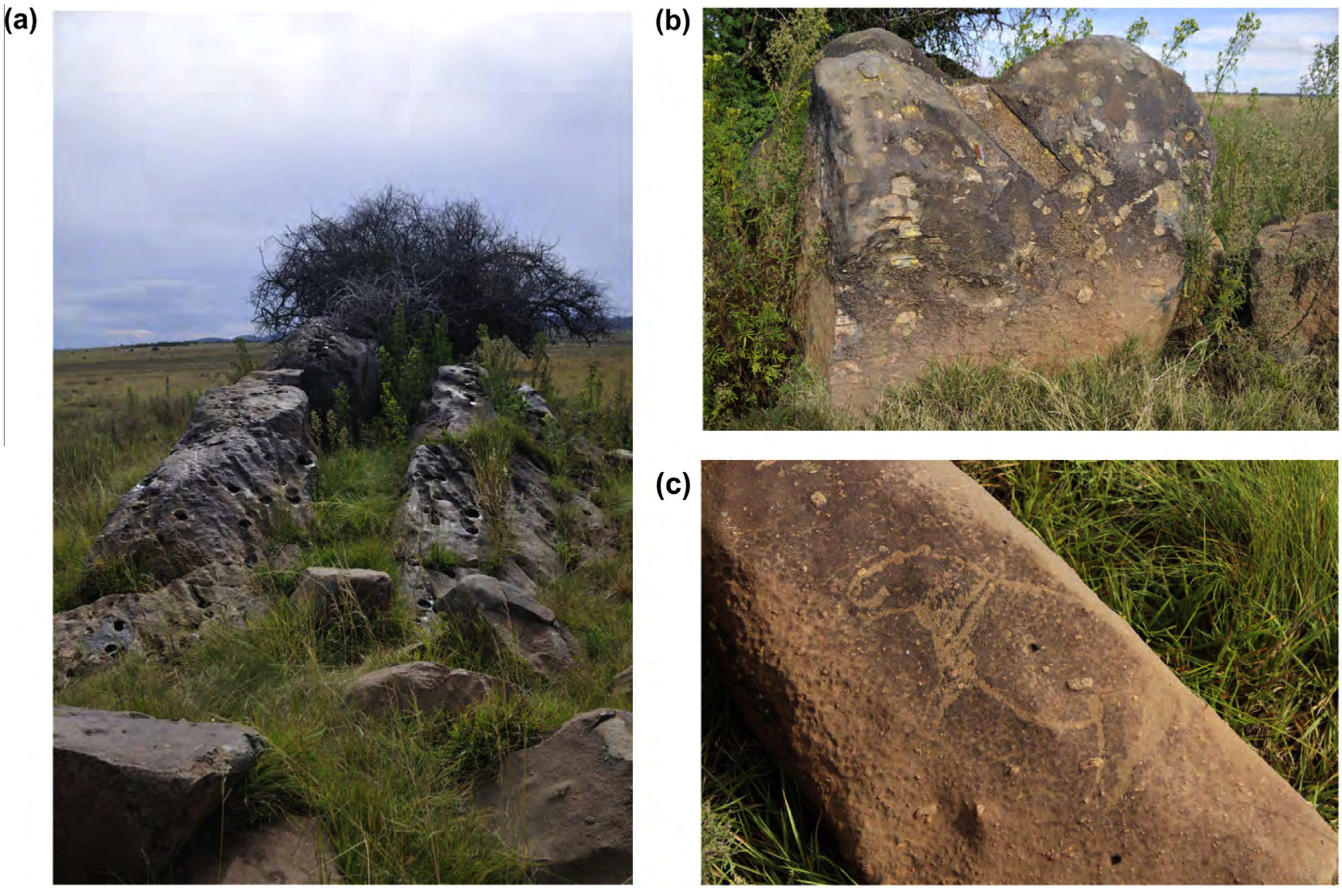


Fig. 40. Vredefort Granophyre (impact melt rock): (a) Part of the well-exposed granophyre dike on farm Daskop in the core of the Vredefort Dome. Note the apparent rarity of visible clasts (and compare against (b)) and the abundance of pits in the rock. The latter is thought by WUR to be the result of selective weathering of granitic clasts. Archeologists, in contrast, prefer spotty dissolution of melt rock below dripping trees. The segment of dike shown is approximately 10 m long. (b) Clast-rich portion of the same dike as shown in (a), just 2 m further along the dike. Note the abundance of granitic clasts and their size variation. Pocket-knife for scale about 10 cm long. (c) One of the marvelous petroglyphs etched onto this dike, which illustrates the spiritual value that this site had for the San people. The petroglyph is about 30 cm long.

observed in clasts within Granophyre (Fig. 12d). Both these textures testify to very strong thermal overprint.

Originally three hypotheses had been suggested to account for the formation of the Granophyre: (i) that it represented a "glorified equivalent" to the pseudotachylitic breccia (Hall and Molengraaff, 1925), (ii) that it was the result of assimilation of much crustal material by a mantle-derived mafic magma of dioritic or lamprophyric composition (Bisschoff, 1972, 1982), and (iii) that it was in fact impact melt rock (French and Nielsen, 1990; Reimold et al., 1990). The aforementioned findings of shock metamorphosed clasts, and the important recognition by Re–Os isotope chemistry of a definite, albeit small, meteoritic component in the Granophyre (Koeberl et al., 1996) finally provided the proof that the impact melt rock hypothesis was correct – and laid the controversy about the origin of the Granophyre and, in fact, that about the genesis of the Vredefort Dome to rest. Reimold et al. (1990) established that the Granophyre could be modeled as a melt mixture resulting of 75% granitoid with <3.5% quartzite, and 25% shale derived from the Witwatersrand Supergroup, consistent with the general clast content of the Granophyre. It should be noted that the bulk chemical work that has been done in the past has obtained analyses for mixtures of groundmass and clasts (much of the latter being quartzite). The groundmass texture does, however, clearly show that this component comprises an intimate assemblage of crystals grown from the melt phase with remnants of partially digested microclasts that, by all means, cannot be separated.

No evidence of a mafic component, such as a Ventersdorp Supergroup-derived one advocated by Lieger and Riller (2012), has ever been detected in samples not contaminated with locally derived epidiorite.

Besides the locations where Granophyre cuts epidiorite, Ventersdorp Supergroup lava and Transvaal Supergroup rocks (dominantly carbonate, some quartzitic and argillitic rocks) do not feature amongst the clast population of Granophyre. This can be explained by considering that the large extraterrestrial projectile that caused the Vredefort impact would have measured between 5 and 15 km in size – depending on its mass (i.e., density) and its equally unknown velocity. It would have penetrated into the target rock by a measure akin to its diameter, where it would then have exploded to create the enormous shock front. Thus, rock melting would have been caused inside the target and not close to surface. This could explain why relatively surface-near Ventersdorp lava would have been vaporized to a large degree. The uppermost target strata of the Transvaal Supergroup would have consisted to a very large degree of carbonates that would have been shock vaporized or at least dissociated into CO_2 and oxides that might have rapidly altered or transformed back to carbonate phases.

6.1.20.3. Metamorphism. Another important aspect of the Vredefort Dome that was recognized very early on in the history of the geological exploration of this region (Hall and Molengraaff, 1925) is the local metamorphic *hot spot* that is centered on the Vredefort Dome. In the Lower Witwatersrand and Dominion rocks of the collar the metamorphic grade is mid-amphibolite and is significantly higher than the lower greenschist Grade of the stratigraphically equivalent rocks out in the surrounding Witwatersrand basin. In

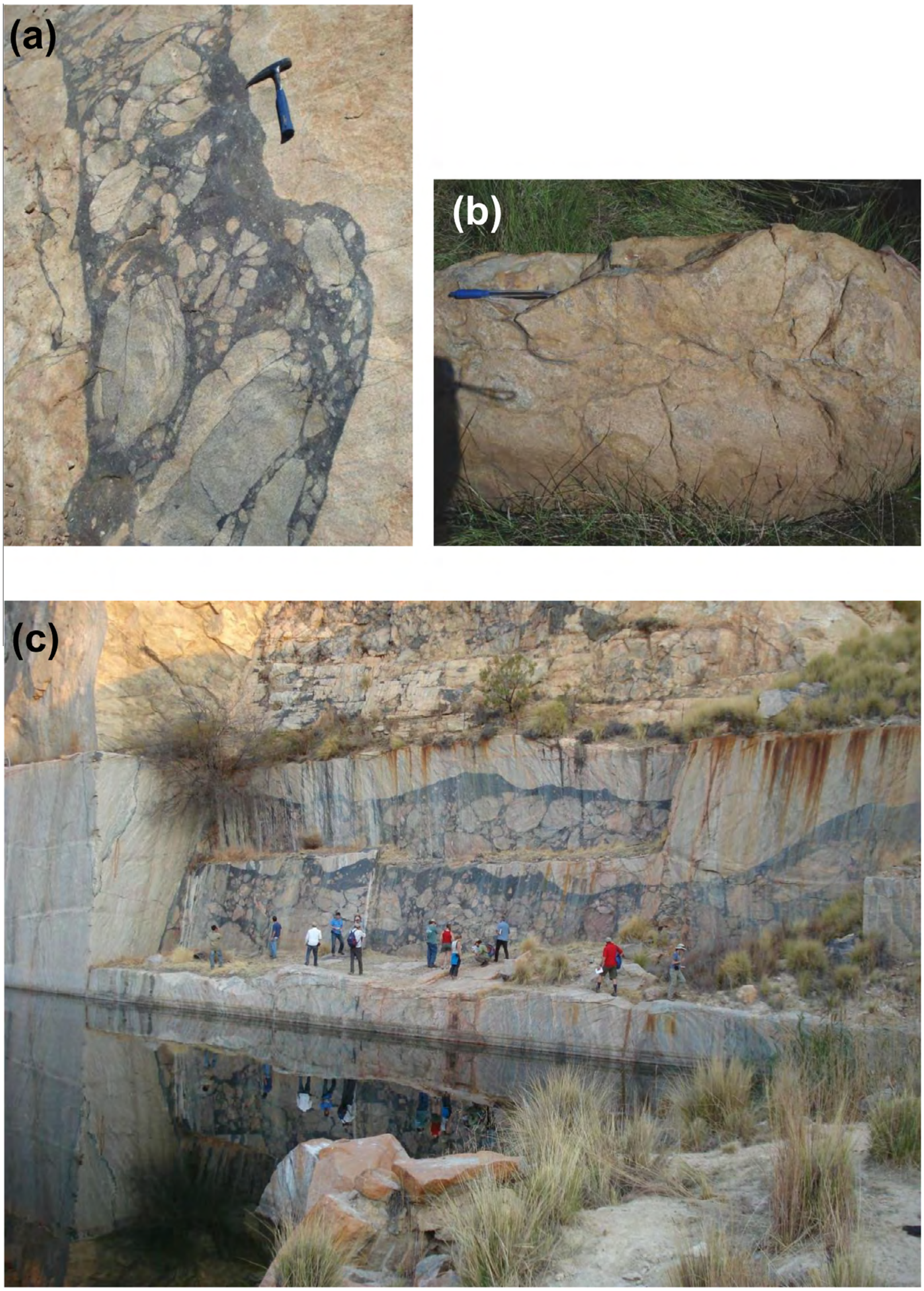


Fig. 41. Various manifestations of pseudotachylitic breccia from the Vredefort Dome (South Africa). (a) Large exposure of pseudotachylitic breccia at Salvamento Quarry in the northern core of the Vredefort Dome. Note the set of radially oriented clasts that obviously are fragments of a common precursor that have been "pulled apart". (b) Millimeter wide, meandering, coesite-stishovite-bearing veinlet of PTB in quartzite from the outer collar of the Vredefort Dome at Kromellenboog. (c) In contrast to (b), here a portion of the Leeukop Quarry exposure of quite massive PTB north of Parys in the northern core of the Dome. Each of these quarry benches is about 3 m high. (d) Pocket of pseudotachylitic breccia in Dominion Group meta-lava (amphibolite) on the property of Kopjeskraal Country Lodge, northwestern edge of the core of the Vredefort Dome. South African 2-Rand coin for scale is 1.5 cm in diameter.

the core of the Vredefort Dome a metamorphic gradient from the center outward involves granulite facies and, further out, amphibolite grade. The enhanced metamorphic grade of the collar rocks was at first perceived, by some, as a contact metamorphic aureole around a hidden pluton (Bisschoff, 1982), until it was shown by Hart et al. (1999) that the high-grade metamorphism of the core rocks represented Archean regional metamorphism. According to Gibson and Wallmach (1995) the metamorphism of the Witwatersrand collar strata followed an anti-clockwise P–T–t path – incompatible with contact metamorphism. Since then it has been

(d)

Fig. 41 (*continued*)

well established that this metamorphic phase occurred much later than the first Archean phase of metamorphism, began prior to the Vredefort impact event, but outlasted it so that impact-generated lithologies could also be metamorphosed (Stevens et al., 1997; Gibson and Reimold, 1999, 2001). It is believed that this metamorphic event was related to the emplacement of the Bushveld complex, some 35 Ma prior to the Vredefort event, which is also supported by argon dating results (Gibson et al., 2000).

Most recently, a detailed study of the textural and chemical characteristics of the high-grade rocks of the granulite terrane historically known as the Steynskraal Metamorphic Zone was carried out by Ogilvie (2010), who showed that these rocks provide a continuum of assemblages that range from extreme disequilibrium, to partial equilibrium, to, finally, full equilibrium. She found that during the peak anatectic granulite event between 3.12 and 3.08 Ga (Lana et al., 2004; Armstrong et al., 2006), full equilibrium conditions were attained for major mineral components during prograde metamorphism. This peak assemblage was then overprinted by shock-induced extreme disequilibrium deformation features. Finally, non-adiabatic decay of the shock front released excess energy into the target rocks, causing post-shock metamorphism. This metamorphic phase is manifest as the diffusion-controlled replacement of garnet by corona reaction products in partial equilibrium.

Following the second metamorphic event, the currently exposed rocks of the Vredefort Dome became subject to the impact event and related impact metamorphism. Detailed petrographic analysis by Gibson and Reimold (2005) established that it is possible to observe evidence for progressively decreasing shock and thermal metamorphism from the center of the dome outward. The shock pressure and shock temperature levels suggested by these authors turned out to be consistent with the impact modeling results (Fig. 2c) of Ivanov (2005).

6.1.20.4. Structure. Like the complex polymetamorphism of the Vredefort Dome, the structural (deformation) history of the region is highly complex and protracted. Only in the last decade or so has a detailed analysis of the structural geology of the Archean Basement Complex (the core of the Dome) been done (Lana et al., 2003a,b,c, 2004; Armstrong et al., 2006), aided by lithostratigraphic mapping and extensive U–Pb single-zircon dating (as reviewed by Gibson and Reimold, 2008).

At least four Archean deformation events with related ductile deformation features have been identified. Lana et al. (2003b) demonstrated that both the amphibolite-grade outer part of the core and the granulitic inner part display the same structural evidence (i.e., these zones are not separated by a structural discontinuity). An earliest S1 foliation is transposed by S2, which results in centimeter-scale S1 isoclinal fold structures within the S2 foliation. The shallowly dipping S1/S2 foliation was then folded during the D3 deformation phase on a scale of meters to decameters, whereby upright, variably plunging open-to-tight folds were generated. D3 folds become tighter in the vicinity of subvertical high strain zones, which range in width from meters to hundreds of meters. The timing of D3 was established to be related to prograde to peak-metamorphic conditions.

Kisters and Reimold (2000) identified a fourth deformation event that involved rotation of the S3 fabric between large, up to several kilometer wide, blocks that are delimited by narrow ductile shear zones. It was suggested that D4 could be related to the impact event, but there is no concrete evidence – such as concomitant development of pseudotachylitic breccias that one could think of – to support this possibility. Lana et al. (2003c, 2006) described D4 as a retrograde mylonitic shearing event that resulted in juxtaposition of the high-grade gneiss terrane against the greenstone complex in the southeastern sector of the dome. Armstrong et al. (2006) provided evidence suggestive of D4 having commenced close to the peak of metamorphism but continuing during temperature waning.

Lana et al. (2003a) interpreted their structural geological data to indicate that the subvertical concentric fabric in the outer, 8 km wide, parts of the core represented the originally subhorizontal S1/S2 fabric that had been rotated by ca. 90° during dome formation (see also Lieger et al., 2009). Wieland et al. (2005), who discussed structure of the northwestern to northeastern sectors of the collar, preferred a larger degree of rotation of up to 120° during dome formation. Gibson and Reimold (2008) stated that "closer to the center of the core, impact-related rotation decreases quite sharply and S1/S2 in the central core retains its shallow orientation." They concluded that the core had a plug-like geometry that is very difficult from the historical "crust-on-edge" model of earlier Vredefort workers (e.g., Hart et al., 1981). It does, however, agree very well with the results of numerical modelling of the collapse phase of the Vredefort central uplift by Ivanov (2005).

A small number of structural studies have dealt with folding and faulting in the dome and its environs. Simpson (1978) mapped the kilometer-scale gentle fold structures in the Pretoria Group rocks (Upper Transvaal Supergroup) of the Potchefstroom Synclinorium to the north of the dome. Lilly (1981) mapped a variety of fault and joint orientations in the collar that he linked to the formation of the dome. He also discussed repeated events that had created "shock deformation in quartz" and were indicative of

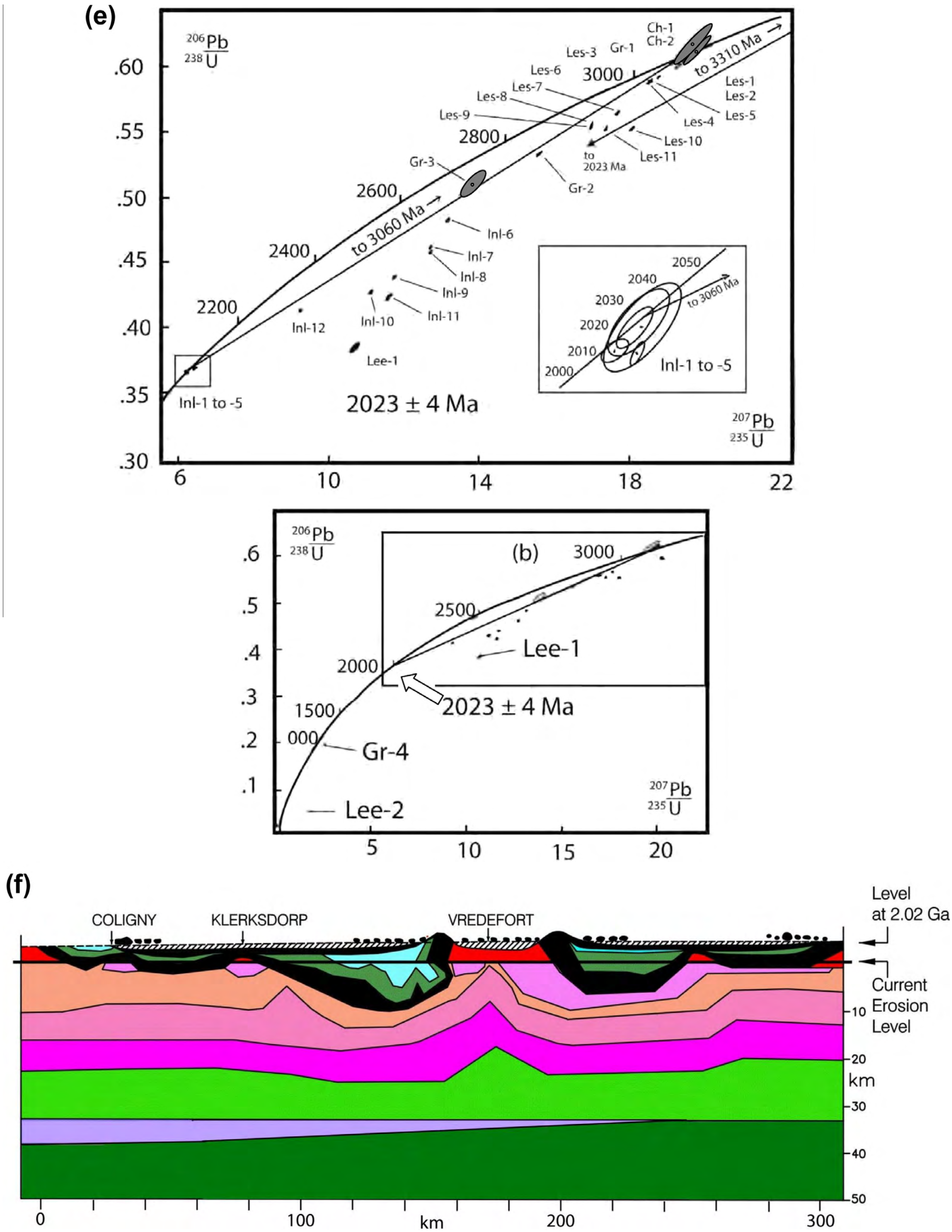


Fig. 41. (e) Northwest (left) – southeast gravity model across the Vredefort impact structure (modified after Henkel and Reimold, 1998). Obviously, the part above the current erosion level has been hypothesized. Crustal structure based on combined reflection and refraction seismic information, as detailed by these authors. Overall the complex geometry of a several hundred kilometer wide impact structure with a large central uplift emerges from this modelling. (f) U-Pb isotope Concordia plot for data for zircon extracted from pseudotachylitic breccia and Vredefort Granophyre (impact melt rock). Modified after figure 5 of Kamo et al. (1996), with permission of the Journal of African Earth Science editor (Tim Horscroft, pers. commun.)

internal shock sources; this has not been supported by any of the further detailed shock investigations carried out since. Brink et al. (1997, 2000) and Wieland et al. (2005) discussed two large pre-impact faults in the northern collar and related them to Ventersdorp-time extensional geology. Lana et al. (2003b) suggested that the northwest-southeast elongation of the dome and the obliquity between the collar strata and the pre-doming metamorphic isograds discussed by Bisschoff (1982) could mean that bedding before doming was not oriented horizontal during the metamorphic event and prior to the impact event. They calculated that a 30° tilt down towards the northwest could account for the perceived asymmetry in collar bedding. Wieland et al. (2005) reported structural measurements from the collar that they interpreted to infer that a 3° tilt was sufficient to account for the asymmetry of the collar. Wieland (2007, also Wieland et al., 2005) investigated the fold structures affecting the inner core strata and determined the tangential shortening related to the uplift of the dome. They determined that impact-related movements were predominantly vertical.

Jahn and Riller (2009) investigated large-scale concentric and radial faulting in the collar of the dome. They concluded: "The faults accomplished constrictional (centripetal) rock flow followed by radial spreading of uplifted and gravitationally unstable rocks in the crater centre. More specifically, concentric faults formed likely as normal faults during transient crater rim collapse, steepened and were transformed to reverse faults during central rock uplift and finally, were overturned during gravitational collapse of the central uplift. Radial faults, by contrast, formed at a later stage of convergent rock flow towards the crater centre and retained largely their orientation during central uplift formation." (Jahn and Riller, 2009, p. 221).

Impact-related folds may have strongly brecciated hinges related to high-strain zones. Fold structures in quartzite may contain locally quite voluminous pseudotachylitic breccia in their hinges; PTB are also significant along some radial faults through the collar (Wieland, 2007; WUR, own observations). Tangential and oblique-tangential, impact-related compressional features known from some smaller impact structures are seemingly lacking around the Vredefort Dome. The Granophyre dikes in the core and at the core-collar contact have tangential and radial orientations with respect to the center of the structure, but their emplacement was late in the cratering process, during the extensional phase related to the down- and outward collapse of the central uplift. Lieger et al. (2009) indicated that the voluminous occurrences of pseudotachylitic breccias in the outer core followed tangential and radial orientations and were, thus, consistent with emplacement during the late collapse phase as well. However, it appears that their mapping locations largely follow radial access roads, so that this preferred orientation remains doubtful (especially in the light of a study conducted by one of us, WUR, in the early 1990s – see Reimold and Colliston, 1994). Lieger et al. (2009, 2011) also concluded that this late emplacement of PTB involved intrusion of impact melt and mingling thereof with local material. It must be stated categorically that to date no valid evidence for PTB formation involving an impact melt component has been recorded: all chemical evidence – including that compiled by Lieger et al. (2011) – as well as clasts populations are readily explained by local derivation of PTB melt from one or more lithologies occurring in the immediate vicinity of the PTB sampling sites (see below).

6.1.20.5. Geophysics. The Vredefort structure was formed within a large block of Archean lithosphere, the Kaapvaal craton. Seismic studies (Doucouré et al., 1996; Nguuri et al., 2001) have shown that the crust-mantle boundary occurs – west of the Vredefort structure – at a depth of about 40 km. Prior to the Vredefort impact and associated with the emplacement of the Bushveld Complex some 40 Ma earlier, the Kaapvaal Craton was locally characterized by a high near-surface thermal gradient of up to 40 K per km (Jones and Gibson and Jones, 2002). These authors estimated that the thermal gradient in the target region for the Vredefort impact was of the order of 15–20 K per km.

Comprehensive geophysical analysis has been conducted on the Vredefort Dome and surrounding Witwatersrand Basin for more than 35 years (e.g., Antoine et al., 1990; Green and Chetty, 1990; Corner et al., 1990; Durrheim et al., 1991; and earlier references therein). The Vredefort Dome is characterized by a distinct, albeit weak central gravity anomaly that has been explained in the past, in accordance with the long-lasting controversy about the origin of the dome, as caused by an underlying pluton, a sheet intrusion, uplifted mantle material, or slightly elevated densities of the basement due to uplift of mid-crustal rocks as a consequence of impact. The area of the so-called Transition Zone between outer amphibolite and inner granulite terrain features a strong, negative aeromagnetic anomaly, which has been interpreted as the result of (1) a magnetite-rich crustal layer (Corner et al., 1990), (2) remagnetization due to the impact event (Hart et al., 1995; Cloete et al., 1999; Henkel and Reimold, 2002), or (3) Bushveld magmatic activity (Gibson and Wallmach, 1995). A reflection seismic study by Durrheim et al. (1991) did not provide detail about the geological structure in the interior of the dome, and only a few, discontinuous subhorizontal reflectors of unknown origin were detected. That this signature was only detected over the extent of the dome was interpreted by Henkel and Reimold (1998) as due to plastic deformation as a consequence of the impact. They suggested that the limits of this disturbed zone may correspond to the extent of the transient cavity (the maximum zone of excavation prior to the onset of crater collapse).

Henkel and Reimold (1998, 2002) conducted comprehensive geophysical modeling by integrating gravity, magnetic, and reflection and refraction seismic information. Their perpendicular NW–SE and NE–SW traverses across the entire Witwatersrand basin lead to complete reconstruction of the (now strongly deformed; Fig. 41f, in NW–SE direction) impact structure and suggested an original diameter of 250 km. They also concluded that the internal structure of the Vredefort Dome was incompatible with the historical crust-on-edge geometry model and that mantle material could not be exhumed at the current erosion surface, as suggested earlier (e.g., Tredoux et al., 1999).

Many samples of pre-impact, impact-generated and even post-impact rocks from the Vredefort Dome have revealed surprisingly high Königsberger (Q) values, i.e., ratios between induced and remanent magnetization (Salminen et al., 2009, and references cited therein). These observations have previously been related to the effect of an impact-generated plasma or magnetic field (Carporzen et al., 2005). Considering that the rocks currently exposed on the dome were, at impact time, covered by 5–8 km of now eroded material, the plasma-magnetization hypothesis was greeted with incredulity. Salminen et al. (2009), as part of their comprehensive paleomagnetic and petrophysical investigation of all major lithologies on the Vredefort Dome and beyond, demonstrated that similarly high Q values characterize also rock samples from the surface of the Johannesburg Dome to the northeast of the Witwatersrand Basin. It was debated whether these high Q values could be caused by lightning strike or by thermal or chemical effects in the original mid-crustal setting and related fluid activity. Nakamura et al. (2010) reported detailed magnetic investigations on individual mineral grains from Vredefort samples and concluded that lightning strike was a definite solution for this high-Q problematic. Finally, Carporzen et al. (2012) laid this controversy to rest by demonstrating that the high-Q values are restricted to material currently exposed on surface. They investigated drill core samples collected to 10 m depth from the Vredefort basement. They concluded that "magnetization at depth is consistent with a thermoremanent magnetization acquired in the local geomagnetic field following the impact, while random, intense magnetization and some of the unusual rock magnetic properties observed in surface rocks are superficial phenomena produced by lightning" (Carporzen et al., 2012, p. E01007).

6.1.20.6. Impact-specific deformation phenomena. Since Shand (1916) the Vredefort Dome is known as the type locality for "pseudotachylite". In structural geology, this term is synonymous with "friction melt", numerous occurrences of which are known from fault and shear zones. In impact structures, however, breccias that are generally similar to pseudotachylite can be formed by several processes (e.g., Reimold, 1998; Reimold et al., 1999a; Reimold and Gibson, 2006; Mohr-Westheide et al., 2009; Mohr-Westheide and Reimold, 2010, 2011), including friction melting, shock melting, decompression melting, and combinations of these processes. It has even been suggested by Lieger et al. (2009, 2011) that impact melt injection into fractures in the crater floor, and concomitant assimilation of basement material by such melt could explain the massive volumes of PTB observed in Vredefort (and in the Sudbury structure in Canada). For this reason, it has been advocated to term such breccias pseudotachylitic breccia (PTB) until such time that the actual genesis of such an occurrence has been resolved and a specific genetic term can be applied.

PTBs occur widespread and sometimes in truly massive form in the Vredefort Dome (Figs. 16b, 41a and c) – e.g., Fletcher and Reimold (1989), Killick and Reimold (1990) or Reimold and Colliston (1994). The sizes of Vredefort occurrences are only exceeded by those at the Sudbury structure in Canada. Significant but much less massive PTB is known from the Brazilian impact structure Araguainha, from Siljan in Sweden, and Dhala in India (see above).

The presence of coesite and stishovite in millimeter-wide Vredefort PTB veinlets (e.g., Fig. 41b; Martini, 1978, 1991) strongly suggests that at least some of these thin occurrences represent shock compression melt (i.e., melting caused by decompression *immediately* after propagation of the shock front). Bona fide friction melts are not known from Vredefort, although a frictional component may be associated with many PTB formations. In some cases, displacement of marker bands or veins in the core granitoids by decimeters to a meter suggests friction along minor faults. In some more extensive, so-called network breccias one occasionally finds apparent boundary conditions of straight and always narrow (< a few cm wide) bands of melt breccia that have exactly the same orientation. Whether this could represent the walls of a faulted block does remain hypothetical though. In the absence of the enormous fault zones that would be required for the generation of massive PTB volumes, ample evidence for local formation and accumulation of melts in dilational sites, and as a contribution to PTB from injected impact melt can be excluded on the basis of a huge chemical data base as well as isotopic results that all favor melt formation from local lithologies only (Reimold et al., 2011b, 2013b), decompression melting upon rapid uplift of shock compressed rock at the beginning of the modification phase of cratering is the preferred process for the formation of PTB (Mohr-Westheide, 2011; Reimold et al., 2011b).

PTB *cannot* be considered true evidence of impact, as such breccias – locally massive friction melts – are known from tectonic sites as well (e.g., Fig. 16a). But the extraordinary abundances of PTB at Vredefort and Sudbury may be unique to very large impact structures formed in crystalline basement.

Shatter cones (Figs. 13, 38a, 42a,b and d), a mesoscopic impact-diagnostic fracturing phenomenon, are widespread in the metasedimentary collar rocks and rarely occur in the core of the Vredefort Dome as well. Nicolaysen and Reimold (1999) observed their co-occurrence with narrow-spaced curviplanar joint sets that they termed multipli-striated joint sets (MSJS; Fig. 42c) and that are also known from other impact structures, for example Sudbury (Fig. 42d). Wieland et al. (2006) reviewed the analytical results on Vredefort shatter cones.

For a long time the apparent lack of proper (i.e., not anomalous) shock metamorphic deformation effects fuelled the debate about the origin of the Vredefort structure (e.g., Grieve et al., 1990; Reimold, 1990; Antoine and Reimold, 1988). However, it was known since the 1960s that quartz in Vredefort rocks often contained single or – more rarely – multiple sets of microdeformations (Carter, 1965, 1968). They do not appear like fresh planar deformation features (e.g., Stöffler and Langenhorst, 1994) but rather represent planar fluid inclusion trails or "strings" of microscopic quartz crystals of different extinction behavior. A detailed transmission electron microscopic investigation of these features by Leroux et al. (1994) revealed that these trails are indeed so-called decorated PDFs that indicate the locations of now annealed PDFs. Leroux et al.'s recognition of basal Brazil twin lamellae along such planes produced the first bona fide evidence of shock metamorphism in Vredefort rocks. Since then, shock deformed (planar fractures, as well as granular shock texture) zircon (Kamo et al., 1996; Gibson et al., 1997; Moser et al., 2011; Wielicki et al., 2012) from the Vredefort Dome, and shocked monazite of likely Vredefort origin (Cavosie et al., 2010; Moser et al., 2011; Erickson et al., 2013) have been recorded. Gibson and Reimold (2005) have studied rocks from a radial traverse across the core of the dome and have described a range of textures that likely correspond to shock metamorphism covering the range of shock pressures from >45 GPa at the center of the dome to 10 GPa in the inner collar. Wenk et al. (2005, 2011) investigated Dauphiné twins in Vredefort quartz and discussed whether this could also be used as evidence of shock metamorphism.

The combined evidence of impact-diagnostic shock metamorphic effects in quartz and zircon – and also in monazite found in the erosion debris off the Vredefort Dome along the Vaal River bed to the southwest of the dome, the occurrence of coesite and stishovite, shatter cones, the Vredefort Granophyre containing a meteoritic component, and the extraordinary volume of pseudotachylitic breccia that does not have a match amongst the known tectonic occurrences of pseudotachylite (friction melt), all this leaves no doubt that the Vredefort structure represents a very large impact structure. The age of the impact event is constrained by the dating of Vredefort Granophyre and pseudotachylitic breccia from the dome and the surrounding basin (Kamo et al., 1996; Gibson et al., 1997, 2000; Spray et al., 1995). The accepted age for the impact event is 2023 ± 4 Ma (Kamo et al., 1996).

Ogilvie et al. (2011) reported an experimental shock deformation study in the shock pressure range of 15–60 GPa for a multimineralic gneiss (including quartz, feldspar, garnet, cordierite, biotite, spinel) held at ambient and elevated temperature just before impact. The results of this study allow a direct comparison with the shock metamorphic deformation that was induced into the high-grade granulite gneisses of the Steynskraal Metamorphic Zone (Ogilvie, 2010), besides providing a detailed analysis of the influence of shock impedance contrasts between directly adjacent minerals on shock deformation heterogeneity on both intragranular and intergranular scales. Shock heterogeneity is the result of shock impedance contrast (shock impedance is defined as the product of the speed of sound and the density of the medium concerned) and is evident as shock amplification where shock impedance contrast is greatest and comparative shock suppression where the contrast is least. Ogilvie (2010) meticulously recorded that this pattern is still obvious in the Steynskraal granulites despite post-shock annealing and metamorphic recrystallization. Ogilvie noted this shock heterogeneity at scales down to micrometers, and cautioned to be meticulous when using shock effects to generate

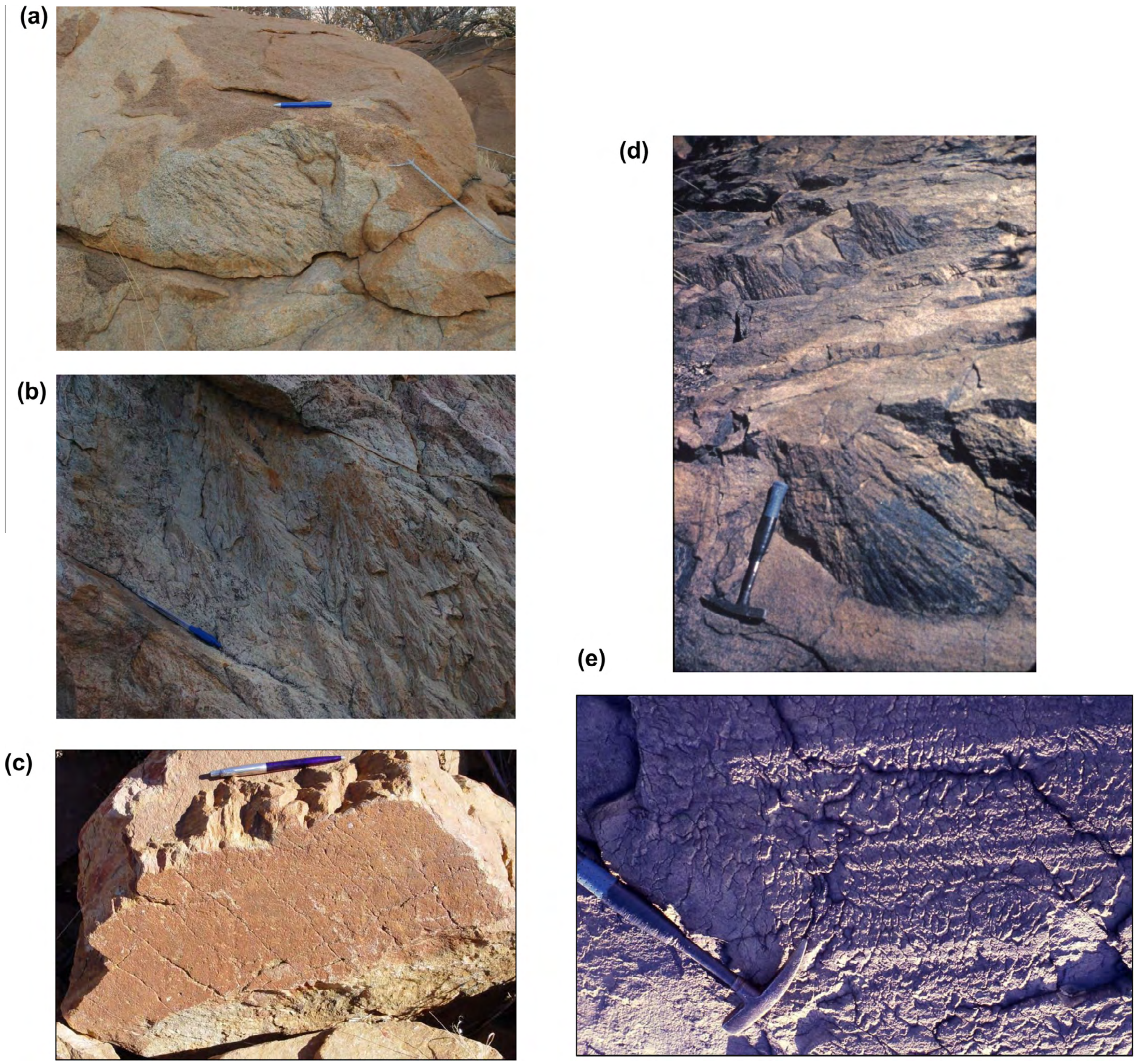


Fig. 42. Shatter cones and shatter cone-related joints. (a) Curved striated fracture surfaces (shatter cones) in medium-grained granite on the property of Kopjeskraal Country Lodge, northwestern core of the Vredefort Dome. Pen for scale 13 cm long. (b) The intricate "horsetailing" effect formed by numerous curved "shatter cones" partially superimposed onto each other. Schoemansdrift bridge, western part of the World Heritage Area of the Vredefort Dome. Pen, for scale, 15 cm long. (c) Several closely-spaced sets of fractures of very different orientations and with striations on the fracture planes (particularly emphasized in the upper part of the image). The sample is from a quartzite boulder found on the alkali granite exposure at Koedoeslaagte, northwestern collar of the Vredefort Dome. These striated fracture surfaces occurring in distinct sets were termed MSJS (multipli-striated joint sets) by Nicolaysen and Reimold (1999). (d) Several, up to 30 cm high shatter cones formed on the prominent (in this image trending from east-northeast to west-southwest) bedding plane in quartzite of the northwestern collar of the Vredefort Dome. Note the distinct relationship between subparallel sets of fractures (e.g., in northeast–southwest and northwest–southeast directions on the top of the large shatter cone in the forefront of the image) and the cone developments. These fracture sets were termed MSJS by Nicolaysen and Reimold (1999) and are found to occur pervasively throughout the collar of the Vredefort Dome, at fracture spacings of less than a cm to several centimeters. (e) Huronian sandstone of the South Range of the Sudbury impact structure, with densely spaced, multiple sets of curviplanar fractures that, in the field, can be recognized as equivalents to the MSJS described from Vredefort.

a regional map of shock distribution, as this work needed to be augmented by independent constraints on shock pressure – such as her own work on the Vredefort granulites and corona mineralogy.

6.1.20.7. Stable isotope studies with relevance for the debate about the genesis of PTB and Granophyre. The first stable isotope study of Vredefort lithologies was reported by O'Neil et al. (1987). They reported $\delta^{18}O$ values for a significant number of core and collar rocks, including granitoids (both from the granulite and amphibolite terranes of the core, and a charnockite from the so-called Transition Zone, see above), pseudotachylitic breccia, and collar metasediment in comparison with equivalent arenites from the Johannesburg area at the northeastern edge of the Witwatersrand Basin. Basic results included that the core granitoids, as referred above at that time classified as OGG and ILG samples, respectively, had similar values, ranging from 7.6‰ to 10‰. The pseudotachylitic breccia samples from the core of the dome yielded values of 7.5‰ to 8.1‰, in all three cases within <0.3‰ of the host granitoids. The lower Witwatersrand collar rocks ranged from 7.2‰ to 11‰, and the Johannesburg samples from 8.2‰ to 10.2‰. Finally, two samples of Vredefort Granophyre gave 7.3‰ and 7.9‰.

Fagereng et al. (2008) significantly added to this data base. They reported $\delta^{18}O$ values for core granitoid gneisses that in their majority lie between 8‰ and 10‰ with a mean of 9.2‰. Data for quartz and feldspar were shown to be consistent with O isotope equilibrium at high temperatures and interpreted as suggestive of minimal interaction with fluids during cooling. The authors applied these data to indicate that the earlier assumption that OGG and ILG represented middle and lower crust (Hart et al. (1990) had to be refuted. Witwatersrand metapelites yielded slightly lower values than measured for the core granitoids, with a mean at 7.7‰. Interestingly, it was observed that comparing data for samples representing a ca. 20 km section through the crust results in a slight gradual increase of $\delta^{18}O$ with depth – the opposite trend of what has been observed normally for crustal rocks. Fagereng et al. (2008) related this finding to their observation that the collar rocks have unusually low values in comparison with meta-sediment from other sites worldwide. They also reported δD values for collar rocks in a range from −35‰ to −115‰. These data are interpreted as being consistent with interaction with meteoric water of δD between about −25‰ and −45‰. Fagereng et al. (2008) suggested that fluid movement through the collar rocks was enhanced due to impact-related deformation resulting in increased permeability.

These authors also analyzed a core granitoid and associated pseudotachylitic breccia sample. The latter had a slightly lower $\delta^{18}O$ than its host rock (7.8‰ and 8.1‰, respectively). δD values are identical and water contents are low and similar. A Schurwedraai alkali granite and associated PTB showed the same $\delta^{18}O$ result (7.0‰ and 7.5‰, respectively). Here, δD values for melt rock and host rock were measured to be similar and water contents were very low. A sample from a PTB vein in metagabbro (epidiorite? [the authors]) of the collar had a $\delta^{18}O$ of 5.7‰ – identical within error with the data for the host. δD values are somewhat different at −74‰ (PTB vein) and −62‰ (host), and water contents are similar. Interestingly they also analyzed two Granophyre samples from the core and obtained identical whole-rock $\delta^{18}O$ values of 7.6‰. This is significantly lower than the values for the surrounding core granitoids (average 9.2‰), but if one considers the contribution of Witwatersrand Supergroup metasediment to the Granophyre melt, with strongly reduced values, this difference is readily explicable. Also, all three studies referred here have indicated that the basement granitoids do show a range of oxygen isotopic values between about 7.5‰ and 10‰. The authors discussed their findings on impact-generated melt against observations by O'Hara and Sharp (2001) for a tectonically generated friction melt (pseudotachylite), in terms of melt generation through partial melting of a precursor material. They concluded that pseudotachylite formation involved the presence of fluid during melting, in an open system. Several authors have in recent decades referred that formation of pseudotachylites and impact-related pseudotachylitic breccias involved preferential melting of hydrous ferromagnesian minerals (such as biotite, amphibole) prior to melting of feldspar or other more refractive minerals (e.g., Reimold, 1991).

A recent publication by Harris et al. (2013) is dedicated to the stable isotope signature of pseudotachylitic breccia, in comparison to host rock, and Granophyre. The stated aim of this investigation was to constrain the mechanism of melt formation and the relationship between "pseudotachylite" and Granophyre. They also report major element chemistry and water contents for their sample suite. In terms of major element compositions they find that "there is an almost complete overlap between the PTB (they call them pseudotachylite) and their host rocks. Unfortunately they follow the Lieger et al. (2011) example, and do not strictly compare the individual PTB-host rock pairs. They point out that Lieger et al.'s data for felsic igneous hosts and the PTB therein cover a wider compositional range than their own, for which they offer no explanation.

The stable isotope results by Harris et al. (2013) can be summarized as follows: Core granitoid gneisses and PTB have almost identical average δD and $\delta^{18}O$ values, at −67‰ and 8.6‰, respectively. Where pairs of PTB and direct host rock have been analysed, these pairs show excellent correlation between $\delta^{18}O$ values (their Fig. 7). Water contents of PTB are very low – which they consider "consistent with isolation of PTB from free water during and since their formation". Vredefort Granophyre is characterized by average δD and $\delta^{18}O$ values of −69‰ and 7.6‰, respectively, as well as similarly low water content. They conclude that "major element and O-isotope composition between the granophyre and the pseudotachylites are not consistent with a simple relationship" (p. 101), thus opposing the claim by Lieger et al. (2011). However, Harris et al. (2013) proceed to conclude that the difference between PTB and Granophyre was due to a higher component of greenstone in Granophyre. This comparison is strange, as locally derived PTB would in nearly all cases not have been involved with a greenstone precursor, *per se*. Thus, it appears that the authors are considering a single source for PTB – not taking into account the entire data base that suggests local generation of PTB from individual host rock types (see above – chemical constraints on PTB formation).

Somewhat confusingly in this paper, this suggestion is then countered by the finding that "[PTB] melt composition is controlled by the immediate surroundings" (Harris et al., 2013). Also, on p. 115 they find that "the isotope data are consistent with local melting of dry rock"; on p. 116 "there was no large-scale connectivity in melt zones at the time of formation", and that "host rock composition is the main control on pseudotachylite $\delta^{18}O$ values". The fact that two groups of PTB of comparatively higher and lower water contents were analysed is interpreted to indicate partial melting of host rock of variable degrees of pre-impact low-T alteration – as opposed to contribution of varying proportions of OH-bearing minerals to the melt.

In summary, a large number of stable isotope data for Vredefort rocks have been generated by now; however, the number of analyses of individual pairs of PTB and host rock is still limited. Also, the situation at the northwestern core-collar boundary, where Granophyre cross-cuts epidiorite and is seriously contaminated by it would lend itself to a detailed stable isotope investigation.

6.1.20.8. Original size of the Vredefort impact structure. Estimation of the original size of the Vredefort impact structure has been approached by different means. Therriault et al. (1997) looked at the regional distribution of shock-metamorphosed and brecciated material, and of shatter cones, in comparison to distribution ranges known from other impact structures. They arrived at estimates for the original diameter of 180 to >300 km. In Fig. 39a some of this regional evidence is shown. Since then, the limit of shatter coning observed by Therriault et al. (1997) was further extended by tens of kilometers by Wieland et al. (2006).

Henkel and Reimold (1998) found that their geophysical modeling results were consistent with an impact structure originally 250–280 km in diameter. In recent years, a diameter of 250 km has been widely accepted. At this size, Vredefort is part of the Big Three – Vredefort, Sudbury in Canada, and Chicxulub in Mexico, the only large impact structures known on Earth that possibly could have been formed as multi-ring impact basins (Grieve and Therriault, 2000; Grieve et al., 2008). These three structures are all preserved to different levels – at Vredefort, a deeply eroded level is currently exhumed, whereas at Sudbury much of the crater fill is retained, including a several kilometer thick impact melt body, and the Chicxulub Structure is completely preserved but also entirely covered by post-impact sediment and, thus, only accessible by drilling and geophysical methods. The numerical modelling of Turtle and Pierazzo (1998) and Ivanov (2005; see Fig. 2c) resulted in comparatively smaller size estimates but did not extend to the end of the modification phase of cratering, during which rim collapse may lead to significant enlargement of an impact structure. Ivanov (2005, 2010, his Fig. 18) showed a model structure for just 400 s after impact, which is ca. 200 km wide and would have been subject to further modification thereafter.

The fact that the original Vredefort impact structure extended over the entire width of the Witwatersrand basin implies that (1) the impact itself was responsible for the preservation of the currently mined ore-bearing strata of the Witwatersrand Supergroup within the confines of the impact ring basin, due to downfaulting along ring faults, akin to what is now observed in "fresh" lunar impact basins, and subsequent cover of these large blocks with impact breccia. (2) Hydrothermal activity would have followed immediately on the impact event, with fluids having circulated as illustrated in Fig. 43 (after Reimold et al., 2005a). This model of impact-induced fluid flow provides a scenario in which the authigenic deposition of gold in the Witwatersrand basin, as for example described by Hayward et al. (2005), could be achieved.

6.1.20.9. Ejecta from the Vredefort impact structure. The Vredefort impact structure is a very old and deeply eroded structure. While remnants of impact breccia – perhaps even including an ejecta component – could well still reside within the deep segment of the ring basin preserved around the northern margin of the Vredefort Dome (see Henkel and Reimold, 1998), the original ejecta blanket around the crater has been entirely eroded. Post-Vredefort sedimentation could involve the deposits of the Waterberg Group of 1700–2000 Ma age (Barker et al., 2006), the closest exposures of which occur to the southeast of Pretoria, more than 200 km northeast of the center of the Vredefort Dome. The bulk of the remaining deposits of the Waterberg Group occur in northern Limpopo Province of South Africa, nearly 500 km from the Vredefort Dome. A single report of an observation of deformation lamellae in quartz grains and in quartz from granite clasts was provided by Callaghan (1986) in a short conference abstract. The accompanying photograph does not allow to identify these lamellae as planar deformation features of shock metamorphic origin. Attempts by one of us (WUR) to obtain access to these thin sections failed, and his considerable efforts to detect possible ejecta beds (spherules or lithic breccias) in long drill cores through the Waterberg sequence from the northern part of Limpopo Province (curated by the Council of Geoscience, Pretoria) were also fruitless.

Chadwick et al. (2001) reported a thick spherule layer from the Ketilidian of southern Greenland. The age of this deposit is rather poorly constrained between ca. 1.8 and 2.1 Ga. These authors suggested that this occurrence could represent remnant ejecta from either the Vredefort or the Sudbury impact. Huber et al. (2011, 2012, 2014) proposed that Vredefort ejecta could have been intersected in drill cores from the Onega Basin, in the Fennoscandian Shield of Karelia, Russia. For a list of known distal impact ejecta layers, see Table 3. The drilled succession there contains spherule beds (Zaonega Formation; Fig. 44), the deposition of which is bracketed by two stratigraphic levels dated at 2050 ± 20 and 1980 ± 27 Ma. This interval includes the formation time of the Vredefort impact structure. Huber et al. (2012) discussed that the spherule occurrences contain an extraterrestrial geochemical signature. These authors discuss some circumstantial evidence, such as an indication from the size of these spherules that the projectile would have been of a size of 20 km, i.e., in the ballpark of the size of a possible Vredefort projectile. Also, the thickness of spherules in the upper part of one of these occurrences would be roughly consistent with the distance of this Karelian occurrence to the paleoposition of the Vredefort site on the Kaapvaal craton (based on an empirical relation by Johnson and Melosh, 2012). They conclude that "several lines of evidence are permissive of the spherules

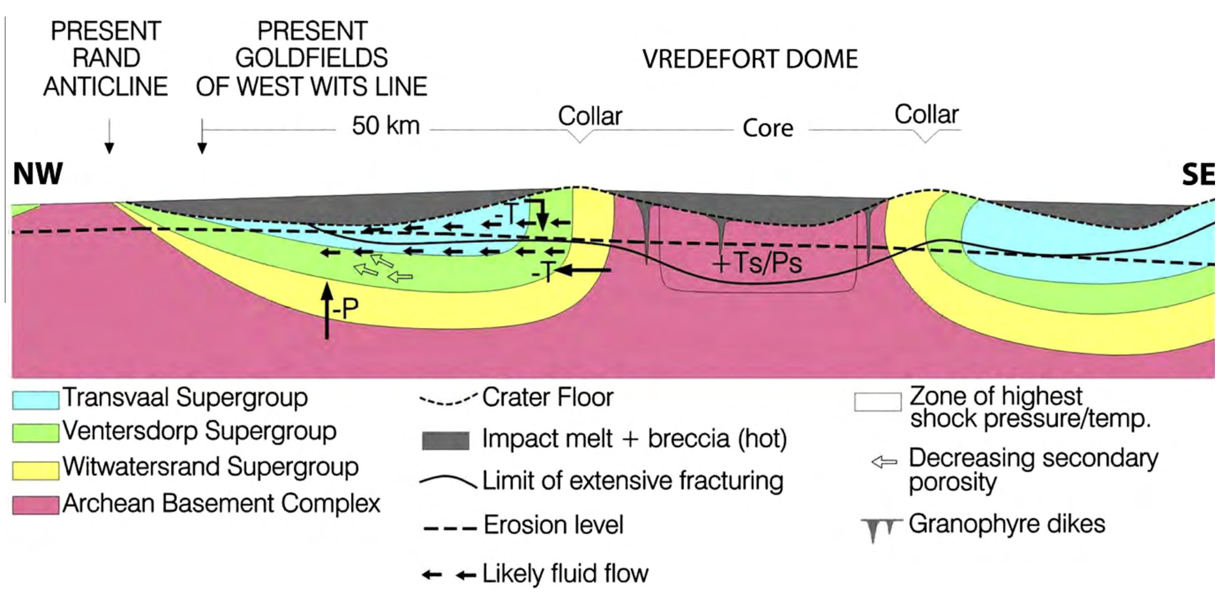


Fig. 43. Schematic model for impact-induced fluid flow in the Witwatersrand basin, as generated by the Vredefort impact. Note, fluid flow is only indicated for the northern, well-preserved part of the ring syncline around the Vredefort Dome. A complex impact structure is shown in cross-section, covered by a thick and hot blanket of impact melt (compare modelling results by Ivanov, 2005). The central uplift volume involves hot (derived from mid-crustal levels) rocks that have been subject to strong shock pressure. Along a profile away from the central uplift, temperature would decrease, and it would also decrease away from the hot impact melt layer above. Outside of the central uplift, in the surrounding supracrustals, lithostatic pressure decreases upward. It is also logical to assume that impact-generated (secondary) porosity would decrease outward. The combination of these effects forces the conclusion that fluid flow would have been primarily laterally away from the Vredefort Dome and enhanced at relatively shallow crustal level. The dashed line indicates schematically the present level of erosion. In summary, fluid flow could have been enhanced to and at the depth of currently mined strata in the outer Witwatersrand Basin (Reimold et al., 2005; see also Hayward et al., 2005a).

Table 3
Currently known distal impact ejecta/ spherule layers (modified after Glass and Simonson (2013)).

Layer	Location	Aggregate thickness[a] (cm)	Age (Ma)	Evidence for an impact origin[b]	Proposed source crater
Australasian microtektites	Indian Ocean; S. China, Philippine, Sulu, Celebes Seas; W. equatorial Pacific; Antarctica	<0.0001 –0.17	~0.8	SM, Mt(L), Ir, Tek	Unknown, but probably in Indochina
Ivory Coast microtektites	E. equatorial Atlantic	<0.001	~1.1	Mt(L), Tek	Bosumtwi
North American mikrotektite layer	Gulf of Mexico, Caribbean, Barbados, NW Atlantic	<0.01–8	~35	SM, Mt(L), Tek	Chesapeake
Cpx spherule layer	Global	<0.0001–0.06	~35	SM, GSph(qt), Mt, NiSp, Cr	Popigai
Nuussuaq spherule bed[c]	Western Greenland	~30–60	~60	GSph(qt), NiSp	Unknown
K–Pg boundary	Global	~0.2–25	~66	SM, Sph(qt, rv), NiSp, Ir, PGE, Os	Chicxulub
Late Triassic spherule bed	SW England, near Bristol	~1.6	~214	SM, Sph(rv)	Manicouagan
Qidong spherule bed	Qidong, South China	Unknown	~372	GSph(L?)	Unknown
Senzeille/Hony microtektites	Belgium	<0.001	~374	Mt(L?)	Unknown
Acraman	South Australia	0.5–3.5	~580	SM, Sph, Ir, PGE, Os	Acraman
Sudbury	Lake Superior region, USA	~25	~1850	SM, Sph(qt,rv), Ir	Sudbury
Grænsesø	Western Greenland	~25	~1850–2130[i]	Sph	Unknown
Zaonega Formation spherule bed[h]	Lake Onega, Karelia, Russia	~10–20	1980–2050	Sph(qt?, rv), Ir, PGE	Vredefort
Dales Gorge[d]	Western Australia	~6	~2490	Sph(qt,rv), Ir, PGE, Cr	Unknown
Kuruman[d,9]	South Africa	~0.5	~2490	Sph(qt,rv), Ir, PGE	Unknown
Bee Gorge	Western Australia	~1–3	~2540	SM, Sph(qt,rv), Ir, PGE	Unknown
Reivilo[e]	South Africa	2–2.5	~2570[i]	Sph(qt,rv),NiSp[j], Ir, PGE	Unknown
Paraburdoo[e]	Western Australia	2	~2570[i]	Sph(qt,rv), NiSp[j] Ir, PGE	Unknown
Jeerinah[f]	Western Australia	0.4–20	~2630	SM, Sph(qt,rv), Ir, PGE, Cr	Unknown
Carawine[f]	Western Australia	~30	~2630	SM, Sph(qt,rv), Ir, Cr	Unknown
Monteville[f]	South Africa	~6	<~2650>~2588	SM, Sph(qt,rv), Ir, PGE	Unknown
S5	South Africa	?	~3230	Sph	Unknown
S4	South Africa	12	~3243	Sph(qt?,rv), Ir, Cr	Unknown
S3	South Africa	~30	~3243	Sph(qt,rv), NiSp, Ir, Cr	Unknown
S2	South Africa	~10–70	~3260	Sph(qt), Ir(?), Cr	Unknown
S6	South Africa	?	~3330	Sph	Unknown
S7	South Africa	?	~3410	Sph	Unknown
S1[g]	South Africa	~6	~3470	Sph(qt,rv), Ir(?)	Unknown
Warrawoona (Apex Basalt)[g]	Western Australia	~5	~3470	Sph(qt)	Unknown

[a] Aggregate thickness estimated from layer thickness and percent spherules.
[b] Cr = Cr isotope data, GSph = glass spherules, GSph(L) = glass spherule with lechatelierite, GSph(qt) = glassy spherule. with quench texture, Ir = Ir anomaly, Mt = microtektites, Mt(L) = microtektites with lechatelierite, Os = Os isotope data, NiSp = Ni-rich spinels, PGE = platinum-group elements with close to chondritic ratios, SM = shock-metamorphosed rock. and mineral grains, Sph = altered spherules, Sph(qt) = altered spherules with quench textures, Sph(rv) = altered. spherules with relict vesicles, Tek = associated with a tektite strewn field.
[c] Also referred to as Disco spherules.
[d] These two spherule layers may be from the same impact.
[e] These two spherule layers may be from the same impact.
[f] These three spherule layers may be from the same impact.
[g] These two spherule layers may be from the same impact.
[h] Three occurrences in the FARDEEP drilling intersections of the Zaonega Formation (Onega Basin) – Huber et al. (2014).
[i] Bruce Simonson, Oberlin College, pers. comm., 2013
[j] Goderis et al. (2013).

being derived from the Vredefort impact, though none of these lines of evidence is conclusive.'' Better chronological constraint for the spherule beds from Karelia is required, before a relationship to the Vredefort impact could be confirmed.

6.2. Impact glasses and tektites

6.2.1. Libyan Desert Glass (LDG), Egypt

Libyan Desert Glass (LDG) is an enigmatic type of natural glass, which occurs in an ~2500 km^2 strewnfield located between sand dunes of the southwestern corner of the Great Sand Sea in western Egypt (Fig. 45a). The glass occurs as centimeter- to decimeter-size irregular and strongly wind-eroded pieces (Fig. 45b). The total quantity of the glass present has been estimated at 1.4×10^9 g, with a much larger original mass assumed (Barnes and Underwood, 1976; Diemer, 1997). The glass is very silica-rich at about 96.5–99 wt% SiO_2 (Fudali, 1981) and shows a limited variation in major and trace element abundances (Koeberl, 1997). Some cristobalite inclusions occur, but otherwise the LDG is perfectly glassy. Although the origin of LDG is still debated by some workers, an origin by impact seems most likely. Fröhlich et al. (2013) noted, based on a detailed Fourier-transform infrared spectrometry study, clear differences between LDG and other natural glasses such as fulgurites or biogenic silica.

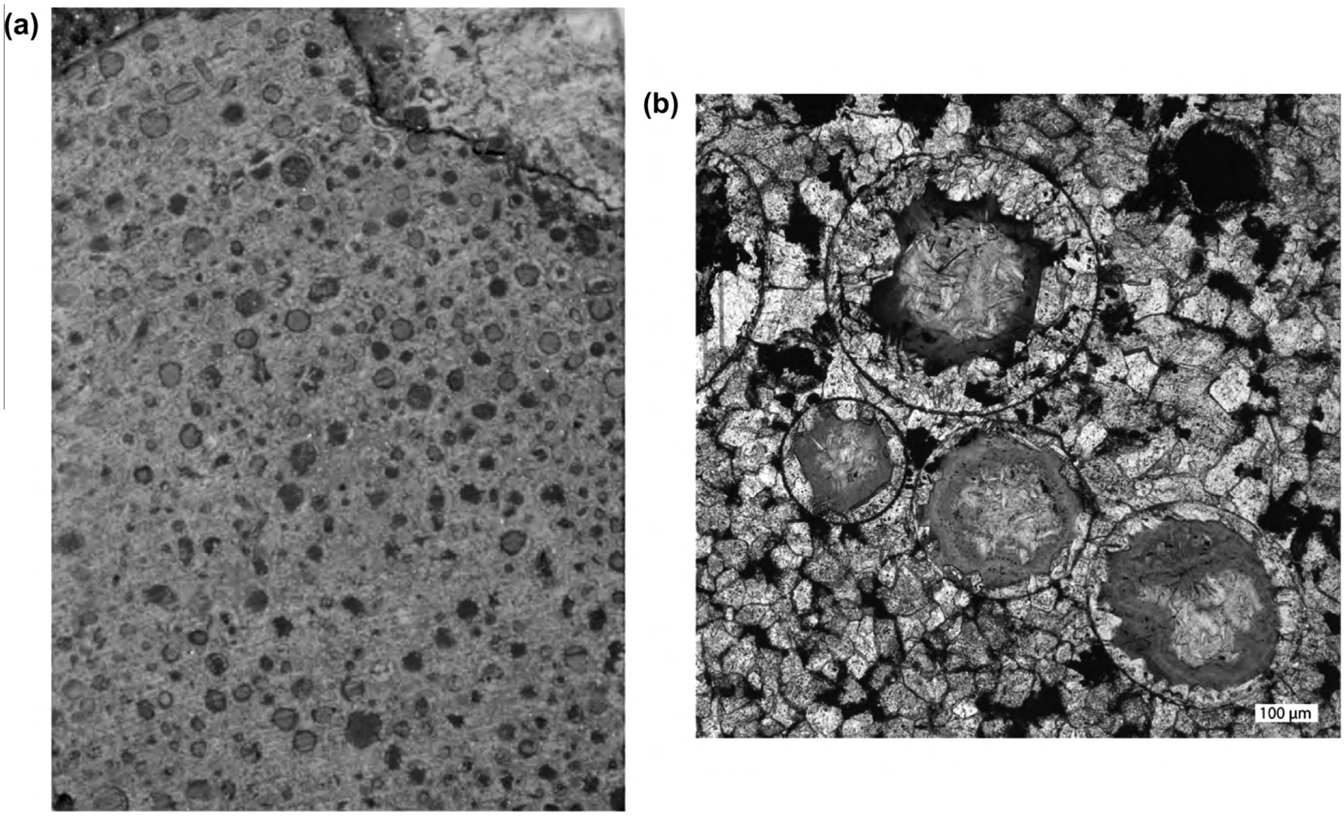


Fig. 44. Spherules of the Kola Peninsula occurrence in the ICDP FAR-DEEP drill core 13A. (a) Sample from 27 m depth. Spherules are approximately 1 mm in diameter. While most spherules have a spherical shape, some elongated, teardrop or dumbbell shaped spherules are apparent. (b) Plane polarized light image of spherules from 27.35 m depth in the same core. Each of the spherules is clearly outlined by a pyrite and apatite rim, separating them from the dolomite matrix. The outer edges of the spherules are altered to calcite, but internal textures are preserved by phyllosilicates in the interior. Scale bar = 100 µm wide. Both images were kindly provided by Matthew Huber, formerly University of Vienna.

There are, however, some differences to "classical" impact glasses, which occur in most cases directly at or within an impact crater. Evidence for an impact origin includes the presence of schlieren and partly digested mineral phases, lechatelierite (a high-temperature melt of silica), and baddeleyite, a high temperature breakdown product of zircon (Kleinmann, 1969). The rare earth element abundance patterns are indicative of a sedimentary precursor rock, and the trace element abundances and ratios are in agreement with an upper crustal source (see, e,g., Koeberl, 1997).

The age of the LDG was determined by K–Ar and fission-track methods. While it is possible to calculate a K–Ar age for LDG (58.3 ± 16.4 Ma; Matsubara et al., 1991), these values suffer from the very low K content of the glass, and other methodologic problems (Horn et al., 1997). Better suited for an age determination of the LDG is the fission-track method, which yielded an age of 29.4 ± 0.5 Ma (Storzer and Wagner, 1977). This age was confirmed with a result of 28.5 ± 0.8 Ma by Bigazzi and de Michele (1996).

There are strong indications for the presence of a meteoritic component in dark streaks or layers of the desert glass (e.g., Barrat et al., 1996). This was confirmed by Koeberl (2000) in an osmium isotopic study. Crustal strontium and neodymium isotopic values (Schaaf and Müller-Sohnius, 2002) exclude a significant mantle component; thus, the osmium abundances and isotopic values confirm the presence of a meteoritic component in LDG.

Giuli et al. (2003) studied the iron local environment in an LDG sample by means of iron K-edge high resolution X-ray absorption near edge structure (XANES) spectroscopy to obtain quantitative data on the iron oxidation state and coordination number in both the iron-poor matrix and iron-rich layers. They found that in the layers with higher iron content, the iron occurs in a more reduced state, which suggests that some or most of the iron in these layers may be directly derived from the meteoritic projectile and that it is not of terrestrial origin.

Greshake et al. (2010) studied a variety of dark streaks that represent a different type of schlieren compared to those which contain a meteoritic component (on the hand specimen scale, there is a slight difference in color). Their findings suggest that LDG formed during a short high-temperature event. Melting of aluminum-rich orthopyroxene bearing target material, which then formed the dark schlieren, seems to suggest an asteroid impact rather than a near-surface airburst. This agrees with the findings of shocked quartz by Kleinmann et al. (2001). Magna et al. (2011) noted that Libyan Desert Glass is characterized by high δ^7Li at $\geqslant$24.7‰, which may represent the previous fluvial history of parental material that was perhaps deposited under lacustrine conditions or in coastal seawater. Finding agreement with the strontium and neodymium isotopic data, which suggest a Pan-African age of the parent material for the LDG, Longinelli et al. (2011) measured δ^{18}O values of bulk rock and quartz from intrusives of Pan-African age and the results obtained were compatible with their values obtained for LDG samples. Thus, geochemical measurements have been essential to determine the origin of the LDG as some type of impact glass, and have also given valuable indications regarding the source material.

Kramers et al. (2013) reported on a diamond-bearing, centimeter-sized diamond-bearing rock fragment that was found in the vicinity of the LDG strewn field. Based on the carbon isotopic composition and anomalous contents of some noble gases (especially Ar), these authors concluded that the fragment is most likely of extraterrestrial origin, and they hypothezised that it could be a fragment of a comet. Stretching the hypothesis even further, they suggested that this could be a remnant from the airburst that

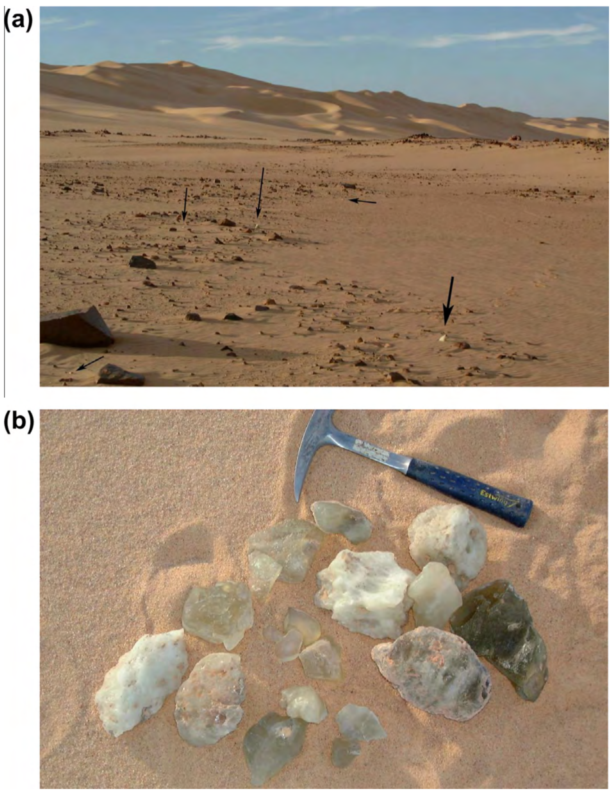


Fig. 45. Libyan Desert Glass: (a) Field occurrence in the corridors between large sand dunes at the southern extension of the Great Sand Sea, Libyan Desert, south-western Egypt. Chunks of glass are marked by arrows. (b) Selection of differently sized and colored Libyan Desert Glass specimens recovered in the field by C Koeberl in 2003, with hammer for size.

created the LDG. There are several leaps of faith and implausibilities in this story. No large chunks of diamonds have been found in comets. It is by no means sure how the LDG was formed – involvement of an airburst has been suggested but is not the most straightforward explanation. No other signatures of an extraterrestrial origin of this fragment were reported. Remnant fragments of the impactors that formed impact structures have not been found on the surface in or around impact craters, because they would quickly fall prey to erosion – if they ever survived the impact. There is no connection between the diamond-bearing rock fragment and the LDG except that they occur in the same region. No causal connection is suggested by any of the available data. Thus, while Kramers et al. (2013) present interesting data on an unusual object, their suggestions are far-fetched and are so far unlikely to have any connection with LDG.

6.2.2. Dakhleh Glass, Egypt – Evidence for a young impact event

The Dakhleh Oasis region of central-western Egypt has been the focus of geo-archeological activity since about 30 years ago (Churcher and Mills, 1999). It was in the course of this project that a lag deposit of dark, glassy material on surface and within Pleistocene lacustrine deposits was discovered (Osinski et al., 2007). Archeological evidence, namely Earlier Stone Age through Middle Stone Age occupation has constrained the age of this layer to ~350–2100 ka. Osinski et al. (2007) presented a first comprehensive mineralogical-geochemical analysis of these glasses that are known locally as Dakhleh Glass. An impact event at that time in this region that displays ample evidence of habitation since >400,000 years before the emergence of *Homo Sapiens* would have left a catastrophic mark on the early inhabitants (Smith et al., 2009).

Dakhleh Oasis (centered at 25°30′N/29°07′E) is a 1200 km^2 wind-ablated depression in the central Western Desert of Egypt, just south of the Libyan Escarpment. The depression is floored by Cretaceous rocks and filled by Paleocene to Pleistocene sediments covered by Holocene playa and eolian deposits. Dakhleh glass (Fig. 46) was found at six locations in the oasis region. The glass is typically highly vesiculated and in hand specimen of greenish-grey to black color. In situ glass comprises masses of <5–30 cm size. Lag glass deposits involve dispersed clusters of irregular and flattened, up to 30 cm sized, masses. Glasses contain abundant microcrystals. The chemical composition of the glasses is highly variable, ranging from 90–100 wt% SiO_2 to CaO- and Al_2O_3-rich

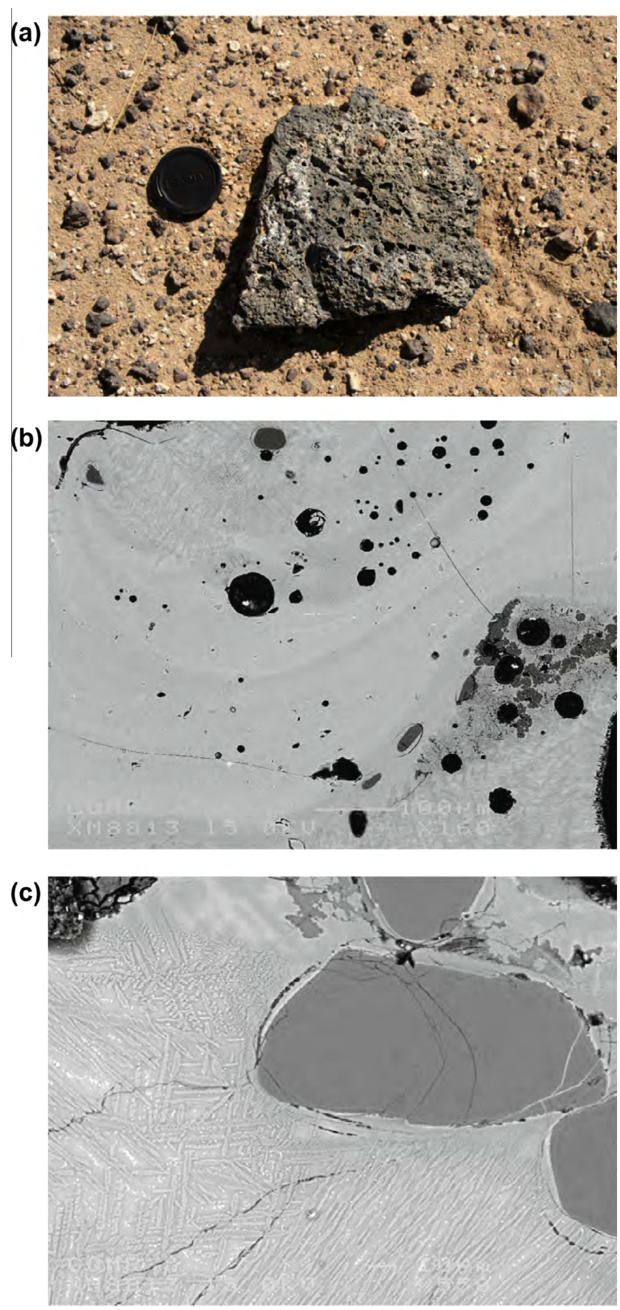


Fig. 46. Dakleh glass: (a) Large melt specimen in the field. Lens cap for scale. (b) Backscattered-electron image showing bright glass, grey quartz clasts, and clino-pyroxene crystallites. Width of image ca. 700 µm. (c) Backscattered-electron image of partially crystallized melt and several grey quartz clasts. All three images were kindly provided by Gordon Osinski (University of Western Ontario, Canada).

varieties. The high-silica glasses represent enclaves within host glass. Compositions vary from sampling site to sampling site, and from specimen to specimen.

Osinski et al. (2007) referred to two ^{40}Ar–^{39}Ar step-heating dating experiments that yielded internal isochrons corresponding to ages of 282 ± 121 ka and 119 ± 45 ka. As evidence for an impact origin of these glasses the following arguments were presented: (1) The chemical composition is different from that of volcanic glasses. (2) There are no volcanic features in the area. (3) The high-silica glass enclaves are interpreted as results of melting of sandstone and related to lechatelierite inclusions in impact glasses. The presence of this phase, which requires formation temperatures in excess of 1700 °C, rule out that the glasses were formed by burning of vegetation or organic-rich sediment. (4) The age of the glass precludes the possibility that these glasses could be of anthropogenic origin.

Bona fide shock evidence has not been found in inclusions in Dakhleh glass. The authors refer to fractured quartz grains in Pleistocene sediment that resemble such deformed quartz from the source area of the Libyan Desert Glass – but they do not show such deformation. In conclusion, there is circumstantial but not direct evidence that Dakhleh Glass could be of impact origin. Osinski et al. proceeded to discuss the lack of a source crater and the possibility that the perceived impact event could have been an aerial burst of an extraterrestrial projectile. Implications on possible target rocks/precursor materials for glass production from the chemical compositions of glasses are discussed in terms of the lithostratigraphy in the region of Dakhleh oasis.

In a subsequent publication, Osinski et al. (2008) provided more detail about the occurrence of these glasses, their petrography and chemistry, and possible origin. By then, the glasses had been recognized over a 400 km^2 area. Largest specimens detected during renewed field work measure 50 cm across. New chemical analyses reveal compositions with up to 25 and 18 wt% CaO and Al_2O_3, respectively. The authors confirm the presence of lechatelierite in the glasses, as well as of burnt sediment, and argue that the presence of clasts and spherules is inconsistent with the character of glass formation by known terrestrial processes. They follow up on their initial suggestion that the Dakhleh Glass could be the result of an airburst of an extraterrestrial projectile by evaluating recent numerical models of airbursts (Boslough and Crawford, 2008), and conclude that this process is currently the best option regarding Dakhleh Glass genesis. They conclude that glass produced by such events "should ... be more common in the rock record than impact craters, assuming that the glass formed in a suitable preserving environment."

Renne et al. (2010) reported the results of several detailed argon step-heating experiments, which yielded a preferred isochron age of filtered data (i.e., excluding data seemingly derived from undegassed clasts) of 145 ± 19 ka, in keeping with the archeological constraints. Applying the calibration corrections recommended by Kuiper et al. (2008) yields a revised age of 146 ka. An impact at that time would have had catastrophic consequences for local inhabitants.

6.2.3. Ivory Coast tektites

Tektites are distal glassy ejecta from hypervelocity impact events; so far only four geographically extended tektite strewn fields have been identified. One of them is centered in the West African country of Ivory Coast. There they were first reported in 1934 (Lacroix, 1934) from a small area of about 40 km radius within the Ivory Coast (Côte d'Ivoire). Although no meteorites have been found in Liberia or Sierra Leone, tektites that are part of the Ivory Coast Tektite strewn field were at some time reported from eastern Liberia, some 780 km from their source, the Bosumtwi impact structure (Preuss and Meyer von Greyhold, 1968). However,

Fig. 47. Image of a typical Ivory Coast tektite (sample IVC-2069; cf. Koeberl et al., 1997b for details about this sample).

Preuss (1969) refuted this claim and presented evidence that this alleged find actually represented transported tektites from the Australasian strewn field. Microtektites were found in deep-sea cores off the coast of Western Africa (Glass, 1968, 1969) and were interpreted as being related to the tektites found on land (Fig. 47). The geographical distribution of microtektite-bearing deep-sea cores has also been used to determine the extent of the strewn field (Glass and Zwart, 1979; Glass et al., 1979, 1991).

A variety of arguments was used to conclude that Bosumtwi is the likely source crater for these microtektites/tektites, including similar chemical compositions (Schnetzler et al., 1967; Jones, 1985) and isotopic characteristics for the tektites and rocks found at the crater (e.g., Schnetzler et al., 1966; Shaw and Wasserburg, 1982), as well as similar ages of tektites and Bosumtwi impact glasses (e.g., Gentner et al., 1964). Koeberl et al. (1998b) found that the oxygen isotopic composition of the metasedimentary rocks and a granite dike at the crater agree fairly well with those of the tektites, and showed that in both a $^{87}Sr/^{86}Sr$ vs. 1/Sr plot and an ε_{Sr} vs. ε_{Nd} diagram, the tektites plot within the field defined by the metasedimentary and granitic Bosumtwi crater rocks. The available geochemical data (Koeberl et al., 1997a, 1998b) support the conclusion that the Ivory Coast tektites were formed from the same rocks that are currently exposed at the Bosumtwi crater. Precise fission track and step-heating ^{40}Ar–^{39}Ar dating on both Ivory Coast tektites and Bosumtwi impact glass established the reliable and currently accepted age of 1.07 ± 0.05 Ma for the Bosumtwi impact event (Koeberl et al., 1997a). Serefiddin et al. (2007) used Be-10 cosmogenic nuclide measurements in Ivory Coast tektites and soils from Bosumtwi to provide yet another link between the crater and the tektites.

6.3. Spherule layers – remnants of distal impact ejecta

In the absence of shock metamorphism in the oldest rocks on Earth (not even shocked zircon crystals have been observed to date amongst the oldest Archean zircon populations investigated), we need to look at "younger" rocks to find the first evidence of terrestrial impact. The oldest known impact structure is the 2.02 Ga Vredefort structure in South Africa – thus, there is a 2.5 billion year lack of impact-crater evidence on Earth. However, the rock record of impact events begins about 400–500 million years after the end of the Late Heavy Bombardment (LHB; see Section 1), in the form of *distal ejecta layers* that are characterized by accumulations of impact spherules (see reviews by Simonson and Glass, 2004; Simonson et al., 2004; Glass and Simonson, 2012, 2013). Impact spherules can represent either melt droplets directly ejected from the crater or formed by condensation from vapor clouds. Glass and Simonson (2012) give the following criteria for the recognition of possible impact spherule layers:

1. The presence of sphere-shaped particles (preferably abundant) among the sand-sized fraction of a layer.
2. The absence of comparable spherules in surrounding strata.
3. The presence of rare spherules shaped like dumbbells or teardrops, indicating that they were formerly molten.
4. The presence of vesicles inside the spherules, again proving that they were molten.

Once such a layer has been recognized in the rock record, it is required to prove the impact association through establishing petrographic or chemical evidence for impact (see Section 1). A list of currently known distal impact ejecta layers is given as Table 3, and macroscopic and microscopic images of two spherule accumulations and thin section imagery thereof are shown in Fig. 48a–d (courtesy of B. Simonson).

6.3.1. Mesoarchean spherule layers in South Africa

Four distinct silicate spherule horizons in the Barberton Greenstone belt, South Africa (designated S1–S4), with ages between about 3.5 and 3.2 Ga, are most probably of impact origin (e.g., Lowe et al., 2003; Glass and Simonson, 2013). It is, however, not yet clear whether S3 and S4 stratigraphically located close to the transition from the Onverwacht Group to the Fig Tree Group could represent

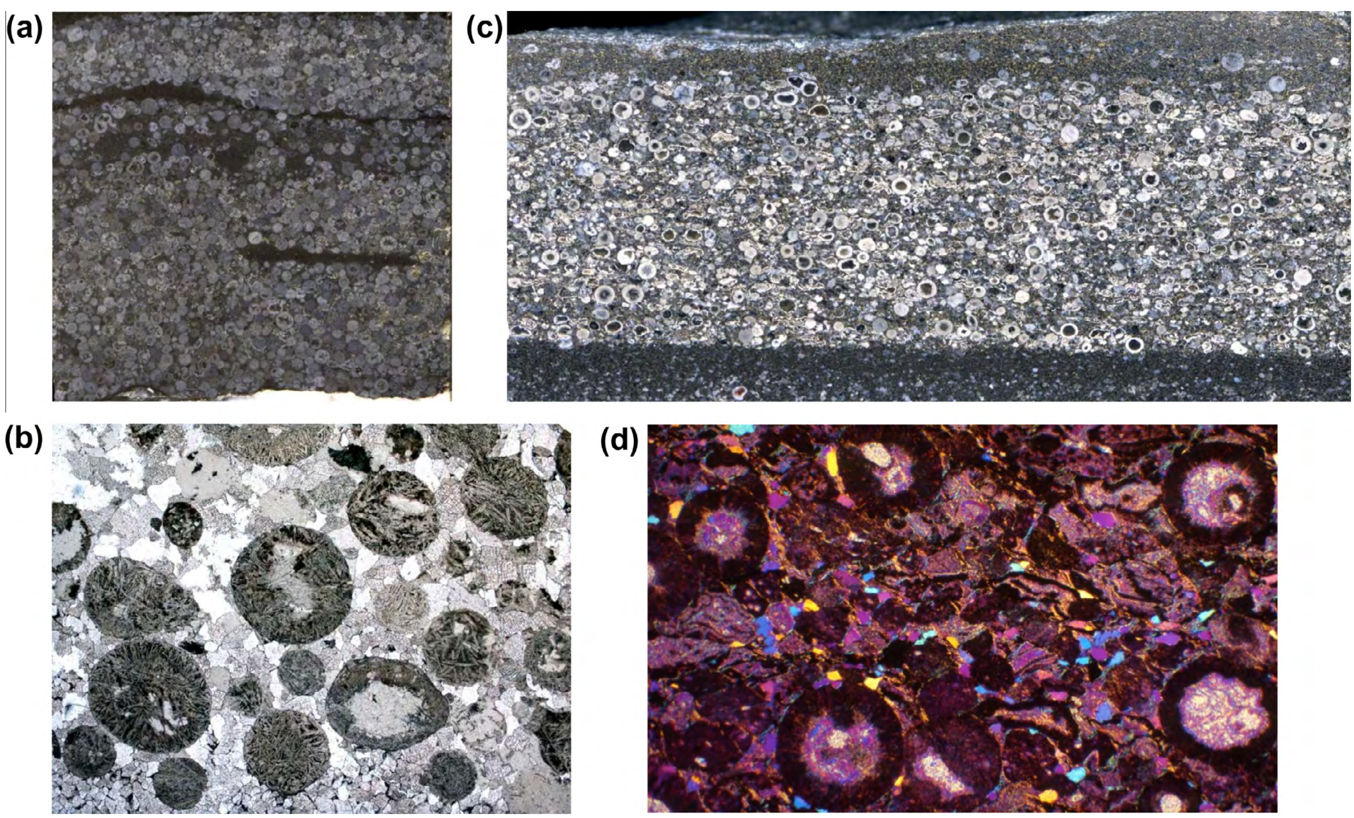


Fig. 48. Impressions of the Reivilo and Monteville spherule layers of the Transvaal Supergroup (South Africa) – courtesy of Bruce Simonson (Oberlin College). (a) Basal ~2.5 cm of the Reivilo spherule layer from a core drilled in the Griqualand West Basin (South Africa). The impact spherules (gray spherical particles), known as microkrystites, were partially crystallized in flight, deposited in relatively shallow water (above wave base), then reworked by unusually high-energy waves and/or currents, possibly generated by the impact itself. The thin, elongated black bodies are intraclasts of shale ripped up by the event that created the layer. This part of the layer is overlain by finer carbonate-rich sand as part of a graded bed. The impact occurred around 2.57 Ga, but the location of the source crater is unknown. This sample would have originally consisted of ca. 60% terrestrial target material and 40% extraterrestrial material based on its 176 ppm iridium content (Simonson et al.,, 2009). Scale: both the horizontal and vertical dimensions of core shown are ~2.5 cm. (b) Reivilo microkrystites (plane polarized light) rich in thin, elongated psuedomorphs that were originally skeletal crystallites of both plagioclase and olivine (and possibly pyroxene). These crystallites were preferentially replaced by K-feldspar and phlogopite, respectively. All other primary phases were also replaced by these minerals as well as sparry carbonate, with the exception of small Ni-rich spinels (Goderis et al., 2013; not visible at this scale). The shapes of these spherules are unusually well preserved because they were protected from compaction by an abundance of sparry carbonate cement. Scale: the diameter of the large circular spherules nearest to the center of the image is ~1.1 mm. (c) Impact spherule-rich zone ~1.2 cm thick in the upper part of the Monteville spherule layer from a core drilled in the Griqualand West Basin (South Africa) where it is ~10 cm thick overall (compare Simonson et al., 2000a,b). Many spherules show a concentric structure with a rim (white) of radial-fibrous fans growing inwards from the outer edge and a core (dark) of clearer sparry crystals filling vesicles and/or replacing residual glass cores. Other spherules were flattened by compaction. The spherules were deposited in relatively deep water (below wave base), then reworked by unusually high-energy waves and/or currents, possibly generated by the impact itself. The finer, darker beds enclosing the spherule-rich zone are rich in quartzose sand transported anomalously far offshore during the deposition of the layer; they also contain rip-ups of early diagenetic pyrite (golden) from the pyritic shale substrate (Simonson et al., 1999). The impact took place around 2.63 Ga, but the location of the source crater is unknown. The spherules originally consisted mainly of terrestrial target material with ca. 1% impactor based on an iridium content of ~5 ppm (Simonson et al., 2000a,b). (d) Impact spherules of the Monteville spherule layer (between crossed polarizers with gypsum plate inserted), with many of them having central spots that appear to be replaced glass cores, although a few vesicles are apparent (for example, the spherule just left of center along the upper edge and the small circular spherule near the right corner on lower edge both contain vesicles filled with sericite). Many spherules were flattened by both brittle and ductile processes during compaction. All primary phases were replaced, primarily by K-feldspar and sericite. The interstitial material is largely sericite, probably a combination of compacted rip-up clasts and cement. The small, non-circular grains (clear/blue-yellow) are very fine to fine quartz sand grains. Scaling info: The diameter of the largest circular spherule (near the lower left edge) measures ~1.1 mm.

the same event. Several other possible impact spherule layers, with textures highly reminiscent of those in S1–S4, have been proposed in the Barberton stratigraphy (e.g., Lowe and Byerly, 2010; Table 1 in Glass and Simonson, 2012, and Hofmann et al., 2006).

The spherules are mostly spherical particles, up to a few millimeters across, but other forms (dumbbell, teardrop, ovoid to elongated with round edges, shard shapes, or irregular shapes) have also been observed. They are found either in dense accumulations in narrow beds (up to a few decimeters thick) sandwiched between shale, BIF and chert layers, or in vastly thicker strata as highly diluted spherule-bearing beds with significant intermittent clastic debris (chert, shale, etc.) of sand to meter size that often have angular geometries. This clastic material has been interpreted to be derived from tsunami-style events, possibly in the direct aftermath of an impact event.

Those intersections (underground mine exposures and drill core intersections) that contain abundant spherules in limited matrix are composed essentially of quartz, secondary phyllosilicates, carbonate (calcite, siderite), sulfate (e.g., barite), and always more or less sulfide and arsenates. Spherules generally contain a lot of secondary K-feldspar. A variety of spherule textural types has been distinguished (e.g., Krull-Davatzes et al., 2012). An important trace component is formed by spinel, which may preserve primary geochemical information. Ni-rich spinel (Schmitz, 2013) has been extensively researched for its extraterrestrial, genetic significance. In this context it should be referred to the work by Reimold et al. (2000c), who noted that some spinels in the Barberton spherule layers are quite enriched in Zn, besides Ni that had been used previously to infer spinel formation from impact-generated vapor plumes (Krull-Davatzes et al., 2006; Ebel and Grossman, 2005). Coarse-grained, heterogeneous layers have been interpreted as reworked deposits and are thought to reflect high-energy depositional events in otherwise low-energy, quiet water depositional environments. Lowe (2013) discusses circumstantial geological evidence involving spherule occurrences in chert dykes in the stratigraphic vicinity of S3/S4 as indicative of crustal fracturing and chert dike formation as a consequence of large-scale impact.

The original mineralogical and chemical composition of the spherules has been almost completely changed by alteration. Spherules do, however, still commonly show original (partial) crystallization textures, which often take the form of lath-like crystals of K-feldspar or other pseudomorphing secondary minerals that clearly indicate that crystallization commenced on the rim of spherules and progressed inward. Where crystallization has replaced an entire spherule, textures may resemble those of barred chondrules in meteorites. Glass and Simonson (2013) have provided a comprehensive volume about distal impact ejecta, reviewing the geological contexts of these occurrences, and the mineralogical and geochemical background on them. The additional spherule layers of still uncertain impact origin require further detailed petrographic and chemical analysis. There are other spherule accumulations known from the Barberton Greenstone Belt that, in part, are considered volcanic acretionary lapilli beds.

Samples from the Barberton spherule layers sometimes show extreme enrichments in the platinum group elements (PGE) – in some cases exceeding the PGE abundances found in chondritic meteorites by factors >2 (also refer to caption to Fig. 48a). This is unlike modern impact ejecta deposits (for example, those at the Cretaceous–Tertiary boundary, or in the late Eocene, see, e.g., Montanari and Koeberl, 2000, for a review), which show siderophile element and PGE abundances corresponding to small fractions of meteoritic abundances. This caused some questions regarding the initial impact interpretation. Spherule layers (3.48 Ma old; Byerly et al., 2002) from the Barberton Greenstone Belt have been interpreted as the result of large asteroid or comet impacts onto the early Earth (see review by Lowe et al. (2003)). The extreme enrichments in the PGE (e.g., Lowe et al., 1989; Kyte et al., 1992; Koeberl and Reimold, 1995; Reimold et al., 2000a,b,c) and other inconsistencies (including questions about stratigraphic positioning of some of the spherule layers and possible tectonic duplication) initially caused Koeberl and Reimold (1995) to question the impact interpretation. A more recent discussion of this is found in Hofmann et al. (2006), who also emphasized that the enormous chemical enrichments of meteoritic tracer elements has still not been resolved.

It has been noted that there is a distinct correlation between the abundances of iridium and arsenic, a very mobile element, in samples from the Barberton spherule layers, all of which were subjected to pervasive transformation into secondary mineral assemblages. The iridium-arsenic relationship might indicate remobilization of both elements; this means that the PGE signature in these samples would not be primary (e.g., Koeberl and Reimold, 1995; Reimold et al., 2000a,b,c) – which, in turn, explains the super-chondritic enrichment factors.

Shukolyukov et al. (2000, 2002) and Kyte et al. (2003) reported chromium isotopic anomalies in samples from three of these layers that seem to support the presence of an extraterrestrial component. A detailed summary, including new chromium isotopic data, and a discussion about the correlation between spherule layers in Australia (see next paragraph) and South Africa, was given by Simonson et al. (2009). New petrographic and geochemical data are also given by Krull-Davatzes et al. (2012). Further detailed analysis, especially for possible presence of PGE enrichments due to meteoritic components, is required. Further analysis is required to investigate the possible secondary nature of Zn-rich spinels.

6.3.2. Neoarchean spherule layers

Other occurrences of unusual spherule layers were reported by Simonson (1992) from the Hamersley Basin in Western Australia. On the basis of similarities to microtektites and mikrokrystites, Simonson (1992) interpreted the spherules as having formed in an impact event and having been redeposited in a sediment gravity flow. Later, three additional spherule-bearing layers were found in the Hamersley Basin sequence, which were also interpreted to be of impact origin (e.g., Simonson et al., 1998). Simonson et al. (2000a,b) also reported the discovery of three similar spherule layers (ca. 2.3–2.6 Ma) in the Reivilo (Fig. 48a and b) and Monteville (Fig. 48c and d) formations of the Transvaal Supergroup in South Africa (Table 3), one of which might be correlated with one of the Australian layers (e.g., Simonson and Hassler, 1997; Simonson et al., 1999, 2004, 2009; Rasmussen and Koeberl, 2004). The Neoarchean spherules have the same petrographic characteristics as the older spherule beds, with K-feldspar pseudomorphs that appear to be devitrification features of quenched melt spherules, and morphologies of spherules that include spheres, teardrops, and dumbbells. Other than their petrographic distinctiveness, the spherules display unusually high Ir abundances and ratios of PGE that are consistent with a chondritic impactor (Simonson et al., 1998, 2000a,b).

As mentioned above (Vredefort ejecta section), Huber et al. (2011, 2014) suggested that a spherule layer intersection (Fig. 44) in a drill core from ca. 2 billion year old strata of the Kola Peninsula represent ejecta from a large impact event that occurred around the time of the 2.02 Ga Vredefort impact event. The so far available age data bracketing the spherule layer deposition time would, however, have to be improved in order to be able to confirm this relationship. Similarly, Chadwick et al. (2001) reported impact spherules from Grænsesø, Greenland, from samples originally identified to contain ooids. The age bracketing of these spherules is less constrained than those found by Huber et al. (2012, 2014), but nonetheless does include the age of the Vredefort impact event.

6.3.3. Shock metamorphism and source craters

Until recently, no shocked minerals have been reported to be associated with any of these spherule layers. It was suggested that this is because the impacts occurred into oceanic crust, which has little or no quartz, and whatever else there was in terms of shocked minerals had long been destroyed by alteration (Simonson et al., 1998). Rasmussen and Koeberl (2004), however, were able to identify one shocked quartz grain in a sample from the 2.63 Ga Jeerinah impact layer of the Hamersley Basin, Australia; this is so far the only evidence of diagnostic shock features in a spherule bed. Spherule layers are usually identified based on the morphology of the individual spherules, the abundance and ratios of PGE, and Cr-isotope anomalies; however, no definitive criteria for the identification of Archean impact deposits have been established, and thus far no proximal ejecta have been discovered in the Archean or Proterozoic deposits. For none of the South African (Barberton and Monteville) or Australian spherule layers have source craters been found, and given the scarcity of the early Archean geological record, it is unlikely that they will ever be found. It is not clear why impact events in the Archean would predominantly produce large volumes of spherules, which are not as commonly found as post-Archean impact deposits (i.e., those for which source craters are known) – although a recent report about ejecta beds from the 1.85 Ga Sudbury impact structure also included observation of spherule horizons (e.g., Addison et al., 2005). M. Huber (pers. comm., 2013) suggests that lack of bioturbation in older deposits probably better preserves spherules than younger deposits where they are more easily destroyed. Another issue could be the different compositions of the Archean, thin atmosphere, and later, denser atmosphere, which may have influenced formation of melt spherules. One difficulty in finding ejecta deposits from known impact craters is that the exact age and stratigraphic position of the impact event must be known to find a correlative distal ejecta horizon, and that is usually not simple to establish (e.g., review by Jourdan et al., 2012). The question regarding how to identify Archean impact deposits remains open and will hopefully be addressed in future studies (but see Simonson and Harnik, 2000, and Simonson, 2003, for discussions on this subject).

6.3.4. Numerical models

Johnson and Melosh (2012) have recently concluded from the spherule layer record that there was a relationship between impactor size and spherule layer size for spherules condensed from the vapor plume of an impact event. From that they calculated that impactors that had been responsible for the formation of the known Archean and Proterozoic impact spherule beds would have been in the size range from about 6 to 70 km. These numbers ought to be viewed with severe caution, as some of the thicknesses of spherule beds that contain much material mixed in with spherules due to reworking of the original – possibly much thinner – impact deposits are clearly larger than their presumed real (original) thicknesses. Some of these beds only contain a few percent of spherules, or even less, and thus could have been miniscule in volume compared to the currently observed stratigraphic thicknesses of up to a meter, or so. These authors nevertheless draw attention to the fact that "the impactor flux was significantly higher 3.5 billion years ago than it is now" (Johnson and Melosh, 2012, p. 75). They take this as evidence that this increased Archean impactor flux is consistent with a gradual decline of the flux after the Late Heavy Bombardment – an issue still ardently debated (e.g., Fernandes et al., 2013). However, because the source craters are not known, the models can only be controlled by a small number of more recent events that impacted target materials of much different composition than the Archean impact events, and for which impactors may have had different compositions and dynamic properties (i.e., angle and velocity of impact), and the atmosphere had a different composition. Although the model likely represents the best estimate that is possible based on current data, it still remains doubtful whether meaningful conclusions can be drawn from such estimates. Johnson and Melosh (2014) present a model for the formation of melt droplets, melt fragments, and accretionary impact lapilli during a hypervelocity impact, based on the results of hydrocode simulations.

6.4. Proposed but not confirmed impact structures

6.4.1. Anefis, Mali

At 18°N and 0.5°W, Rossi (2002) observed a ca. 3 km wide structure in Neogene sedimentary strata on ASTER imagery. It is located on a subhorizontal plateau and cut by drainage. The southwestern part of the rim is partially dissected. Rossi stated that "the general aspect and the noticeable erosion of the structure do not suggest a recent age...". No ground truth has been presented yet.

6.4.2. Azenak structures, Niger

Several tens of kilometers south of the town of Agadez, at about 16.5°N and 8°E, two structures of 0.5 and 1 km diameter were noted by Rossi (2002) on ASTER imagery. The distance between these structures measures about 9 km, and they are aligned in east–west direction. The bedrock is Early Cretaceous but some Quaternary playas and eolian deposits are noted in the area as well. Only the western, smaller structure has a distinct rim. Both structures may be filled with eolian deposits. No ground truth information has been presented yet.

6.4.3. Bangui, Central Africa

Girdler et al. (1992) proposed that a 600 km Magsat-based magnetic anomaly, also investigated by surface investigation and aerial magnetics, over the Bangui Basin of Central Africa (it is centered at about 04°22′N/18°33′E, Girdler et al., 1992) could represent a gigantic and very old impact structure. This, of course, is an interesting observation, as this area in Central Africa and its counterpart in Brazil are also known to host carbonado occurrences, for which several groups have advocated an extraterrestrial origin (e.g., Garai et al., 2006; see also discussion of carbonado and framesite genesis by Heaney et al., 2005, and references therein). The anomaly is somewhat elliptical, with a short axis/diameter of some 550 km. According to Girdler et al. (1992), the amplitude of the anomaly varies between −1000 nT at ground level and −20 nT at satellite altitude. This anomaly coincides with a similarly sized anomaly of low gravity. These authors used for their magnetic database a contour map constructed from Magsat data, whereby the Bangui anomaly was observed as a strong, isolated signature.

In contrast, re-processing of Magsat data by Antoine et al. (1999) indicated that this anomaly forms part of a larger feature that begins in the Bangui region but continues towards the southwest. There, it widens "like the wake of a ship" and extends into the Atlantic Ocean beyond the Mid-Atlantic Ridge. Antoine et al. concluded that the Bangui anomaly is inseparable from a larger anomalous trend that they termed the Bangui-Atlantic magnetic feature. Boundaries of this trend are, to the north, the Cameroon-St. Helena volcanic line, and, to the south, the Walvis Ridge. These authors resolved that "The genetic link between the Bangui-Atlantic feature and the Bangui anomaly remains equivocal. We provide an explanation for the Bangui-Atlantic magnetic feature – namely, that it reflects a zone of thinner ocean crust bounded to the north and south by relatively thicker crust."

6.4.4. Bangweulu Basin, Zambia

Master (1993) proposed that the Bangweulu Basin (centered at 11°10′S, 29°54′E) on the block of same name in Zambia could represent a 150 km multi-ring impact basin. He indicated the basin covered a ca. 150 km wide area that is occupied by Lakes Bangweulu and Kampolondo, and the Bangweulu Swamps. The regional basement comprises 1.8 Ga granitoids and ca. 1.1 Ga Katangan cover rocks. To the north cover rocks of the Mporokoso Group of 1.8–1.3 Ga age form the arcuate Luongo Fold Belt that partly defines the perimeter of the outermost ring at 125 km radial distance from the center of the alleged Bangweulu structure. Master noted that drainage into the basin is centripetal. At 100 km radial distance, he noted an arcuate watershed to the west of the structure. Also arcuate shapes of islands in Lake Bangweulu caught his eye; they are alleged to follow the arcuate northwest boundary of the structure.

Master emphasized that in contrast to the Great Lakes region, the Bangweulu Basin area was seismically inactive and, thus, would be unrelated to rifting. He referred also a positive aeromagnetic intensity anomaly over the central Bangweulu depression, as well as a positive magnetic anomaly over the central part of the basin, surrounded by a concentric low. And a roughly circular gravity anomaly outlined by a −140 mgal contour could be found in the regional Bouguer gravity field, centered on Lake Bangweulu. On satellite imagery of Central Africa Master claimed to have noted a roughly circular outline of the basin. These geomorphological and geophysical observations were used by the author to arrive at the conclusion that the Bangweulu Basin represented an eroded remnant of a multi-ring impact structure that postdated the Katangan Supergroup. He indicated that a ground search for impact evidence was planned for 1994 – but apparently this has still not taken place. The Bangweulu basin remains an unsubstantiated proposition of an impact structure.

6.4.5. Bateke Plateau structure, Gabon

Master et al. (2013) referred to a possible impact structure centered at 14°27′29″E/0°38′45″S in eastern Gabon. The ca. 7.1 km diameter structure was located about 5 km west of the border with the Republic of Congo. The structure was identified on Landsat imagery of eastern Gabon, in a region dominated by Paleogene to Neogene sedimentary rocks of the Bateke Plateau that unconformably overlie Archean basement. Master et al. (2013) discussed the results of digital elevation modeling of SRTM data that indicated that the structure comprised two nested toroidal rings with an intermediate ring-shaped depression. The outer toroid had a diameter of 5.8 km and a width of 1.3 km, which combines to a diameter of 7.1 km. The inner ring feature had a diameter of 1.4 km and a width of 700 m. Master et al. considered this morpohology to be consistent with that of an impact-generated complex, central peak-ring structure. They excluded some other genetic possibilities, such as young igneous intrusions, diapiric structures, and karst structures. These authors, thus, suggest that the Bateke Plateau Structure could represent a post-Neogene complex impact structure and that the degraded morphology is suggestive of a Pleistocene rather than a Holocene formation.

6.4.6. Chituli structure, Luangwa Valley, Zambia

The presence of a 3.8 km wide, generally circular structure centered at 32°24′E/11°15.33′S on the western flanks of the Northern Luangwa valley of Eastern Province of Zambia was reported by Master (2001). He named this structure after the Chituli River that flows through this area. The structure was said to be somewhat circular but polygonal. On aerial photographs and Landsat imagery, Master discovered a prominent outer ring of rocks with a positive relief. Apparently, it was Sykes (1978, also 1994), who first drew attention to this structure that was investigated as a possible carbonatite or kimberlite. He found that the area was overlain by a ferruginous sandstone of likely Karoo age. However, the structure does not seem to display any geochemical or geophysical anomalies as known from other alkaline intrusives in Zambia.

According to Master, the structure is surrounded to the south by porphyroblastic Chituli-Lufila gneisses of the Muva Supergroup of the 1.3–1 Ga Irumide Belt. To the east, sandstones of the upper Luwumbu Formation of the lower Karoo Supergroup occur. He thinks that sandstones occurring in the interior of the structure could also belong to this formation. The northern half of the structure was surrounded by sandstones and carbonaceous mudstones of the lower Luwumbu formation. Master (2001) proposed the Chituli structure as a possibly Neoproterozoic or Late Paleozoic impact structure. His argument for that is based on the observation that the alleged structure is located on deformed rocks of the Irumide Belt but infilled with undeformed Karoo rocks – in an area that lacks magmatic activity and as the morphology seemed to resemble that of eroded complex impact structures. No tangible evidence has been obtained on the ground since.

6.4.7. El Mrayer, Mauritania

At 22.5°N and 7.2°W, Rossi (2002) recognized a ca. 2 km wide structure in Cambro-Ordovician basement, in an area that has several prominent longitudinal dunes trending in NE–SW direction. Dunes also cover the structure in some sectors. The morphology of this crater-like structure is complex: a subcircular inner depression is bounded by a concentric, deformed zone. The author envisaged a small, possibly secondary crater feature to the east, against a longitudinal dune, which he interprets to suggest a young age for the main structure, as the small secondary had not been eroded yet. No ground-truth has been presented yet.

6.4.8. Faya Basin, Chad

Faya Basin is a ca. 2 km wide, almost circular structure located at 18°11′N/19°33′E) in northern Chad (Fig. 49). It was discovered by Schmieder and Buchner (2007) through a combined Landsat and Shuttle Radar Topography Mission data investigation. It is located some 55 km east-northeast of Faya (Largeau) city, close to Odofu oasis in the district of Borkou-Ennedi-Tibesti. Faya basin was formed in sandstones of likely Late Devonian age that form the Borkou plateau. This information provides the only currently available age limit. The remote sensing observations suggest that the morphology of the structure resembles that of small, complex impact structures (e.g., BP, see above). The feature comprises an elevated rim surrounded by concentric faults, an annular depression, and a small central elevation. Like Aorounga, the Faya Basin

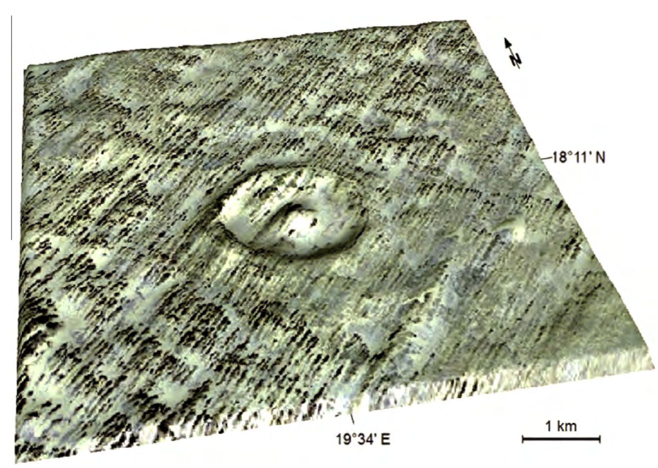


Fig. 49. Faya Basin structure, Chad. The structure measures ca. 2 km in diameter. 3D-representation based on superposition of Landsat 3-2-1 RGB plus panchromatic Band 8 data (resolution 15 m) and SRTM data (resolution 90 m). 5× vertical exaggeration. Courtesy of Martin Schmieder (Curtin University, Perth) and Elmar Buchner (Universität Neu-Ulm).

structure is transected by conspicuous, NE–SW trending yardangs. These and small-scale longitudinal dunes are abruptly cut off at the basin's margin. The authors think that large parts of the basin are covered by eolian deposits and salt. They mention that the remote sensing data suggest the inner slope of the rim is much steeper than the outer one. The rim seems to be elevated above the surroundings by a mere 10 m. The central topographic high is somewhat more pronounced and stands up to 35 m above the deepest part of the basin's floor, and 10–20 m above the outer terrain. Profiles of SRTM topographic data provide cross sections of the basin that are rather similar to those of complex impact craters.

Schmieder and Buchner (2007) did not find any indication of volcanic activity in the area (P.M. Vincent, pers. comm. to these authors, 2006). No evidence of salt diapirs, sinkhole structures, or glacial erosion features could provide alternate explanations for the genesis of the Faya Basin structure. Unfortunately, the prolonged civil war in northern Chad, with large tracts of land polluted with landmines, has made geological groundwork in this region a very hazardous undertaking.

More recently, Schmieder and Buchner (2010) added more detailed observations on Faya Basin from Spot 5 imagery, specifically concerning the central topographic high. It covers an area of ca. 250×150 m and represents a slightly triangular complex. They noted an apparent SW–NE trending section with ridges on each side of the bisecting axis, and interpreted the detailed imagery as suggestive of steep dips of the sandstone strata and report comparisons with central uplift structures of some Martian craters.

6.4.9. Foum Teguentour, Algeria

Lambert et al. (1981) investigated the 8 km diameter Foum Teguentour (26°14.5′N/02°25′E) and the 0.8 km diameter *Mazoula* (28°24′N/07°49′E, see below, section 6.5.12) structures in Algeria and found that there was no evidence for an impact origin. Foum Teguentour forms a large bull's eye ring pattern. The authors concluded that "Although the high circularity and morphology are consistent with an impact origin, the prominence of ductile deformation, the nature of the formations (clay-gypsum with sandstone intercalations), the type of folds, the relationships between the structure and a surrounding plateau, and the lack of any evidence of shock effects better support a diapiric origin".

We do feel, however, that further analysis of this structure is required. Particularly, detailed ground geophysical analysis may provide further hints at the origin of this structure. Thus, Foum Teguentour is retained in section 6.4.

6.4.10. Gogui, Mauretania

Based on a survey of ASTER satellite imagery, Rossi (2002) identified a 500–600 m diameter structure at 15.5°N and 11.4°W. He stated that the feature shows a relatively "pristine morphology". Bedrock in the area is said to be Paleozoic metamorphic rock. The structure is characterized by a circular rim and a flat bottom. On the outside of the structure, presence of ejecta is hypothesized. The age of this structure is judged as rather young because of the "pristine aspect". No ground truth has been presented yet.

6.4.11. Highbury, Zimbabwe

Highbury is a near-circular area of about 15 km width, centered at 17°05′S/30°09′E. This area was first noted by S. Master (University of the Witwatersrand) in 1985 on Landsat imagery. In 1993, a South African–Austrian–German expedition to Zimbabwe, which included both authors of the present review, took place to investigate several sites of possible impact structures, including the Highbury area (Master et al., 1994). The regionally occurring rocks at Highbury are arkoses and meta-dolomites of the Deweras Group, with adjacent occurrences of the Striped Slates and Mountain Sandstone members of the Nyagari Formation of the Lomagundi Group. The circular structure is visible on Landsat imagery where some contrast between the agriculturally used main area of the structure with the hills on the eastern and western sides, which are formed by Lomagundi strata, is evident. In the north and south the Striped Slates and Mountain Sandstones are abruptly terminated. In this image the Highbury area is indicated as perhaps being as wide as 15 km, but perhaps up to 25 km. These authors did refer to locally metamorphosed carbonates, carrying tremolite and wollastonite, and an early (Jacobsen, 1962) discovery of a small granophyre occurrence in the Munwa river-bed, intrusion of which had been mooted as the possible cause of skarn development. The outline of the alleged Highbury structure is somewhat pear-shaped and narrowing to the northwest. Master et al. (1994) noted a small elevated area in the central part of the structure, elevated by some 80 m above the surroundings.

During a half-day visit to the area in 1993, 39 samples, mostly arkoses and sandstones, were collected in both the central and exterior parts. In thin sections, quartz crystals show ample subplanar fluid inclusion trails, initially thought to resemble Vredefort fluid inclusion trails (Master et al., 1994). Master et al. (1994) even reported bona fide PDFs and some pockets of fresh glassy material. However, this has never been confirmed and the thin sections available to us do not indicate such features. Instead, fluid inclusion trails clearly lacking in planarity prevail and are not comparable to true impact deformation.

If it were confirmed through further geological work that the Highbury area represented an impact structure, the upper age for this event would be constrained by some north–south-trending faults of Magondi (1.8 Ga) age. Master et al. (1994) discussed that the structure was offset in the southwestern sector by dextral wrench faulting of possible late Irumide (1 Ga) age, which would then be a limit for the lower possible age. A granophyric granitoid, which was sampled in the southern part of the structure, whereby the limited extent of outcrop did not allow to determine whether this was a large boulder or actual outcrop – in the hope that it might be impact melt rock (due to its superficial resemblance to Vredefort Granophyre – see above) – yielded a U–Pb zircon age of ca. 1.4 Ga (S. Master and R.A. Armstrong, pers. comm., 1996).

Without further detailed geological analysis of this area, it remains entirely inconclusive whether the Highbury area could be the eroded remnant of an old impact structure.

6.4.12. Ibn-Batutah, Libya

Remote sensing investigations by Ghoneim (2009), based on Landsat ETM+, dual-band (L and C) and dual-polarization (HH and HV) radar (SIR-C), and SRTM data, suggested the possible presence of a 2.5 km diameter and ca. 25 m deep circular basin feature at 21°34′10 N″/20°50′15″E on the Hamadat Ibn-Batutah plateau of southeastern Libya. This singular geological structure is formed in Nubian sandstone of likely early Cretaceous age. Most of the feature is surrounded by a circular rim of just a few meters elevation over the surrounding terrain. Ghoneim's imagery illustrates that the structure strongly disrupts the regional paleo-drainage (her Fig. 3).

Topographic profiles constructed from SRTM data suggest a simple bowl-shape for the structure. This could be the result of complete infill of a rather young impact structure of this size, with a possible central uplift feature like that observed at the BP impact structure (of similar diameter) completely obscured by the fill with eolian deposits. Ghoneim (2009) discusses that alternate modes of origin such as magmatism, diapirism, karst dissolution, and glacial or fluvial erosion are not supported by the evidence at hand, so that the impact hypothesis remains the most reasonable. She declared that "verification of this hypothesis will require collection and analyses of rock samples from in and around the structure, although the area is currently almost inaccessible for environmental and security reasons." This includes the harsh and inaccessible landscape of the area of interest, the security situation along the Chad-Libya border, and the currently uncertain security situation in much of Libya in the wake of the dramatic political developments of 2011 that lead to the institution of a new political dispensation.

6.4.13. In Ezzane, Algeria

Bonin et al. (2011) drew on meteorite impact during the Quaternary as the cause for the formation of circular structures (in the area around 23°29′N/11°14′30″E) within Ordovician Tassili sandstone of the Tuareg Shield of the Eastern Hoggar region of southern Algeria. Five such structures occur in the In Ezzane volcanic province, a 500 km^2 plateau along the Algerian-Niger border. These nestled circular structures have diameters of 4–9 km and are aligned along a north–south trend. They truncate each other on the southern sides. Bonin et al. mention a reconnaissance field trip that established that there was a total absence of the products of volcanism that one could relate to these structures. Less than 500 m outside of the rims, the Tassili sandstone was reddened and dissected by networks of fractures filled with fine-grained, brown material. The rims of these 5 structures represent topographic highs and are composed of brecciated sandstone with dark veining (mainly goethite) in the form of networks that cluster in zones parallel to the rim. At the edges of dark veins, sandstone is shattered and angular fragments of it are included in veins. The authors infer that the veining and mineralization therein involved large volumes of fluid and they refer to hydrofracturing. The age of these structures is estimated at Quaternary due to their occurrence in 2 Ma old basanite and the overall fresh appearance of these crater-like features. Bonin et al. (2011) conclude that all currently available evidence was suggestive of an impact origin of these structures.

Obviously, in the light of what constitutes accepted evidence for impact (see above; French and Koeberl, 2010), this conclusion cannot be supported. However, these structures are undoubtedly intriguing and deserve further research.

6.4.14. Jaraminah, Libya

Dunford and Koeberl (2009) reported on two possible impact structures noted first on aerial and satellite imagery in western Libya – *Tmisan* (see below) and Jaraminah. Dunford (2008) also visited the structures and obtained samples.

Jaraminah is located at 26.540°N and 10.588°E. This location is about 15 km south of the Thamad al Jaraminah well in the municipality of Awbari, and about 70 km from the border with Algeria. The structure has a diameter of about 2.2 km and comprises a "remarkable" set of ring features that are interpreted as "uplifted and erosionally truncated strata" with outward directed dips. An apparently uplifted terrain occupies the central area. It has an irregular outline and a diameter between 300 and 400 m. The innermost part of this terrain shows a conical morphology. An intermediate ring occurs at about 600 m from the center and is conspicuously circular. A series of outer rings that are closely spaced forms the outer anticlinorium. They seem to represent a series of indurated sedimentary layers. The northwest–southeast trending fabric of the environs is deflected at the structure (see also Dunford, 2008).

Twelve samples from Jaraminah were studied petrographically. These siliciclastic rocks did not show any diagnostic shock deformation but are said to contain tectonic deformation lamellae and possibly planar fractures. Therefore, the origin of this structure remains uncertain and the structure requires further field analysis and laboratory analysis.

6.4.15. Jebel Hadid, Libya

A 4.7 km wide structure, comprising a series of concentric ring forms (Fig. 50), was detected by Schmieder et al. (2009) centered at 20°52′12.43"/22°42′17.73E in the southern Al Kufrah basin of Libya. The structure seemingly occurs entirely in Nubian sandstone and would therefore be of post-Cretaceous age (see also *BP* and *Oasis* structures, above). The structure was recognized by remote sensing, using Landsat-7 Thematic Mapper Plus (ETM+) satellite images and Shuttle Radar Topography Mission (SRTM) terrain elevation data. A pre-Pliocene age limit is also proposed by the authors based on the presence of a Late Miocene to Pliocene drainage system that overprints the structure.

Schmieder et al. (2009) proceeded to discuss alternative modes of origin, besides impact cratering, and concluded that magmatism, diapirism, sand volcanism, karst dissolution, or glacial erosion did not provide adequate options and, consequently, arrived at the conclusion that this structure likely represents a significantly

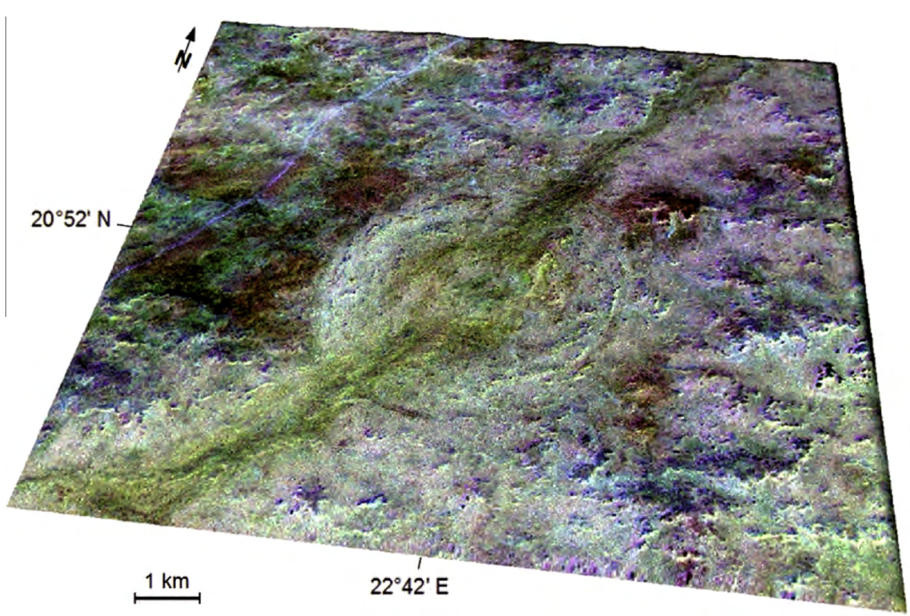


Fig. 50. Jebel Hadid structure, Libya, 4.8 km diameter. False-color (bands 7-5-4 RGB pan-sharpened) Landsat data superposed onto SRTM data, 10× vertical exaggeration. Courtesy of M. Schmieder (Curtin University, Perth) and E. Buchner (Universität Neu-Ulm).

eroded, complex impact structure. They noted a strong similarity on remote sensing imagery between Jebel Hadid and Tin Bider in Algeria. And they observed the possibility that a structure like this could well have potential for hydrocarbon exploration, as the Al Kufrah basin has a high potential for hydrocarbon deposits. Bar any field visit and detailed ground proofing that might provide confirming evidence for impact, this genetic proposition remains unconfirmed.

6.4.16. Jwaneng South structure, Botswana

Master et al. (2009) and Master (2010a,b) reported a pronounced, 1.3 km wide, circular geophysical anomaly at about 24°46′E and 24°42′S, some 15 km south of Jwaneng (a prominent diamond-mining town) in Botswana. They termed this structure the Jwaneng South structure. It was discovered by airship-mounted gravity gradiometer survey in the course of kimberlite exploration. Besides a prominent Bouguer gravity in these aerial data, the structure is said to be prominent in airborne EM, ground gravity, and CSAMT imagery. A number of diamond drill holes in this structure revealed a sedimentary rock sequence comprising evaporitic lacustrine carbonates with plant fossils that are overlain by sediment breccia and sandstones of the Kalahari Group. The maximum thickness of sediment is given as ca. 300 m. These studies showed this feature to have a circular, bowl-shaped geometry. In addition, below the sedimentary rocks a <60 m thick unit of allogenic breccia occurs that features a series of breccias with granitic fragments, grading downward into brecciated mafic rock (according to Master et al. resembling Karoo dolerite in terms of chemistry) and eventually unbrecciated granite of the 2785 Ma Gaborone Granite Complex. Kimberlites or other intrusive rocks had not been encountered. Master et al. stated that "the rocks had suffered intense shock deformation (although impact-diagnostic PDFs have not been found)". In the light of this, we need to point out that the reference to "shock deformation" is premature in the light of complete lack of proper evidence. The deformation style indicated in this paper is of the brittle kind, with angular rock fragments. However, Master et al. proceed to refer to "features found in known impact structures" – including mosaicism, deformation bands, and lamellae in plagioclase and alkali feldspars, and cleavage in quartz, as well as "gries"-textured breccias and multiply-striated joint surfaces. Master (2010a,b) adds to these findings an observation of breccia dikes consisting of polymict clasts including subrounded fragments of lava of different sizes, and individual plagioclase crystals, all of which are enclosed in a fine-grained altered dark matrix (resembling pseudotachylite), rich in clay minerals and red iron oxides. These dikes are alleged to cross-cut intensely brecciated mafic lava.

In conclusion, none of these reported features relates to unambiguous evidence of impact deformation. In fact, the impact hypothesis advanced for this structure remains unsupported by any concrete proof for impact (shock) deformation. The fact that mafic lavas and related brecciation occur in this structure could be considered a hint at possible internally triggered deformation. Jwaneng South is a very interesting, and indeed intriguing, structure that deserves further investigation.

6.4.17. Karas, Namibia

Corner (2008) suggested the existence of a somewhat circular structure of 300 km diameter, located in the area around 26°S/19°E at a deep crustal level to the northeast of Keetmanshoop in southern Namibia. The subsurface structure had a central magnetic high and outer ring structures out to a radial distance of 300 km. The existence of this structure was inferred based on interpretation of magnetic and gravity anomalies (see also Miller, 2008a,b). For structural reasons Corner (2008) gave an age of pre-1200–1350 Ma for this alleged impact event. He further suggested that this event had "produced a zone of long-lived crustal weakness that has influenced the region up to the present", and he points out that "several of the present rivers have taken advantage of parts of ring-shaped features..." (Miller, 2008a, p. 26-1). No bona fide evidence for the presence of such an impact structure is known.

6.4.18. Kogo, Equatorial Guinea

A 4.67 km diameter structure was discovered at 1°11′N/10°1′E by Martinez-Torres (1995, also referred by Master, 1998) in SIR radar imagery, in the tropical rain forest of Equatorial Guinea, in the district of Kogo. The structure is located on Precambrian gneisses and seems to be faulted away on its western side by Neocamian faults related to the opening of the Atlantic. This would make the structure, according to Martinez-Torres, older than about 145 Ma. Master (1998) discussed the carbonado fields of Central Africa (see also *Bangui*, above) and Brazil and noted that the Kogo structure is situated exactly between these two carbonado fields. He referred to a systematic decrease in abundance and maximum size of carbonados with distance away from the Kogo structure (after Bardet, 1974), and concluded that Kogo "is proposed as the source crater for impactogenic (or meteoritic) carbonados of Bahia and the Central African Republic".

6.4.19. Lac Iro, Chad

Garvin (1986) identified Lac Iro at 10°10′N/19°40′E in southeastern Chad as a possible impact structure on SIR-C radar and Landsat data. This sub-circular lake (according to Garvin, possibly an ephemeral lake) of up to 13 km width is located just north of Bahr Salamat, and 130 km northeast of Fort Sarchambault, in Quaternary alluvium of the Chari Embayment (Martin, 1978). Small granitic intrusions and outcrops of Paleozoic basement occurred close to the lake (Nickles, 1952). Garvin noted that the lake seems to deflect the Bahr Salamat River. The circularity of much of the lake's circumference is conspicuous, but in the southwest the shoreline is more irregular. Garvin interpreted this as a possible effect of regional runoff. The local geology is not known. It is interesting to note that the lake is comparable in size to the El'gygytgyn impact structure in northeast Siberia, a large part of which is covered by a lake as well. Ground-based work, and scrutiny of the regional information perhaps collected in the course of oil exploration, is indicated to further investigate Lac Iro as a prospective impact structure.

6.4.20. Minkébé and Mékambo structures, Gabon

Recompilation of the national airborne magnetic data of Gabon by Antoine et al. (2000) revealed two large – 90 and 50 km diameter, respectively – circular anomalies. The structures are located not far from each other, with about 250 km between their respective centers. They occur in a remote, sparsely populated region of equatorial rain forest, in the areas of the Minkébé and Mékambo 1:200,000 topographic map sheets of Gabon – which has leant them their names.

Minkébé is the larger of the two structures and is centered on 1°21′15″N/12°24′29″E. The structure straddles the border between the Woleu-Ntem and the Ogooué-Ivindo provinces. It is characterized by a circular rim, which is not transgressed by regional structural fabrics. Within the structure, fabrics trend NW–SE, but outside, to the west and north of the structure, they are oriented NE–SW and ENE–WSW. The eastern half of the structure displays a subdued magnetic relief that is part of a concentric belt continuing for some 50 km beyond the rim. The structure comprises a central plateau of 660–940 m.a.s.l., which is surrounded by the concentric valleys of the Mvoung and Nouna rivers. In the northern part of the structure is a watershed that separates it from the north-flowing Ntem river. Country rocks around the structure are

Archean granites, gneisses and greenstones of the North Gabon Block, of 2.76–3.1 Ga ages.

The smaller *Mékambo* structure is centered on 0°55′39″N/13°40′25″E. It is located in the Ogooué-Ivindo province, but its northern part extends into Sangha province of the Republic of Congo. The structure is recognized as a 50 km wide circular region of subdued magnetic relief that also shows an intense central magnetic anomaly. The structure is surrounded by a ring of high-relief magnetic anomalies extending beyond 30 km from the structure. This belt is, according to Antoine et al., caused by ferruginous greenstone belts. There is a central plateau of >500 m elevation in the interior of the structure, which is surrounded by a concentric depression, in which the west-flowing Zadié and Liboumba rivers have their beds. Like the Minkébé structure, the Mékambo structure is also located mainly in Archean granite-greenstone terrain of the North Gabon Block. The northern part of the structure is apparently covered by flat-lying sedimentary rocks of the Sembé-Ouesso Group of 2.1 Ga age, which are intruded by E–W trending magnetic dikes.

Antoine et al. (2000) drew attention to the fact that these large aeromagnetic anomalies are quite conspicuous in the aeromagnetic pattern of the North Gabon Block. They proposed that the two structures could be the result of impact, perhaps even related to a twin impact event such as those that formed the East and West Clearwater structures in Canada. If this were confirmed, the structures would be the result(s) of very old impact event(s): they would post-date the >2.8 Ga Archean basement and be older than the Paleoproterozoic Sembé-Ouesso Group. Antoine et al. (2000) also discussed the carbonado conundrum, which is referred in the present publication in connection with the *Bangui* and *Kogo* impact proposals. Antoine et al. (2000) considered the two Gabonese structures as possible sites of origin for the Brazilian and Central African carbonados, and favored them over the poorly constrained Bangui "structure" and the, in their opinion, much too small Kogo option.

Obviously due to the remote locations of the two Gabonese structures, no reports of efforts to verify the impact hypothesis for the Minkébé and Mékambo structures of Gabon have been published.

6.4.21. Mora Ring, Cameroon

Using SIR-C radar and Landsat data, Garvin (1986) identified several prospects of possible impact structures in Central Africa. This includes a feature that he termed "Mora Ring" and that is located in northern Cameroon, just east of the Mandara Mountains and west of the Maroua Tertiary volcanics, at 11°N/14°E. He states that the structure lies equidistant between the city of Mora at the edge of an alluvium-covered region at Lake Chad against the Cameroon basement rocks (granites and mica-schists), and the town of Gétalé in the lower savannas abutting granitic uplands (according to the map of French West Africa by Nickles, 1952). The Mora Ring structure had a topographic expression comprising two ring forms around a central complex of hills. Garvin observed that faults in adjacent basement were apparently truncated at the outer ring of the structure. The outer ring diameter is at least 7 km wide and was described as having tens of meters relief and a polygonal outline. The central hill complex was quite prominent, and according to Garvin rises by more than 100 m above the moat at its base. Garvin emphasized that the general morphology of the structure could be consistent with an occurrence of a ring dike or eroded volcanic cone sheet, but he also observed that this structure at Mora was unique in this region of Cameroon. He concluded that it could represent an impact structure but also an eroded volcanic complex, a differentiated weathered pluton, or a large, isolated ring dike like those known from central Nigeria. Age information on the structure remains speculative, but it must be younger than nearby Paleozoic granites. If the structure were of volcanic origin, it would likely be of Tertiary age (Nickles, 1952).

6.4.22. Mousso, Chad

Mousso structure in northern Chad (Fig. 51) was proposed as a possible impact structure by Buchner and Schmieder (2007). This ca. 3.8 km diameter structure is centered on 17°58′N and 19°53′E, about 80 km east of Faya (Largeau) town near Mousso oasis in the Borkou-Ennedi-Tibesti district. The structure was recognized on Landsat 7 ETM+ and SRTM imagery. It has a circular rim with concentric faults, an annular basin, and a central elevated area. The inner basin and the central elevated area have diameters of 3.2 and 1.8 km, respectively. These characteristics are similar to those of complex impact craters, and Buchner and Schmieder (2007) find a particular resemblance of Mousso to the confirmed Spider impact structure in Australia (Abels, 2005).

The area around Mousso displays numerous NE–SW trending yardangs, barchan dunes, and salt diapirs. The rim of the Mousso structure is in parts superimposed onto the edge of a sedimentary plateau. It is observed as a semicircular escarpment in the eastern, and as a dark annular trace in the western domain. In the eastern part, concentric semicircular faults along the rim suggest partial, inward-directed slumping off the rim. The central peak area seems to comprise radially oriented arcuate ridges. The central peak is elevated by 30–40 m above the ring basin and by ca. 10 m above the rim crest. The area of the Mousso structure consists of flat-lying late Precambrian and early Paleozoic sedimentary rocks. The authors refer to maps of the area (e.g., Wacrenier et al., 1958) and suggest that the structure could, in its northern part, be located in Cambro-Ordovician, and perhaps in subsidiary Silurian strata. Following Furon (1963), this could be an up to 500 m thick sequence of quartz-rich sandstones. This package provides the only age constraint for a presumed impact event, at <542 Ma (after Gradstein et al., 2005).

Schmieder and Buchner (2007) discussed in some detail the possible genetic options for this structure and concluded that "no endogenic geological processes such as magmatism, diapirism, karst dissolution, and glacial or fluvial erosion can conclusively explain the formation of the Mousso structure within a large area of flat-lying early Paleozoic sandstones". While field-based confirmation of this opinion is obviously required, the unstable civil situation in northern Chad and the fallout from prolonged civil war have so far not allowed this to be undertaken.

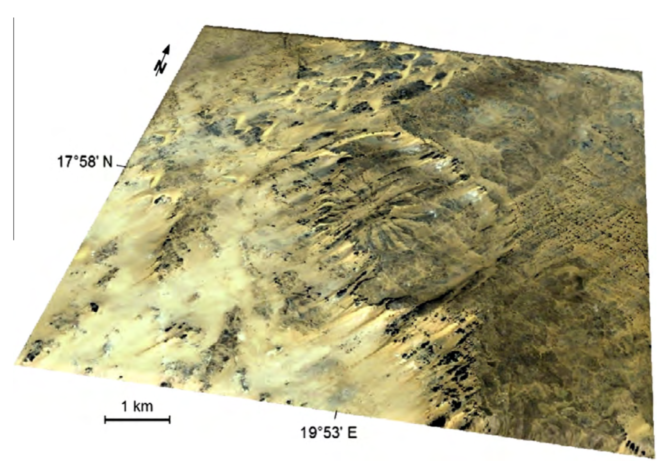


Fig. 51. Mousso crater structure, Chad, 3.8 km diameter. Combination of SPOT satellite data and SRTM topographic data, with a 3-fold vertical exaggeration. Courtesy of M. Schmieder (Curtin University) and E. Buchner (Universität Neu-Ulm).

6.4.23. Ntwetwe, Botswana

In the large (600,370 km^2) country of Botswana, to date only one impact structure (*Kgagodi*) has been confirmed (see above). Besides Ntwetwe, three other structures have been proposed to be of impact origin (*Jwaneng South; Okavango Delta feature; Unnamed, Botswana*).

A ring structure of 7 km diameter was reported by Master (1994) to occur at 24°50′E/20°55′S at about 32 km east-northeast of Rakops on the Batete River, and about 17 km north of Xhumo on the Batete River northeast of Lake Xau, near the western edge of *Ntwetwe* Pan. From this pan, which is the westernmost one of the Makgadikgadi saltpans in the Kalahari Desert of northeastern Botswana, he derived the name for this feature. The structure was detected on a Landsat satellite image published by Short et al. (1976). It is described as comprising a central dark area which might have a positive relief and that is surrounded by a light-colored ring and a further, outermost, ring which could also have a positive relief. Master (1994) cautiously discussed the possibilities that this feature could be (a) an optical illusion of no geological significance, the result of a chance arrangement of lithologies and/or colors with a roughly circular or ring-like geometry (compare our interpretation of the *Bir Anzarane* structure of Morocco, section 6.5.4, below), (b) the eroded remnant of a nearly circular oxbow lake within meandering-stream deposits, (c) the surface expression of an igneous intrusion (ring complex or pipe-like intrusion such as a kimberlite, or (d) a complex impact structure with a central uplift.

Option one was rejected because of the regularity of the features observed. Option 2, the oxbow lake, was judged the strongest by Master, because of the general appearance of such deposits in the wider area. Considering the significance of kimberlitic intrusions in the Kalahari region, option 3, an igneous intrusion, was also not rejected. However, on the basis of the possible central elevation, resembling central uplifts of complex impact structures, the kimberlite option was not preferred, in contrast to the impact structure possibility. Master (1994) proposed that verification would be readily possible by checking whether the central area exposes sedimentary rocks of the Karoo Supergroup that might have been uplifted from their normal regional position below some 100–200 m of Kalahari Group sand cover. Since then, no further scientific evidence has been reported for this proposed structure.

6.4.24. Okavango Delta feature, Botswana

According to Henshaw (1997), there was oral history (allegedly two incompatible stories) relating a lake in Botswana with a fireball (S. Master, pers. comm.). This 22 m diameter, 3–4 m deep lake, also referred as Khurunxaraga crater, occurs at 23°20′E/19°50′S, about 35 km north of the town of Maun. It is formed in a calcrete formation. No further information about this site is known to us.

6.4.25. Omeonga, Democratic Republic of Congo

The Omeonga feature, or as it has been known for a number of years as well, *Katako-Kombe* (name proposed by A. Walemba, Council for Geoscience, Pretoria), is located in the central region of the DRC. The structure is centered at about 3°35′11″S and 24°29′10″E. It has been named Omeonga after the village closest to the structure. Only a few years ago Omeonga was first listed on a website of possible or probable impact structures (http://www.impacts.rajmon.cz). And it has only been these last two years that some more detailed information has become available. Monegato et al. (2011) compiled remote sensing information and matched it with regional geological knowledge. They reported an annular drainage pattern encircling a 45 km wide area that revealed an about 20 km wide central, rather smooth area. This central zone was slightly elevated above a surrounding ring depression corresponding to the bed of the Unia River (termed Lonya River by Ferrière et al., 2011). These authors concluded that the morphology was similar to that of several confirmed impact structures and they attempted to discard several other hypotheses for the origin of this structure (magmatic activity, salt diapirism, karst dissolution) on the basis of published regional information – leaving impact as the most likely process to have generated this structure.

Only one group of researchers, led by Ludovic Ferrière (Natural History Museum, Vienna) has, to date, endeavored to visit the remote Omeonga area in order to investigate the impact suggestion. Ferrière (2011) and Ferrière et al. (2012) report difficult access and seriously limited local outcrop conditions. While much of the structure could be surveyed on the ground, only rare outcrops could be sampled and contacts between lithologies have remained obscure. Samples are reported to comprise various sedimentary rocks, mainly sandstones, of likely Pleistocene–Pliocene age. Neither shatter cones nor breccia occurrences were detected. The authors exclude the possibility that volcanism could have anything to do with the formation of this structure, as volcanics were not observed. Ferrière (2011) and Ferrière et al. (2012) presented results of preliminary petrographic analysis of their 32 samples. They concluded from this that "conclusive microscopic shock-metamorphic features" have not yet been detected but that the observation of conspicuous planar fractures in quartz and a few quartz grains with possibly 1 set of planar deformation features appeared promising". Clearly, further analysis is required. Ferrière (2011) suggested a slightly smaller diameter, in comparison to Monegato et al. (2011), of 38 km.

6.4.26. Oun, Chad

Two interesting structures were identified by González and Alonso (2006) in northern Chad (see also *Uri* below). They used a combination of GoogleEarth and NASA World Wind imagery that are freely available on the internet. For the study region of Chad, this method provided a resolution of 15 m per pixel.

Both structures are located close to the border between Chad and Libya. The structures were recognized because of their circularity and the fact that they clearly stand out from the regional geological environment. The respective centers of the structures are 50 km apart, and they are located some 250 km north of the confirmed Aorounga impact structure. The authors draw attention to the fact that the three structures discussed are all aligned.

Oun is the more northerly located structure of the pair; it is located at 21°44′N and 19°20′E. It measures 8 km across and comprises three annuli with intermittent ring depressions. The authors think that the irregularities of the outermost ring could be the result of partial cover with eolian deposits. The ring is discontinuous in the north due to the presence of what they called a "fracture". Inner rings seem to be more continuous. Dips of the strata seem to be radially outward. The innermost ring encloses a central depression that is crossed by a prominent drainage system. There are alleged landslides on the inner ring wall, and the authors say that "some volcanic build-ups have developed from fractures that affected the outermost ring – presumably after the formation of the structure." While it appears suspicious that volcanic deposits should occur in this structure, we do not discard a possible impact origin off-hand – ground-truthing is required but may be difficult in this border region, which is said to hold numerous landmines dating from the war with Libya in the late 1970s. And the present civil war in northern Chad should not be forgotten either...

6.4.27. Ouro Ndia, Mali

Rossi (2002) proposed – based on ASTER-based remote sensing reconnaissance – an impact origin for an about 3 km diameter structure recognized at 15°N and 4.5°W in Mali. The feature is partially filled with a subcircular lake. It is located in the Niger delta, on Quaternary deposits, and was recognized as an isolated landform that is dissimilar to other small lakes in the region with regard to both size and shape. The author identified a drainage system on the rim and possible landslide deposits in the northeastern part of the structure. Morphology and geomorphological and geological setting suggested to Rossi a relatively recent origin. A lobate pattern outside of the rim was tentatively interpreted as possible fluidized ejecta. No ground-truth has been presented yet.

6.4.28. Ras Zeidun, Egypt

Barakat (2011) suggested that a circular structure of 7 km diameter occurred within the basement complex of Egypt's Eastern Desert, just south of the Wadi Zeidun gold field, and that this structure could be a possible impact structure. No location values for this feature were presented. Barakat referred to regionally unusual jasperite that contained primary quartz grains with "features characteristic of impact" (ibid). This interesting notation leaves high expectations for further findings.

6.4.29. "Reitz" structure, South Africa

Repeatedly the possible presence of a large – variably 150–500 km wide – ring structure centered about 25 km southwest of the hamlet of Reitz in the Free Sate Province of South Africa has been suggested (Hargraves and Fuller, 1981; Antoine and Andreoli, 1995). This location would be centered or near 26°30′S/27°59′E. Hargraves and Fuller constrained this alleged structure by the arcuate northwest rim of the Central Rand Group basin and the Pongola basin to the southeast, and considered a centrally located, high-amplitude, positive Bouguer gravity anomaly of regional extent. Antoine and Andreoli discussed a model of the magnetic anomaly over this region and referred to a circular structure of more than 150 km width. There is no confirming evidence for this region to represent a large impact structure.

6.4.30. Rwanda crater structures

Denaeyer and Gérards (1973) proposed the possibility that some crater-like structures in Rwanda might be of impact origin but did not provide any supporting evidence.

6.4.31. Saghira, Libya

El-Baz and Ghoneim (2007) referred to a "cone-shaped crater" of 5.5 km^2 area extent alleged to occur just north of Al Kufrah oasis. As this structure – at 2.7 km diameter – is small, they called it "Saghira" meaning *small* in Arabic. No further detail was provided, not even the exact locality.

6.4.32. Setlagole, South Africa

At the 2008 Large Meteorite Impacts and Planetary Evolution IV Conference at Vredefort, South Africa, Anhaeusser et al. (2008) presented a first report about a possible, newly discovered, large and old impact structure, around 25°7.5′E and 26°22.5′S, on the western Kaapvaal Craton of South Africa. More recently, Anhaeusser et al. (2010) reviewed the available evidence in more detail. The discovery is based on the recognition of a 25–30 km wide magnetic anomaly in >2.79 Ga old granite-greenstone rocks, in an area that also hosts spectacular megabreccia outcrops in a stream-bed near the village of Setlagole (North West Province). Anhaeusser et al. (2010) describe the breccia as comprising "angular to rounded clasts of TTG gneisses, granites and granodiorites, with lesser amounts of amphibolite, calc-silicate rock and banded iron-formation, as well as unusual dark grey, irregular, mostly centimeter- to decimeter-sized clasts that show evidence of fluvial behavior [we think that this could imply that the clasts are partially rounded] and plastic deformation during incorporation into the breccia." The largest clasts reach several meters in size.

Interestingly, Anhaeusser et al. also report evidence of fluvial transport in the form of thin sandy to gritty layers that exhibit crude bedding and upward-fining. The breccia matrix is said to be variable but dominated by angular mineral clasts. No lava, dolomite or quartzite clasts have been observed yet, and this is taken as evidence that the breccia formed prior to the deposition of the Neoarchean Ventersdorp and Early Proterozoic Transvaal supergroups. Lower greenschist metamorphic grade of the clasts suggests that the breccia postdates the 2.79 Ga amphibolite-grade regional metamorphism. The age of the breccia is, thus, bracketed between this upper limit and the age of the Ventersdorp Supergroup (ca. 2.71 Ga, Armstrong et al., 1991). The fact that the magnetic anomaly interrupts a host of dikes of possible Ventersdorp age seems to support this chronological constraint as well.

Unusual, plastically deformed, dark clasts with altered matrix are suggested as bodies of original impact melt rock. Importantly, however, no unequivocal shock-diagnostic evidence for impact has been reported yet from the breccia itself or the granitic gneisses outcropping in the central part of the magnetic anomaly. Anhaeusser et al. (2010) advance the hypothesis that the megabreccia could represent a mass or debris flow in a marine setting triggered by an impact tsunami or resurge. Other genetic modes (diapirs, fold interference patterns, volcanism, glacial deposition of the breccia) did not find favor with the authors due to lack of supporting geological evidence in this region of the western Kaapvaal craton. Consequently, an impact origin remains the preferred mode of origin in the opinion of these authors.

This extensive breccia occurrence is certainly intriguing. Further investigation of the region should be carried out, and besides searching for shock deformed quartz, shocked zircon or monazite should be considered as possible tracers of impact as well.

6.4.33. Sinamwenda, Zimbabwe

A small, only 200 m wide, near-circular depression near the southern shore of Lake Kariba, at 17°11′42″S/27°47′30″E, in Zimbabwe was proposed as an impact structure by Master et al. (1995). The site is located about 4.8 km south-southwest of the Sinamwenda Research Station. It is formed in an area underlain by Middle Triassic sandstones and grits of the Escarpment Grit Formation of the Upper Karoo Supergroup. Master et al. (1995) arrived at the conclusion that this depression could result from meteorite impact on the basis of the following observations: crater-like morphology, overturned bedding along the rim, abundant allegedly striated joints, and enhanced microdeformation of crater rim samples. Regarding the latter, we note that no comparison with sandstone samples collected away from the site was reported.

Originally spotted on aerial photographs, the structure was first visited by a student excursion led by one of the authors of this report, Clive Stowe, whose mapping had to be abandoned when he and his team were chased away by elephants. Only in 1994 did three others (Master, Walsh, and Robertson) manage to visit the site. They carried out a magnetic survey and collected a suite of sandstone samples. They found that the unbreached crater rim is elevated a few meters above the surrounding sandstone plateau. The interior of the structure is filled by younger sediments that possibly represent Cenozoic eolian deposits of the Kalahari Group. The first visit in 1970 revealed steeply dipping and overturned outcrops of sandstone in the northwestern rim section, that could be linked to the Escarpment Grit Formation and that was overlain by stratigraphically older shales of the Upper Madumabisa Mudstone Formation. In 1994, numerous sets of joints with variably oriented parallel striations were observed in outcrops of the northern and eastern rim, and likened by Master et al. (1995) to the MSJS fracture phenomenon of Nicolaysen and Reimold (1999). These so-called MSJS, first described in collar rocks from the Vredefort Dome, are an integral aspect of the shatter cone development, and it has been, inter alia, observed in the Sudbury Structure. Such striated joints were not observed in the flat exposures of sandstone in the environs of Sinamwenda. No bona fide shatter cones have been reported from Sinamwenda.

The magnetometer traverse only revealed a 2 nT variance of the data across the structure; it failed to yield an appreciable anomaly. Master et al. compared with the magnetic work at the *Save* structures in southern Zimbabwe (see section 6.5.18, on *Zimbabwean structures* below) and concluded that this lack of an anomaly refutes the possibility that Sinamwenda could be the site of a volcanic plug – which is in keeping with the total absence of volcanic products in the area around the structure. Finally, petrographic analysis of Sinamwenda sandstone samples did not yield evidence of bona fide shock metamorphism, but it looks as if crater samples have relatively higher proportions of fractured quartz grains than samples from further away. Pockets and deformation bands of cataclastic breccia are quite abundant in crater samples. Consequently, Master et al. (1995) preferred an impact origin for Sinamwenda, although the final proof is still outstanding.

Master and Robertson (2009) reiterated this information and discussed the magnetic traverse further – as being significant in that it does not suggest the presence of a magnetic volcanic pipe below the Sinamwenda structure.

6.4.34. Temimichat Ghallaman, Mauritania

A seemingly hexagonal, 700 m wide structure named Temimichat Ghallaman occurs at 24°15′N/09°39′W in Mauretania, about 150 km northeast of the Tenoumer impact crater. Its slightly polygonal shape (Pomerol, 1967; Rossi et al., 2003) resembles the geometries of the Tswaing and Meteor craters that have been explained by analyzing the regional strain field and linking the angular geometries of these structures with prominent joint orientations in regional geology (Poelchau et al., 2009). A gravity study of Temimichat Ghallaman by Fudali and Cassidy (1972) indicated that the structure is rather shallow compared to other impact structures. However, this alone is not sufficient indication to discount a possible impact origin for Temimichat; notably the very interesting Colônia structure in Brazil was recently described with a rather shallow geometry (Riccomini et al., 2011), and it was suggested that a significant degree of erosion of an originally normal-shaped impact crater may be responsible for this shallow form.

The Temimichat Ghallaman structure was referred early on by Monod (1954), together with other crater-like features in Mauritania. Pomerol (1967) reported the occurrence of basaltic rocks in the area. No definite evidence of shock metamorphism has so far been found at Temimichat Ghallaman. Rossi et al. (2003) reported new observations from this structure. According to them, the basement rocks include granitoids and gabbros. Rim height ranges from a few meters to a few tens of meters. The structure is clearly eroded to a considerable degree: the crater rim is discontinuous, rounded, and partially covered by eolian deposits. Rossi et al. interpreted the hexagonal appearance of the crater as the result of differential erosion, with low-lying parts of the rim corresponding to the presence of gabbroic dikes (composed of mostly plagioclase and amphibole) that may be less resistant to erosion than the granitoid basement.

The crater interior is entirely covered by eolian sand. Rossi et al. (2003) observed vein structures in the granitoid rim sections that they compared with "pseudotachylite". While they use this genetic term (equivalent to friction melt rock), they state that they apply it in a non-genetic way; but as the term "pseudotachylite" is synonymous to "friction melt", a non-genetic application is obviously non-sensical. These veins are said to only occur in the rim of the crater-like structure, not in the surrounding occurrences of granitoid. The authors report that these veins of dark material are related to faults and narrow shear zones and contain glassy material with fluidal texture, and that shear zones apparently contain melted granitic clasts of mm to cm size.

Outside of the structure no ejecta blanket is apparent; only eolian and fluvial deposits were noted by these authors. Sporadic large blocks also occurred in the wider environs of the crater-like structure. Rossi et al. consider the significant degree of degradation of the crater structure to be indicative of a rather old age of formation, older than the much better preserved Tenoumer structure, for example, which was formed in similar target lithology.

Rossi et al. considered the occurrence of pseudotachylitic veinlets as possible evidence of impact, as they could only observe it along the rim exposures. No definite evidence for impact has been recorded to date, and we must emphasize again that pseudotachylitic breccias of uncertain genesis do *not* represent diagnostic evidence of impact. Nevertheless, further work on these veins and at the crater structure is warranted, and Temimichat Ghallaman does seem to be a good prospect for possibly obtaining evidence in support of an impact origin.

6.4.35. Terhazza, Mali

Rossi (2002) noted a 1 km wide structure in Mali on ASTER imagery, at 23°N and 6°W. This site is located several hundred kilometers to the east of the El Mrayer structure proposed by the same author. Terhazza is located in Cambro-Ordovician basement. It has a morphology that suggests a complex, concentric structure. Rossi suggests that this observation, in the light of the small diameter, may indicate that it actually pertains to a central uplift feature of a possibly much larger impact structure. Like at *El Mrayer* (see above), longitudinal dunes are prominent in the vicinity of Terhazza. No information about local geological investigations has been published yet.

6.4.36. Tigraou, Algeria

A very interesting prospect is the 1 km wide Tigraou structure (Fig. 52) at 35°02′32″N/1°54′12″W in northwestern Algeria, near the border to Morocco. Kock (1901) was the first to draw attention to the feature but interpreted the nearly circular, about 1 km wide,

Fig. 52. Panoramic view of the Tigraou crater structure, with its prominent limestone rim. Image courtesy of Charaf Chabou, Ferhat Abbas University, Setif, Algeria. The structure is about 1 km in diameter.

bowl-shape structure as the product of volcanism. Sadran (1958) noted the absence of volcanic products and related the structure to "crypto-explosion".

Recently, Chabou (2011) investigated this crater structure of cryptic origin. He reports the crater-like structure to occur in Early Jurassic limestones. The southwestern rim rises 75 m above the floor of the structure; the northern and eastern rim segments have been eroded. Eolian deposits cover the flat central part of the crater floor. At the top of the southwestern rim the limestone strata are oriented about vertical, and, locally in a quarry, they are said to be even overturned. Both monomict and polymict breccias are described along the lower crater rim. The main component of these is formed by angular limestone clasts. Chabou (2011) also reports the occurrence of a melt breccia close to the crater floor, comprising a fluidal-textured glassy matrix surrounding mantled inclusions of brownish flow-banded glass around limestone fragments. The matrix is said to contain abundant quartz particles and local schlieren of silica glass. These observations are intriguing, but further work is required to obtain unambiguous evidence of an origin of the structure by impact.

6.4.37. Tmisan, Libya

Dunford (2008) and Dunford and Koeberl (2009) reported on two possible impact craters in western Libya. They identified these structures from Landsat 7 and SRTM data. Interestingly they reported also that RADARSAT-1 data were not suitable for a study of small structures such as these candidates, named Tmisan and *Jaraminah* (see above, section 6.4.14), at 3.2 and 2.8 km size, respectively.

Tmisan is located at 27.423°N and 13.407°E, about 15 km southeast of a village of the same name in the municipality of Ash Shati, some 110 km east-northeast of the city of Sabha. The outer diameter of 3.2 km is indicated by a set of nearly circular but somewhat undulatory ring forms, 500–700 m wide, thought to represent upturned and truncated strata. The dip direction may be outward. The inner part of the structure comprises a somewhat irregularly shaped plain of 600–700 m width. Some radially oriented ridges seemingly extend from the center. In the central area is an apparent topographic high with a scalloped rim, which may be plateau-like and has a width between 600 and 800 m. The Tmisan structure seems to be located in the B'ir al Qasr Formation of Eifelian age. This formation consists of a lower fine-grained sandstone unit that is overlain by coarser-grained sandstones. The formation is overlain by a conglomerate bed and, finally, a coarsening upward succession of sediments of the Idri Formation.

Forty samples collected by A. Dunford from the Tmisan site were investigated by optical microscopy. No definitive evidence of shock deformation, such as PDF, was identified. Planar fractures could be present though, and Dunford (2008) mentions the observation of shatter cone-like features in the field. It remains entirely open though, whether these observations could relate to ventifacts. In conclusion, in the absence of any direct evidence for an endogenic origin of this structure, and of bona fide evidence of impact, the Tmisan structure remains in the category for possible impact structures.

6.4.38. Tongo, Cameroon

Through personal communication from Tiolo A. Temenou of the Université de Yaoundé, Cameroon, we learnt that a further site in that country had been considered an impact structure and had been the subject of a MSc thesis (Temenou, 2010). The location of interest is in the western part of Cameroon, in the district of Nord Makombé, at 4°51′–5°00′N/10°21′–10°31′E, northwest of the village of Ndikiniméki. From a digital elevation model supplied by T.A. Temenou, the proposed structure is somewhat ovoid shaped and 10 × 12.5 km large. A series of partially curved ranges forms the outer rim zone of this area. An annotated landscape image is marked with the location of a "central uplift", but we fail to see this expression in the digital elevation model. A topographic profile from NE to SW across the structure shows prominent, up to 500 m high outer terrain. In the interior of the structure several prominent, up to 350 m high, elevations occur, but they extend over the entire floor of the structure. A series of steep faults dissects the inner part of the structure. The inner slope of the SW rim shows a gradual slope inward over some 4 km distance.

The ovoid structure is prominent on drainage maps, where a strong radial and inward directed pattern characterizes the structure itself. However, there are a number of drainage lines that breach the outer "rim" section, with the most prominent drainages from the northeast and north and outflowing in the south/southeast. The alleged structure is located in granitoids of the Precambrian basement. Various kinds of breccias are described in the thesis, but evidence at hand does not allow to judge whether this could be related to an impact event. One interesting image of a hand specimen could represent a melt rock or a mylonitic rock. In terms of microscopic evidence, Temenou cites shock metamorphic effects: kink bands in biotite and PDF in quartz. While the former is not of shock-diagnostic value, images supplied of the latter are in part interesting but remain inconclusive. Recently, several rock specimens obtained from Temenou were studied by optical microscopy by one of us (WUR), who failed to detect any petrographic evidence supporting shock metamorphic deformation. Further investigation of the breccias and of the possible shock metamorphic features is desirable, but at present it is not possible to classify Tongo as a confirmed impact structure.

6.4.39. Unnamed Geophysical Anomaly, Okavango Delta, Botswana

Based on aeromagnetic data interpretation, Cooper et al. (2010) proposed the possible occurrence of a sizable impact structure in the Okavango Delta of Botswana, centered at 19°07′40″S/23°18′12.7E. The ca. 15–20 km wide, pre-Cenozoic, apparently complex structure is located in Neoproterozoic basement that is expected to be composed of gneisses and massive granites containing also some migmatized amphibolitic gneisses. It is buried beneath Paleogene to Holocene Kalahari Group sediments (Partridge et al., 2006).

The authors interpret the aeromagnetic data to reveal a possibly 15–20 km wide structure in the form of a circular region with a subdued magnetic intensity that also reveals a central magnetic peak characterized by an amplitude >700 nT. The feature is not visible on either GoogleEarth imagery or in Shuttle Radar Topography Mission (SRTM) data. It is interesting to note that the regional magnetic fabric is continuous across the anomaly that delineates the possible structure. Cooper et al. (2010) modelled the anomaly and from that suggest that the actual structure could be covered by 200–500 m of sediment. This model also indicates that the inferred central uplift could be ∼5 km wide.

The fact that the regional magnetics is, in subdued form, continuous across the interpreted structure is taken by the authors to suggest that the circular feature is not a deeply rooted structure, such as a ring-dike, pluton, or other igneous feature, but rather that it is restricted to the upper basement. The interpretation that this anomaly could represent an impact structure is based on the complex anomaly pattern with a central positive anomaly that mirrors the geometry of a complex impact structure. Age constraints for this proposed structure are limited, with a minimum age being indicated by the Tertiary upper age of Kalahari Group sediments. An upper age limit is given by regional fabrics related by Miller (2008b) to the Damara orogeny.

A detailed gravity study over the region of the proposed impact structure might provide supporting evidence, short of final confirmation that would obviously require drilling into the central anomaly.

6.4.40. Unnamed, Angola

Roger Swart of Blackgold Geosciences, Windhoek, Namibia, recently communicated the GoogleEarth image of an intriguing structure in Southern Angola, at 15°12′07″S and 12°45′8″E. The somewhat ovoid, ca. 900 m long (in NE–SW direction) structure (Fig. 53a) is located in Precambrian schists. No constraint on its age exists, and so far it appears as if it has not been visited on the ground. The appearance in this image suggests that it is an old, degraded structure. It is awaiting ground-based investigation.

6.4.41. Unnamed, Libya

Seismic evidence for a possible impact structure in the Al Hamada Al-Hamra Basin of Libya and obtained during hydrocarbon exploration was mentioned by Ben Musa and Baegi (2009). Two seismic lines denoted NG 238 and NG 464 allegedly show a crater-like structure on depth-contour maps near the top of the Awaynat Owenin Boundary and on the top of the Cambrian and Ordovician formations of the basin. They reported that the alleged structure had a similar morphology to that of a complex impact structure, with a raised rim and annular synform and a central uplift feature. The diameter of the structure is given as 2 km and an estimate of 300 m uplift for the strata of the central uplift is reported. They suggest a post-Carboniferous age, as mainly Late Carboniferous and Triassic rocks are disturbed in the area concerned.

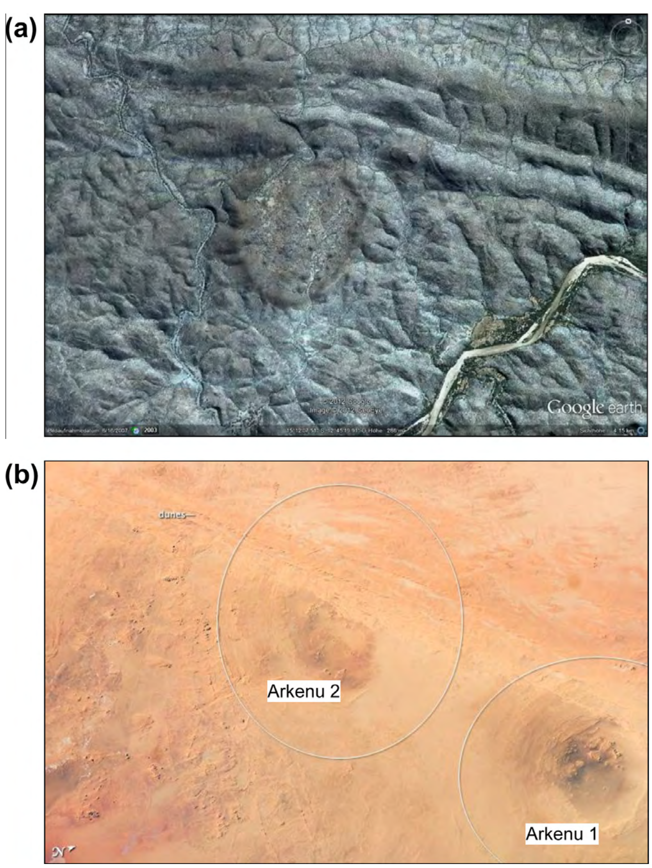


Fig. 53. (a) GoogleEarth image of the Unnamed Structure in Angola, courtesy of Roger Swart, Windhoek (Namibia). The structure measures about 0.9 km in the long (NE–SW) direction. (b) The Arkenu structures of southeastern Libya. Circles indicate the originally estimated sizes of the structures of 6.8 and 10.3 km. Width of area shown ca. 25 km. Astronaut photograph ISS017-E-20538 of October 2008. Courtesy NASA Earth Observatory (http://www.earthobservatory.nasa.gov/).

It appears that only recovery of samples from this structure in the course of drilling will be able to verify the nature of this intriguing feature. No further information regarding the location of this structure is available.

6.4.42. Unnamed, Sudan

Di Achille (2004) applied varied satellite imagery in a regional study of northeastern Sudan. At 37°55′E/17°57′N, close to the border between Sudan and Eritrea and near the Red Sea, he observed a nearly circular structure that interrupts the regional N–S trending fabric of the Baraka Suture. Basement in this region of the Red Sea Hills is composed of volcano–sedimentary rocks of the Neoproterozoic Nubian Shield. The circular feature is some 5.5–6 km in diameter and is obviously covered by sediments. Parts of the rim at the eastern and western limits of the structure are collapsed. Two drainage channels breach the rim in the northern and western parts of the structure. The somewhat asymmetric appearance of the structure is thought by Di Achille to be the result of erosion. The elevation from the floor of the structure to the rim is of the order of 350 m and there seems to be a central elevated area that is raised by some 75 m above the surrounding terrain. This observation is interpreted by the author to suggest a remnant of a central uplift. While he considers the possibility that this circular structure could be an eroded, complex impact structure, he also remains conscious of the fact that the volcano-sedimentary rocks of the regional geological setting might be related to this structure. A comparison with the generally larger volcanic features of the region and the typically smaller size of maars lead him to conclude that the most reasonable explanation for this crater-like feature was found in an impact hypothesis.

6.4.43. Unnamed craters and ring complexes, Sudan

Sparavigna (2010a) placed remote sensing observations (survey of satellite data, via Google MAPS and ACME Mapper) onto the internet, in which she claims – on morphological grounds – the possibility that two circular features within a host of crater-like structures and ring complexes – in the Bayuda Desert and north of the Nakasib Suture could represent impact craters. One of these is located at 19°12′47″N/35°59′00″E, and is reported to resemble a ring or donut of ca. 3 km width. The other structure was located at ca. 21°17′25″N and 33°30.22′E and was 6–7 km wide. No ground-truthing has taken place since.

Furthermore, in the Bayuda Desert region, a 10 km wide structure was identified by Sparavigna (2010b), with locations given as 18°3′25.52′N and 33°30.22′E, some 40 km west of Berber town. While Sparavigna remained ambiguous about the origin of this feature, McNally (2010) preferred this to be a previously undiscovered impact crater. Here, too, ground-truthing is lacking entirely. However, a discussion on the internet (http://www.theepistlesofpaul.blogspot.com/2011/11/possible-sudanese-extraterrestrial.html) has been provoked by Sparavigna's suggestion. In the context of published geological information about the Bayuda Desert area concerned (Barth and Meinhold, 1979, 1981; Barth et al., 1983; Woolley, 2001), it appears that the proposed impact structure could well be of volcanic origin and might correspond to the Singeir ring complex of Woolley's.

6.4.44. Unnamed, Tunisia

A crater-like structure of simple bowl-shape was noted by Tomlinson (1999) on a Landsat image of Central Tunisia, at 35°45′N and 09°08′E. The site is located some 12–15 km from the town of Makthar. The structure is about 5 km wide. Tomlinson describes a well-defined rim on the northern and eastern sides of the structure, which is also said to be affected by small-scale faulting. The other sides of the crater-like structure are less well-defined and extensively modified by larger scale faults, other linear features, and possibly by slumping. Strange for a structure of this size is that no evidence of a central uplift or of concentric ring structures was observed. Apparently, the structure is also weakly evident on the local topographic map sheet by a slight perturbation of the contour lines and a "circular" deflection of a cross-cutting stream-bed. On the geological map of the region, the structure is shown as an isolated outcrop of Mid-Cretaceous marine sediments of polygonal shape. Obviously, the structure deserves to be investigated. Youbi et al. (2011) referred to this site as *Jebel al Bukrah*.

6.4.45. Uri, Chad

In addition to the *Oun* structure discussed above, González and Alonso (2006) reported on a second structure identified by remote sensing analysis. The Uri structure of about 5 km diameter is located at 21°17′N/19°20′E, in a quite mountainous terrain. Like its "cousin" Oun, Uri displays a multi-ring morphology. It is said to be more eroded than Oun. There are two concentric rings, separated by a ring depression, and the authors even hint at the possible existence of another ring feature at 4 km from the center. Linear fractures in the rings are reported to be associated with volcanic build-ups, which to us suggests that this structure may well represent an igneous complex, but which is interpreted by González and Alonso as indicating post-impact volcanic activity. This needs to be resolved through detailed ground-based geological analysis.

6.4.46. Velingara, Senegal

Master et al. (1999) first proposed that a 48 km diameter, alleged multi-ring feature, termed the Velingara structure, in Senegal could represent a buried impact structure. Velingara was identified on Landsat MSS and TM images, and in NOAA-AVHRR imagery, and is centered at 13°02′13.2 N/14°7′40″W in southern Senegal. The center of the structure is located about 12 km south-southwest of the town of Velingara in Haute Casamance Province. The northern margin of the structure abuts against the border with Gambia, whereas the southern limit is about 20 km north of the border with Guinea-Bissau. The area is typically flat and has only low (<70 m.a.s.l.) elevations, and is covered by thick ferruginous laterites. There are essentially no outcrops. Master et al. (ibid) reported that the structure was formed in Mid Eocene marine sediments of the coastal Senegal Basin and was buried by up to 90 m of post-Eocene continental sediments. Drilling and a resistivity survey support that Neoproterozoic or Paleozoic basement of the Mauritanide Belt is subcropping in the central parts. This has been taken by this author as an indication of basement uplift in the central part of the structure. A positive Bouguer gravity anomaly is associated with the northern part of the structure. Master et al. (1999) suggested that Velingara could be a shallowly buried complex impact structure with a central uplift, and that, if this were confirmed, this structure may be of similar age to several other impact structures formed in the Late Eocene.

Master et al. (1999) also drew on similarity with the remote sensing expression of the 40 km wide Brazilian Araguainha impact structure (Lana et al., 2007) that has a sizable central uplift. The central part of the Velingara structure is a broad depression, the Anambé basin, into which centripetal drainage flows, which creates a swamp in the central area. Drilling for phosphate exploration in the central part of the structure revealed that basement occurs at shallow depth, below only 3 m of sand cover. Several samples studied petrographically did not reveal any bona fide shock deformation effects.

A follow-up on the initial report by Master et al. was presented by Wade et al. (2002), who focused on Synthetic Aperture Radar (SAR) interferometry that yielded a digital elevation model enhancing the morphological detail of the structure (such as the centripetal drainage system). This was also exploited in a FOCUS EARTH report by Wade et al. (2001a, www.esa.int/esapub/bulletin/.../bul106_13.pdf). Wade et al. (2006, 2001b) reported discovery of possible shocked quartz in laterite overlying the Velingara structure – a report that needs to be confirmed.

6.5. Disproven candidates

6.5.1. Aflou, Algeria

The Aflou structure in Algeria, located at 34°00′N/02°03′E about 12 km SSW of the village of Aflou and some 80 km WNW of Laghouat represents an oval depression of about 3 × 5 km extent. Lambert et al. (1980) discussed this structure as being faintly visible on aerial imagery. They visited the structure as it had been referred as a possible impact structure by Marks et al. (1972), who arrived at this conclusion because they failed to detect any other obvious cause for the existence of this feature. They had also collected what they termed "vitreous" samples in the vicinity of Aflou, which they then interpreted as impact melt. Lambert et al. (1980) reported that Aflou occurred within Cretaceous sandstones and conglomerates of the Ed Dor formation. The 1:200,000 aeromagnetic map of the region showed a well-defined circular anomaly centered on the Aflou structure.

They could not report either a continuous rim at Aflou, nor evidence of disturbance of the strata, or any shock effects in the rocks of the walls. The strata exposed in the walls of the structure have variably inward and outward dips in the different sectors of the structure. No breccias were found. The rocks that intrigued Marks et al. (1972) were identified as basaltic to andesitic volcanics – diabases that underwent retrograde metamorphism. Lambert et al. (1980) concluded that these volcanics are younger than the sedimentary strata filling the structure, and – consequently – younger than the formation of the elongated structure. Thus, an interpretation of Aflou as an impact structure and the igneous rocks as impact melt is not reconcilable.

Lambert et al. (1980) drew on structurally controlled magmatic activity in the region. They reported that the formation of Aflou may be purely erosional at a site where local geology represented a weakened crustal spot at the intersection of several structural trends. An alternate mode of formation could be dissolution (karstification) in the underlying El Richa limestone or gypsum units. Marks et al. (1972) had suggested a relatively young age for the structure, perhaps upper Tertiary. The regional magmatic activity may well be of late Tertiary or Quaternary age.

6.5.2. Al Mouilah, Algeria

The Al Mouilah (also known as El Mouilah) structure of 4.5 km diameter and 130 m depth is located at 33°51′N/02°03′E, 80 km west of Laghouat, in Algeria. It was investigated by Lambert et al. (1980). They selected this feature for its morphological appearance: it is nearly completely circular, has steep walls and even a central dome. These authors even likened the appearance of Al Mouilah to the geometry of the Steinheim basin impact structure in Germany, although the Algerian feature does not have a similar-sized central uplift. And yet, they could not find any field or laboratory evidence that would permit to conclude that this structure could be the result of impact: they failed to observe a raised rim and could not detect any breccia occurrences, or other deformation in the massive limestone (and minor sandstone) that forms the "crater" wall. The strata exposed in the crater wall consistently show shallow (5–15°) dips outward. Sandstones exposed in the central "dome" are flat-lying or gently dipping, similar to the regional strata. Again, Lambert et al. (1980) found that these rocks are undisturbed/undeformed. The authors interpreted this large crater-like structure as the result of a recent collapse event "due to dissolution in the thick underlying limestone and gypsum formations" or that it could have been caused by "purely erosional" processes. Evidence for volcanic activity in the area of this structure was not detected. While the cause of formation of this structure has not been proven yet, the analysis by Lambert et al. (1980) has conclusively shown that there is no evidence for impact, so that this structure has been categorized as a disproven impact structure.

6.5.3. Arkenu structures, Libya

Paillou et al. (2003) alleged the existence of two impact structures in Libya, at 22.1°N/23.8°E (Arkenu 1) and 22.05°N/23.72°E (Arkenu 2), some 250 km south of Al Kufrah oasis on the eastern margin of the Al Kufrah basin. They had first noted the two somewhat circular features on satellite imagery and then followed up with a field visit. The structures (Fig. 53b) were reported to be 10.3 and 6.8 km in diameter, and their centers were about 10 km apart.

As evidence of impact, the somewhat circular morphologies, presence of shatter cones, and presence of planar deformation features in quartz of alleged impact breccias were given by Paillou et al. On this basis, Arkenu 1 and 2 even made it into the Earth Impact Data Base (http://www.passc.net/EarthImpactDatabase/index.html) – however, for limited time only. Objections were voiced within the impact cratering community almost immediately, as neither the imagery of alleged shatter cones, nor that of presumed shock deformation, was considered satisfactory. The allegation of the presence of impact breccia was based only on the unconfirmed report of findings of shock deformation. The appearance of such breccia as presented in the discovery paper did not allow to elucidate an origin by impact or other terrestrial process. Significantly, Paillou and co-workers have not followed up since with additional evidence that could have possibly confirmed their initial allegations.

In contrast, Di Martino et al. (2008) reported about a field visit to the Arkenu area. They could not detect any shock deformation in quartz and reported that the shatter cones alleged by Paillou et al. were the result of wind erosion (ventifacts). They categorically stated that they could not find any evidence that would support an impact origin. According to Di Martino et al. the local geology involves Paleozoic sandstones and siltstones. The strata in the areas of the structures are impregnated by Fe-oxide minerals. In Arkenu 1, the sandstones are quite well preserved in the structure's interior. In Arkenu 2 they are disaggregated and contain massive magnetite deposits. The authors interpret the local geology as the result of partial "digestion of sandstones by a subvolcanic intrusive body (now partially outcropping within the crater area)". In Arkenu 2 they observed a first mafic hypabyssal phase, followed by granite that is locally preserved in the northern sector of this structure. They concluded that these crater-like features could be the result of intrusion of two nearly cylindrical sub-volcanic pipes, which was accompanied by hydrothermal venting and dike injection.

Most recently, Cigolini et al. (2012) reported further evidence in support of a volcanic genesis of the Arkenu bodies from field and petrographic work. They did not observe shock metamorphic evidence in samples of sandstone from the Arkenu circular structures and also state clearly that the alleged shatter cones have an origin as wind-ablation features. They support the conclusion that the two features represent volcanic stocks and interpret their existence as a consequence of intrusion of syenitic porphyritic rocks into the Nubia Formation sandstone. These volcanics are part of a "rather simple and eroded ring complex". They make a case for hydrothermal activity subsequent to volcanic intrusion, which deposited massive magnetite-hematite, coeval with the emplacement of silicified dikes in the environs. Finally, they observed "plugs of tephritic-phonolitic rocks and lamprophyres (monchiquites) inject[ed into] the Nubian sandstone along conjugate fracture zones...".

Baegi and Ben Musa (2009) also discredited the alleged Arkenu impact structures and referred to their magmatic geology. Based on these recent developments, Arkenu has now been removed from the Earth Impact Database. The history of this posting, however, poses a clear warning that even a serious impact crater database should be used with caution – and cited evidence ought to be carefully scrutinized.

6.5.4. Bir Anzarane and Co – The Moroccan record

Since 2011, a concerted effort to investigate sites of possible impact structures (or determine criteria for the discrimination of possible structures first identified by remote sensing) has been made by staff and students of the Hassan II Casablanca University (Chaabout et al., 2011), in cooperation with French and German impact workers. So far, only a single site with evidence of impact deformation has been proposed (Sadilenko et al., 2013 – see section 6.1.1 above, *Agoudal* – occurrence of shatter cones) in the territory of Morocco of considerable size (712,550 km^2). This suggests – in comparison with the record of, for example, neighboring Algeria (2,381,741 km^2, 4 confirmed impact structures) – that a number of impact structures may remain to be discovered in Morocco.

Starting out with meticulous screening of GoogleEarth and YahooMaps imagery, around twenty potential sites of interest were recorded based on their circular or subcircular morphology. These locations (Fig. 54a) have then been scrutinized against available topographic and geological information, which lead to elimination of a number of features in the Timahdit/Michelifene region of north-central Morocco, where intense Jurassic volcanism is recorded, and in the area around the town of Smara in southeastern Morocco where mud mounts had been mistaken for possible impact structures. Several of the investigated sites are featured in Fig. 54b–d.

First ground-truthing was done in June 2011 at Bir Anzarane, a ca. 1.5 km wide feature located at 23°3′13.59″N/15°23′35.33″W in the desert of the Moroccan Sahara territory, southeast of Dakhla town. The Bir Anzarane structure seemed promising because of its high degree of circularity over more than half of its circumference – with the remaining sector being obscured by dunes. However, initial field observations (poor morphological expression, similar strata orientation in the center and in the area along the perceived rim, lack of breccias and any other possible indication for impact-related deformation, and, particularly, a pattern of light and dark areas seen in GoogleEarth likely being the result of ferruginous quartzite scree and variably grass and brush-vegetated patches), as well as complete lack of indications of shock deformation in quartz-rich mylonites from this area, suggest that this target should also be excluded. Further evaluation of a second, smaller circular observation to the northeast of Bir Anzarane was obstructed by the perceived danger from landmines in this area close to the border with Mauretania.

In the course of two expeditions in the first half of 2013, five further sites identified by S. Chaabout were investigated by a Moroccan-German team. The first (Tetouan, 35°35′01.30″N 5°24′19.68″W) was revealed as a limestone quarry. On the recommendation of Mr. Abdelmagid Abatalib (Casablanca) a second structure at 30°18′5.06″N, 7°31′23.950″W near Taznaght town was visited. It turned out to be a site of preferential erosion within a major fault zone. Tectonically produced cataclasite characterizes the interior of this feature. Less than 200 m from this fault, adjacent to a tight bend in the river, is a near-circular quarry that was also mistaken as a possible impact crater.

Travelling towards Imilchil, at +31°45′46.80″, −5°30′41.58″ a small, absolutely circular feature was observed. Close inspection of this 20 m wide and 1.5 m deep structure, as well as confirmation sought from local people, revealed this to be a location for threshing wheat by a donkey walking around a pole placed at the center of this feature.

In 2012 newspapers and other media in Morocco alleged the occurrence of two impact craters at 32°13′1.23″N, 5°32′28.07″W (Lake Isli, Fig. 54b) and 32°11′46.29″N, 5°38′11.80″W (Lake Tislit, Fig. 54c). The argument put forward by a researcher from Agadir University included circularity of these lake structures, relative vicinity (23 km) to the area of the Agoudal meteorite find around N31°59′4″N, 5°30′55″W, and alleged shock deformation in the form of PDFs in quartz in samples from the lake areas (Ibhi et al., 2013; Wikipedia: http://www.fr.wikipedia.org/wiki/Double_crat%C3%A8re_d'Imilchil_et_de_Tislit).

The recent expedition by staff and students from Casablanca University in conjunction with one of us (WUR) revealed no lithology containing quartz grains of sufficient size to possibly identify planar deformation features. In fact, no shock deformation was observed at all in any of the samples investigated. The rocks making up the areas around both lakes comprise limestone, marl, and mudstone. Shatter cones were not observed. To our knowledge, no Agoudal meteorites have been found to date near the lakes. The closest known strewn field is that of the Agoudal iron meteorite (Chennaoui Aoudjahane et al., 2013; and above: *Agoudal*). Results of detailed mapping proved that there is no significant rock deformation in the environs of the lake that could be linked with a cataclysmic origin of these two structures. They also force the conclusion that the previous interpretation that the lakes were formed within a synclinal basin setting as a result of tectonics (Michard et al., 2011; Ibouh, 2004) is sound. A detailed paleo-climatic study of Lake Isli was published by Zeroual (2001). A first report of our field results has been published by Chaabout et al. (2013). *In conclusion, to date no unambiguous evidence for the existence of impact structures at Isli and Tislit lakes has been reported.*

Finally, a circular structure of about 1 km diameter at Michelifene (33°24′47.46″N, 5°4′42.64″W; Fig. 54d) was investigated. It had previously been studied by, e.g., Martin (1981) and Gigout (1955). It has long been known as a volcanic explosion crater. The country rock consists of limestone that is locally slightly deformed (down warped, perhaps as the result of slumping, and fractured. Breccia occurs prominently in the south-eastern sector of the structure. It comprises a carbonate-rich, microclast-bearing, fine-grained groundmass carrying a population of lithic clasts (limestone, volcanic lithology) and melt bodies. In the main breccia mass, the lithics may reach sizes from millimeters to 60 cm, in addition to blocks of limestone of many decameter sizes. Melt bodies were observed over the size range from millimeters to 50 cm. The shape variation of melt clasts ranges from angular to well-rounded, but also includes highly irregular bodies. At some sites further outward from the deep central depression a 10 cm to meter thick layer of similar breccia was observed that, however, is characterized by a grain size population that is strongly skewed towards smaller sizes. At the thin section scale, the same clast population observed macroscopically is found (glassy or slightly devitrified volcanic melt, besides various carbonates, with or without microfossil content). It is obvious that this breccia is related to the volcanic eruptive activity, and thus there is no evidence that the structure itself should be of impact origin.

The remaining sites in Fig. 54a have not been visited yet by this group; however, there is literature about them available. A large number of spectacular coral-stromatoporoid reefs of sometimes distinctly circular shapes occur in the area around Smara (Wendt and Kaufmann, 2006). A circular structure at Twihinat in the south of Morocco has been described as a volcanic megastructure with carbonatites and volcanic breccias (ONHYM, 2011a). Another structure at Lamlaga, also in the south, is also seemingly of volcanic origin. It is characterized by distinctly circular magnetic and gravity anomalies, and is known for presence of Rare Earth Element and Nb mineralization (ONHYM, 2011b). The third structure of this group, Glibat Lafhouda, is also associated with carbonatites and mineralization of Fe, Rare Earth Eelements, Nb and Ta, and U (ONHYM, 2002).

6.5.5. Bushveld Complex, South Africa

The Bushveld Complex of northern South Africa, a 500 × 350 km wide body (Fig. 55), is known as the world's largest layered igneous intrusion (e.g., Cawthorn et al., 2006) and has particular significance because of its enormous base metal and particularly platinum-group element, Cr, and V resources. Ages for the mafic and felsic formations of this huge intrusion have been summarized by Cawthorn et al. (2006). They range from 2054 to 2061 Ma (further references therein). The debate about the genesis of this unique deposit is still ongoing, with endogenic processes having been generally favored for the last decades.

Another hypothesis that has been repeatedly proposed is an origin by multiple impact, sometimes involving a combination of *Vredefort* and Bushveld impacts (Hamilton, 1970; Rhodes, 1975; Elston, 1995, 2008). The arguments given in support of this hypothesis have involved structural, stratigraphic, and mineralogical considerations, mainly focusing on some inliers of pre-Bushveld rocks as well as the roof rocks of the complex, the Rooiberg Group felsites (for a review of the Rooiberg group: Buchanan, 2006). The Rooiberg sequence has been considered to represent a sheet of impact melt and impact breccia, which was topped by sediments generated by erosion of the unstable rims of impact structures. Elston (1995) referred to paramorphs of tridymite that indicated that some melts had been superheated (as one would expect for impact melt). It should be noted that paramorphs of tridymite are also known from volcanic rocks (e.g., Green and Fitz, 1993).

Also shock deformation in quartz was alleged as evidence for impact in the Bushveld region by Labuschagne (1970). French (1990) could find no evidence of shock metamorphism in the many samples investigated by him. Also Buchanan and Reimold (1998) and Joreau et al. (1996) investigated the alleged evidence of shock metamorphism proposed by Labuschagne and found that the features proposed by him are different from shock-diagnostic planar deformation features (PDF). Joreau et al. demonstrated by transmission electron microscopy that the "Bushveld features" were in fact low-strain deformation features that could well be the result of normal tectonic activity. To date, no definite evidence of shock metamorphic deformation has been reported for rocks from the Bushveld Complex – so that the impact hypothesis for this complex should be treated with great caution.

On an aside, Reimold and Minnitt (1996) reported shatter cone-like features in Transvaal Supergroup rocks from east of the Bushveld Complex. They emphasized the similarity of these features and associated multiple joint sets to impact-generated shatter cones and multipli-striated joint sets (MSJS) but concluded that the site of observation represented a fossil water-fall and the deformation features were, in fact, *percussion marks* (also known in sedimentology as *impact marks*).

6.5.6. Delmas sinkhole, Mpumalanga Province, South Africa

On the first edition of the official 1:50,000 scale topo-cadastral map of South Africa (sheet 2628BB Kendal), a "Meteorite Hole" is indicated at coordinates S26°08′57.4″, E28°46′37.5″ (WGS84) some 10 km east of the town of Delmas. These maps were in the past also used as base maps for geological field mapping by the then Geological Survey of South Africa (now Council for Geoscience, CGS), but this inscription on the topographical map was not transferred to the official 1:250,000 scale geological map sheet 2628 East Rand that covers the area. To date no official record describing this map inscription could be found at the CGS or the Surveyor

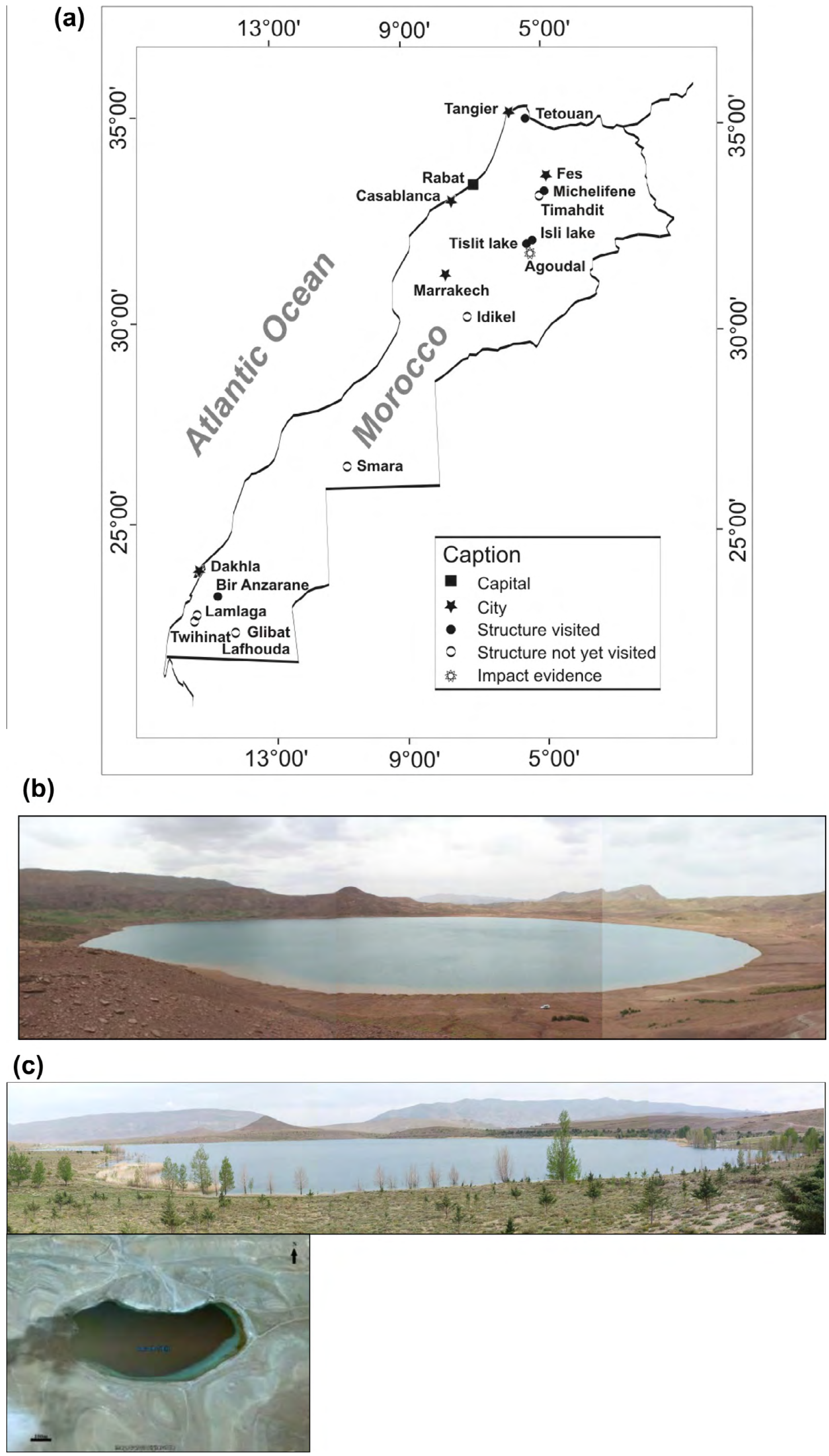


Fig. 54. (a) The geographical locations in Morocco of some circular structures under investigation as possible impact structures (by S. Chaabout, University Hassan II, Casablanca). (b) Panoramic view – towards the northeast – of Lake Isli. The near-circular lake is ca. 1.4 km wide. (c) Panoramic view of Lake Tislit, view towards the north. The lake is ca. 1 km wide. Also shown is the GoogleEarth image of Lake Tislit indicating that the lake could be considered slightly ovoid with a slightly longer east–west axis. However, it does have a kidney-shape instead, with the terrain on the northern side representing part of the northwesterly trending, folded, anticlinal structure in the upper part of the image.

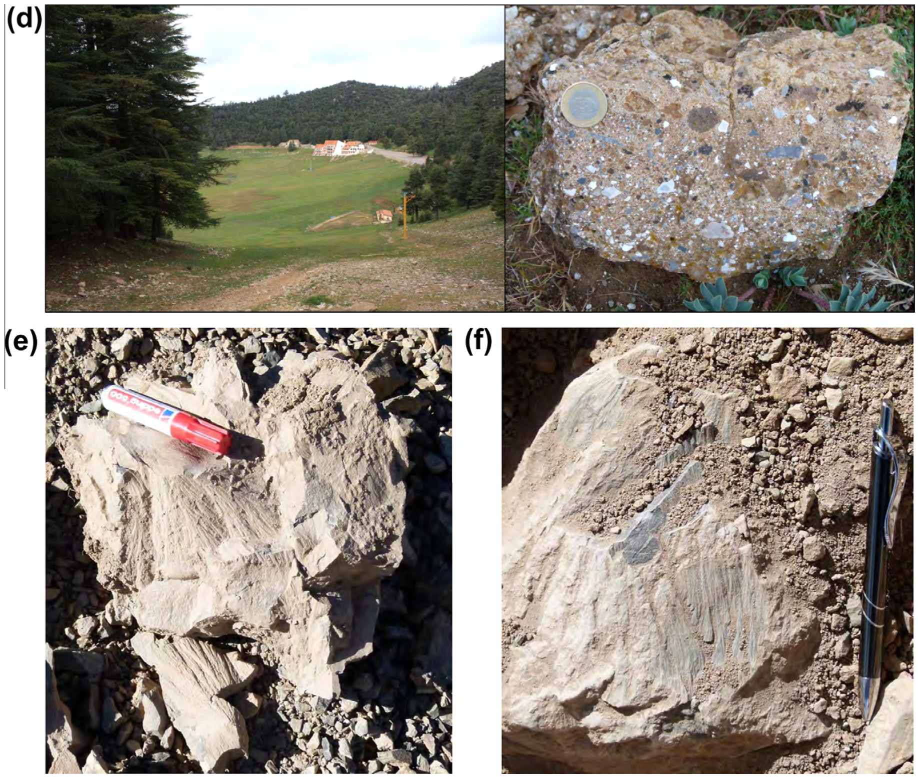


Fig. 54. (d) Partial view of Michelifene chaldera, from a limestone block on the inner limestone crater wall. The breccia shown on the right is composed of fragments of volcanic rock and limestone country rock, and cemented by a carbonate-dominated groundmass. The block shown is ca. 20 cm wide. (e and f) Two field impressions of shatter cones from the Agoudal shatter cone discovery site (pens for scale ca. 10 cm (e) and 14 cm (f) long.

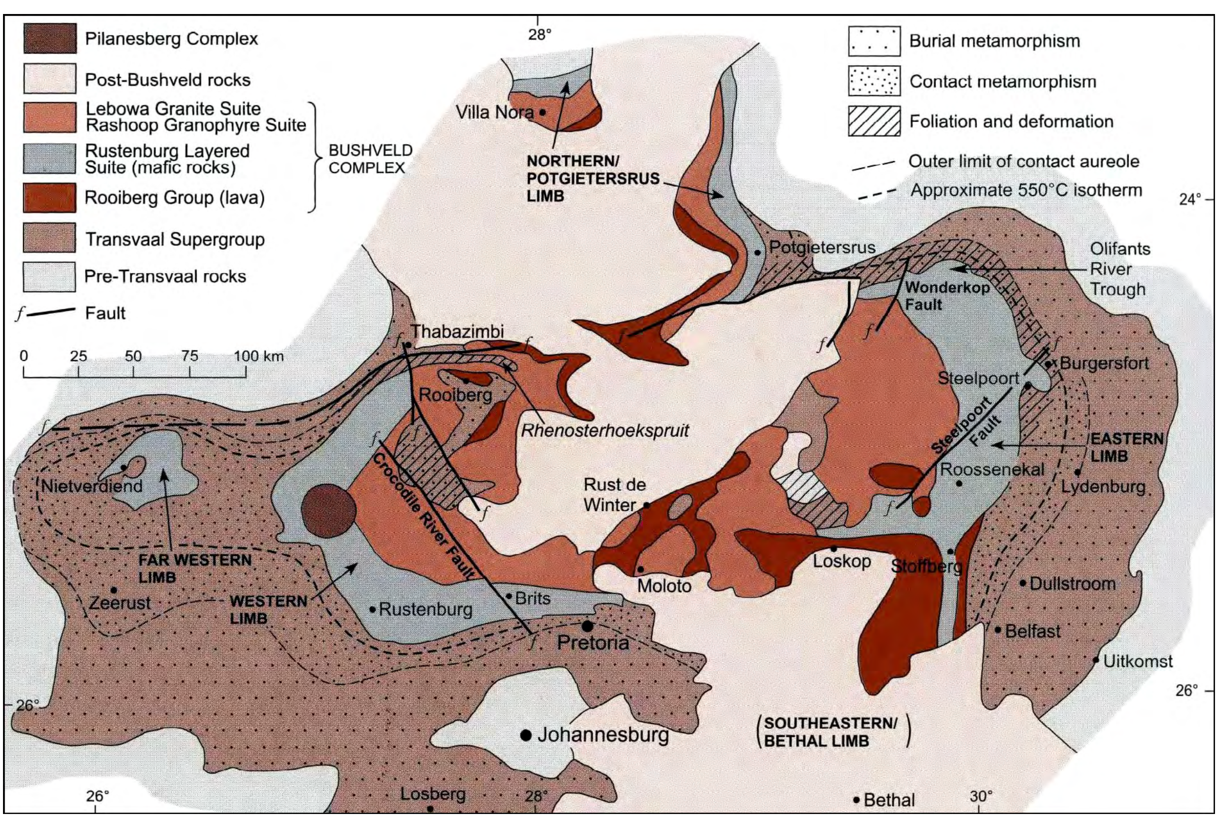


Fig. 55. Simplified geological map of the Bushveld Complex, after Cawthorn et al. (2006). Reproduced in slightly modified form with the permission of the Council for Geoscience (Pretoria) and the Geological Society of South Africa (Johannesburg).

General's offices, nor did enquiries with local farm owners reveal any evidence to explain the origin of this inscription. Attention was drawn to this notation in 2011 by Mr. Reinie Meyer of Pretoria, who also led the subsequent investigation.

The aerial photo shown in Fig. 56a reveals the concentric structure of this feature, which probably gave rise to the identification as a "meteorite hole" on the maps. Geologically the area is underlain by the ~2500 Ma old dolomite rocks of the Malmani Subgroup conformably overlain by the thick sedimentary sequence of the Transvaal Supergroup (~2600-2150 Ma). These formations are dipping at 10–20° to the NNE. The area has been subjected to an extensive period of erosion lasting almost 2000 Ma culminating during the glacial Dwyka period (ca. 300 Ma) and then followed by the deposition of the extensive cover of the Permian age Karoo arenaceous and argillaceous sequence. During this extended period of erosion the area was probably already subject to extensive karstification in the carbonaceous formations. It was proposed that a massive pure silica deposit (the Delmas silica deposit, currently mined) in close proximity to the alleged "meteorite hole" could be the result of filling of a mega-sinkhole in the dolomite formation by pure arenitic quartz beach sand during a transgression period (Martini and Horn, 1996).

The area was inspected during a visit in April 2012 guided by Reinie Meyer. The "hole" is an almost circular feature with a radius

(a)

(b)

Fig. 56. The Delmas sinkhole. (a) Overview of the apparently circular feature (note the slightly elevated ring around the central, rectangular depression. In the background part of a waste berm from the local silica plant is visible. The photograph was taken by Patrice Zaag (MfN Berlin) from another waste berm to the west of the sinkhole, which is related to an extensive coal-mining activity in the Karoo strata overlying the Transvaal Supergroup carbonates in which the sinkhole is developed. (b) The open "pit" of the sinkhole illustrating the subhorizontal layering of the carbonate rock.

of approximately 70 m and extending probably to a depth of at least 50 m in the center. All the evidence gathered during this inspection points to an old subsidence feature developed in the dolomite country rock, which subsequently developed into a large sinkhole. Near horizontal dolomite layering is still visible on the eastern side of the sinkhole (Fig. 56b), which argues strongly against the presence of a meteorite impact structure. The area is prone to the development of sinkholes due to the overexploitation of the large groundwater resources associated with the karstic terrain for agricultural purposes.

Based on the evidence gathered in the field and from some historical information gathered from local residents, it is concluded that the feature identified as a "meteorite hole" on official topographic maps is in fact a large, near-circular sinkhole that had developed over a period of probably more than a century.

6.5.7. El Baz, Egypt

A circular, crater-like feature of 4 km diameter was located by El-Baz (1981) at 24.2°N and 26.3°E on Landsat imagery in the Great Sand Sea of western Egypt. It is located to the east-southeast of the *Oasis* structure in Libya, in the midst of a vast field of longitudinal dunes. The structure is characterized by a sharp, crenulated rim crest, a terraced wall, a discontinuous inner structure of about 1.6 km width, which the author thought to be perhaps the remnants of a central uplift structure, and some blocks outside of the rim in an irregularly textured, 2 km wide swath – particularly on the east side of the crater-like structure.

The structure was interpreted to be located in sandstones of the Nubia Formation. El-Baz (1981) likened its morphology to the appearance of some impact structures on the Moon and terrestrial planets. He compared the morphology of the structure in the Western Desert with the attributes of Meteor impact crater in Arizona, but also noted that this structure may have formed by a circular diorite intrusion. After its erosion by wind, only the indurated and metamorphosed sandstone would remain to form the circular structure. In a subsequent paper, El-Baz and Issawi (1982) related this crater structure to crypto-explosive activity.

Orti et al. (2008) defined the El-Baz structure as a crater-like feature delimited by basalt dikes intruded into quartz arenitic bedrock. Consequently, no evidence for an impact origin of this structure has been related to date. The observation by Orti et al. (2008) of volcanic activity around this site leads us to believe that no tangible support for an impact origin is on the table – and thus the structure is listed amongst the discredited impact proposals.

6.5.8. Gilf Kebir crater field, Egypt

In 2004, Paillou et al. proposed that at least 13 of the numerous elliptical to roughly circular structures visible on satellite imagery over the region of the Gilf Kebir plateau in the western Egyptian Desert could represent impact structures. The area of concern is located between latitudes 23°14′–23°32′N and longitudes 23°17′–27°27′. As with the Arkenu discoveries (above) by the same group of authors, shock deformation in the form of planar deformation features in breccias and shatter cone-like features were cited as positive evidence for impact. Similar to the Arkenu story, the evidence was not satisfactory.

Two years later, Paillou et al. (2006) counted not less than 1300 crater-like structures in this region and extended the number of possible impact craters to a minimum of 62. This time, they did, however, discuss the possibility that these structures might not be of impact origin but rather represented hydrothermal vents.

In 2005, an Italian expedition (Orti et al., 2008) visited this region and carried out detailed field work, including structural geological analysis and geophysical studies, thereafter followed by petrographic and geochemical investigations. These authors also presented a detailed geological account of the geology of the region surrounding the Gilf Kebir plateau. This 8000 km^2 plateau, which is elevated by some 700 m above the general desert level, comprises a cover of Quaternary sand dunes and sheets overlying Cretaceous to Lower Tertiary sediments of the Nubia Formation. In southern Egypt, these sandstone-dominated strata are interbedded with lavas and tuffs, and there are alkaline intrusive bodies related to Cretaceous and Lower Oligocene magmatic activity (Issawi, 1982; Klitzsch et al., 1987) as well. South of the Gilf Kebir trachyte and phonolite plugs and cones testify to late Paleozoic (Hercynian) volcanic activity. Such bodies have also been observed in the wider region to the south, whereby many such occurrences have sandstone rims around basalt bodies, lending the complexes a caldera-like appearance.

Orti et al. (2008) found that of the 62 structures singled out by Paillou et al. (2006) ten structures could be associated with basalt dikes. Volcanics observed are mostly of trachyte and olivine basalt composition, but phonolites, rhyolites and microsyenites also occur. Breccias observed were identified by Orti et al. as intraformational sedimentary breccias. Another breccia type was noted in rims of some of these structures and was tentatively interpreted as the result of tectonism. No characteristic shock deformation could be found in samples of both these breccia types. Observed microdeformation features are regarded the product of normal tectonic deformation. Neither did they observe bona fide shatter cones. Instead, striations on local rock surfaces are revealed as superficial features that are not related to fracturing. As these features are also observed away from crater-like structures and seemingly related to main wind directions, they are likely the result of wind erosion. Geophysical surveys by the authors at some of these circular features showed that they do not possess the bowl-shape geometries one would expect for small impact structures. Consequently, Orti et al. (2008) arrived at the conclusion that "the circular structures detected in the Gilf Kebir area are related to endogenic geological processes typical of volcanic areas, such as extensive geothermal fields that have been activated after the deposition of a sedimentary cover".

6.5.9. Kebira, Libya

El-Baz and Ghoneim (2007) followed up their internet fanfare of 2006, with which they announced the discovery of a very large impact structure in the eastern desert at the border between Libya and Egypt. This claim of the presence of a 31 km diameter impact structure at 24.40N/24.58E (Fig. 57) and announcement of this feature being the "largest impact crater identified in the Sahara" was based on Landsat ETM+ image interpretation and Radarsat-1 data, which indicated to the authors a discontinuous outer rim, and a "group of prominences forming an inner ring." They reported that the host Nubian sandstone had undergone severe erosion, and that the crater morphology had been severely affected by both eolian and fluvial erosion. SRTM data indicated to them that drainage including a major river was responsible for the degradation of the structure. Based on these findings only, without any supporting field or petrographic evidence, the claim of a "Kebira" impact crater was launched. Lacking any modesty, the authors also chose the name *kebira*, which in Arabic means *large*. Finally, a further quite unsupported add-on was the suggestion that this crater structure was "possibly the source of the silica glass fragments that abound on the desert surface…in southwestern Egypt" (El-Baz and Ghoneim, 2007).

A host of rejections of the Kebira impact claim has since appeared on the internet, including some imagery of flat-topped sedimentary plateaus in the center of the structure. However, no proper geological investigation has been published to date, so that the inclusion of this structure in this list of discredited proposals of impact structures must be judged carefully. It is our personal belief that there is no tangible evidence on the table for the existence of

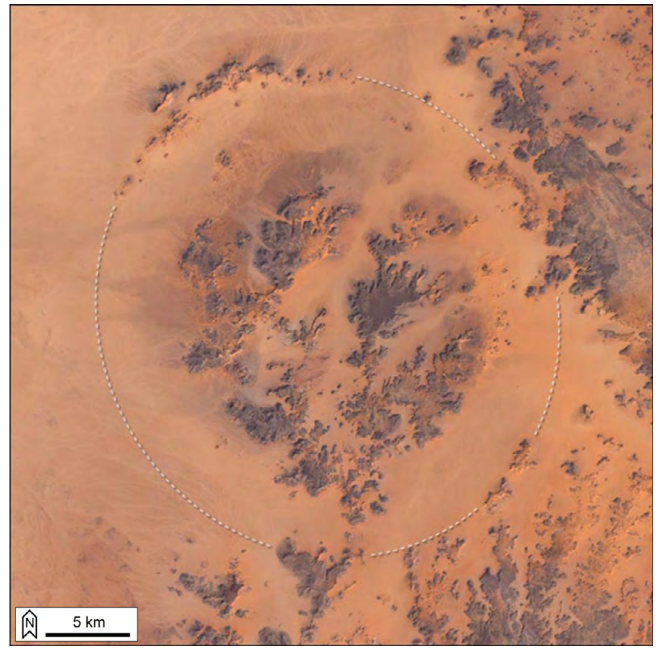


Fig. 57. Satellite image of the so-called "Kebira crater" (NASA Landsat image). The supposed crater outline according to El-Baz and Ghoneim (2007) is indicated by the white dashed line. No crater is actually visible and no indication of an impact origin has been found.

an impact structure at Kebira, and unless this is presented in the future, we regard this claim of such an impact structure as unfounded and unsubstantiated.

6.5.10. Lac Télé, Republic of Congo

Garvin (1986) drew attention to Lac Télé, a ca. 6 × 8 km sized, conspicuously oval-shaped lake located at 1°20′N/17°10′E in the tropical rain forest region of the Republic of Congo, based on a remote sensing study. Garvin noted that the lake's drainage pattern is strongly radial, as had been observed for several impact structures as well. Lac Télé is located west of the town of Epena between two branches of the Likouala river. Garvin could not report any information that could have elucidated the origin of the lake, and referred to an eroded impact, a buried and collapsed volcanic caldera, or basement subsidence as possible options for the formation of this lake.

Master (2010b) provided a comprehensive analysis of Lac Télé, in its geological context. He emphasized the importance of this area as a biodiversity and conservation hotspot, which includes the conservation of a number of endangered mammal species such as gorillas and chimpanzees. Master reviewed the outcomes of a number of expeditions to the lake that had been conducted in recent years, focusing mainly on biodiversity issues, but the results of which also had implications for the hypothetical impact origin of the lake. According to Master, Lac Télé is located in the northeastern part of the intracratonic Congo Basin, in a region that is dominated by Holocene alluvium. The lake is located within the basin of the Likouala aux Herbes river, which is a multi-branched meandering river flowing over very low gradients. The river basin is characterized by rain forest and swamps. Previous bathymetric studies demonstrated that the lake has an average depth of only 4 m, is rich in organic-rich silty sediments, and that the lake bottom geometry resembles that of a flat-bottomed dish. The author reminds that such isolated lake ecosystems are not unique in the Congo Basin. Rather, there are several similar small, isolated, and shallow lakes surrounded by rain forest and marshes, for some of which an origin due to damming of drainage by neotectonic faults had been suggested. And such an origin, due to the lake's location over neotectonically reactivated, seismically active regional lineaments, is also assumed by Master (2010b) for Lac Télé. Particularly, the inconsistence of the morphology of the lake with that of an impact structure of this size (complex crater geometry with a comparatively deep interior basin surrounding a central uplift, an overturned rim, ejecta blanket) were cited by Master as argument against the possibility that Lac Télé could represent an impact structure.

6.5.11. Lukanga Swamp, Zambia

A 52 km impact structure was postulated based on the occurrence of some breccias and of an aeromagnetic anomaly in the Lukanga Swamp area (around 14°24′S/27°54′E) of central Zambia by Vrána (1985). He alleged that quartz in these breccias showed planar deformation features of shock origin. The swamp is located in central Zambia, about 100 km to the west of Kabwe town. It covers a grossly rhomb-shaped area and is generally surrounded by flat terrain. Along the southern margin of the swamp extends the Nyama Dislocation Zone (a fault zone that below is referred as the NDZ). The swamp is set in Proterozoic basement rocks, and this suggested to Vrána that the impact structure was deeply eroded and old. It was only in 2000 that an effort was made to follow up on this proposal with some field work, which lead to Katongo et al. (2002) giving a comprehensive account of the geology of this area and report on the results of the expedition there, and of subsequent petrographic and geochemical investigations.

During the field work at Lukanga, 96 samples were collected at scattered outcrops in the Nyama Dislocation Zone and at several other sites around the swamp. In particular, samples were obtained at locations previously mentioned by Vrána (1985). Outcrops along the NDZ form a continuous trend of boulders that form loose ridges or occur in isolation. Mapping and sampling involved sheared quartzite, sedimentary quartz breccias, metapelite (meta-siltstone and shale), and fault breccia. Outside the NDZ, muscovite schists, muscovite-bearing quartzite, granites and gneisses were mapped. Katongo et al. (2002) discussed that structures and textures in the meta-sediments are of sedimentary origin, and the fault breccias display tectonic deformation fabrics. Quartz in sedimentary breccias is characterized by widely spaced, randomly oriented non-planar but occasionally subparallel fluid inclusion trails, which are likely akin to those features previously, and erroneously, misidentified as shock-induced planar deformation features. In the absence of shock deformation, any geochemical signatures that might indicate involvement of a meteoritic component, and of aeromagnetic signature that could be suggestive of the presence of a large impact structure, Katongo et al. (2002) concluded that the suggestion by Vrána could not be substantiated. They rather favored from their synthesis of regional structural information that reactivation of movements along the Nyama and the Kapiri-Mposhi Dislocation Zones might have lead to the development of the Lukanga Swamp – maybe as late as during the Cenozoic.

6.5.12. Mazoula, Algeria

An 800 m diameter, multi-ring feature, with a 300 m wide anticlinal dome that rises some 30–35 m above the surrounding, horizontally disposed strata, is located at Mazoula, at 28°24′N/7°49′E in Algeria. Lambert et al. (1981) investigated the structure. They report that the area is capped by a flat-lying massive carbonate layer with outward dips along the flanks. There were no particular rock disturbances, no fractures and no evidence of breccias, which lead them to conclude that Mazoula is not of impact origin, but represents a Cretaceous rudistic reef and/or salt diapir.

6.5.13. "Meteor Strike", Kwa-Zulu-Natal, South Africa

Brandt et al. (1999, 2001) investigated rumors surfacing from the mid-1990s amongst the geological community of South Africa of the possible existence of two meteorite impact sites in northeastern KwaZulu-Natal Province of South Africa. The area of concern is the western shoreland of Kosi Bay lagoon. A site marked "Meteor Strike" was even found on the Kosi Bay topographic map sheet at 26°56.6S/32°47.0E, directly adjacent to a mission station with the poignant name "Star of the Sea". Brandt et al. in 1999 undertook a field visit to the area and made the following findings:

A ca. 65 × 170 m, elongated, north–south-trending, bowl-shaped depression was found about 500 m north of the mission, within the area of the highest local dune. No deformation that could be related to an impact event was observed in the horizontally oriented eolinites, where surface sands had been removed by wind and human activity. No anomalous rock types were in evidence, and a ground-magnetic survey of the depression did not indicate any presence of magnetic material such as certain types of meteorites. Extensive interviews with local people, including several tribal chiefs (indunas), did not yield any reference to possible meteorite impacts in the time of human memory. The name "Meteor Strike" apparently only surfaced in the late 1940s to 1950s, the time when the local mission was renamed from St. Francis to Star of the Sea. Induna (meaning *chief*, in the Zulu language) N. Tembe was credited with the saying: "no local mentioned a meteor crater until the white man came along…and white men know so much more about meteor craters…". Brandt et al. concluded that the reference to a meteorite impact in this area is the result of a mere coincidence of the local occurrence of sizable wind blow-out holes due to the persistent north–south direction of the wind in the region and the presence of the mission station of a somewhat leading – or misleading – name.

A second locality 8 km further north, at 26°53′S and 32°51.2E, on the flank of a north–south trending dune, was also investigated by the team. The 50–60 m wide depression was partially degraded by human recreational activity, and originally it could have been circular. Again, a magnetic survey did not yield any evidence supporting an impact origin. This part of the coastal area around Kosi Bay contains a large number of blow-outs on many dunes. Brandt et al. concluded that there was no evidence for impact here either. The conspicuous holes are likely the result of eolian process, perhaps supported by local seismic activity. Human-inspired change of the surface geology can also not be excluded.

6.5.14. Nyika Plateau structure, Malawi

Master and Duane (1998) investigated hearsay reports (Mossmann, 1972) of a small impact event that had formed an 80 m impact crater in 1959 on the Nyika Plateau (around 09°30′S/32°15′E) near the border of Malawi and Zambia. Sketchy notes of a short visit to the area by Mossman and secondhand oral information (bright light, loud explosion) lead the authors to investigate the area of the Chilinda Pine Plantation in the Nyika National Park of northern Malawi. They detected that the site of the alleged explosion coincided with the location of the Chilinda mudslide of 23 April 1960. Master and Duane found detailed documentation and obtained oral history about the mudslide, which seems to have been an example of a mudslide due to waterlogging of clayey soils along an unstable slope. Their findings and the reports of plantation workers that remembered the event of 1960 apparently exclude the possibility that a meteorite strike might have occurred.

6.5.15. Richat, Mauritania

The Richat dome was first described by Richard-Molard (1948b, 1952) and by Monod (1952, 1954). On satellite imagery this enigmatic structure appears as a giant "bull's eye" feature (Fig. 58). It is located in the Adrar region of central Mauretania, centered at

Fig. 58. The about 50 km wide Richat structure of western Mauritania. For details see text. Space Shuttle photograph, mission STS-41G of October 1984. Image No. 17-33-110. Courtesy Lunar and Planetary Institute, Houston.

21°04′N/11°22′W. The Richat structure has long been proposed as being of impact origin (Cailleux et al., 1964; Freeburg, 1966). Evidence cited in favor of this origin includes highly circular morphology and rarity of this morphology for very large structures of endogenic origin, indication of stratigraphic uplift in a region of flat-lying sedimentary strata, an inward directed drainage pattern, extensive outcrops of brecciated chert and quartzite, and a report by Cailleux et al. (1964) of coesite occurring in a brecciated quartzite.

Dietz et al. (1969) reported observations made on the ground. They described Richat as a circular topographic depression of some 38 km diameter. In later literature it has been measured variably at 45 km (Netto et al., 1992) or as "at least 40 km" in diameter (Matton et al., 2005), and on recent webpages values up to 50 km diameter have been given. Measuring from the outermost clearly visible ring feature on satellite images, we obtained a value of 44 km for the width. The structure occurs at the edge of an extensive plateau formed by flat-lying sedimentary rocks capped by Chinguetti sandstone of lower Paleozoic age. The rocks exposed in the interior of the structure represent the lower part of the succession and comprise carbonates, cherts, sandstones, quartzites, and siliceous shales. As a result of differential erosion of rocks of varied resistance, the dome consists of a series of annular rings. Dips of the beds increase towards the center, but the most central "eye" is occupied by flat-lying limestone and meta-arkose. Dietz et al. (1969) reported evidence of volcanism in the form of small dolerite dikes or sills at several sites within the structure. They also refer to zeolitized volcanic tuff from the environs of Richat and determined a strike orientation of 30°NE for a well-developed fault system that trends in the direction of *Tenoumer* impact crater and the *Temimichat* structure of still unconfirmed origin.

Dietz et al. could not report any direct evidence of impact. Shatter cones were not detected, megabreccia is missing at Richat, and overturned or at least upturned strata are missing as well. No injected breccias, or something resembling pseudotachylitic breccias, could be observed either. And microscopic analysis of several specimens, including some from the locality sampled by Cailleux et al. (1964) as well as specimens from the central breccias, failed to reveal shock metamorphic evidence. In agreement with Richard-Molard and Monod, who had advanced an endogenic origin for Richat, the latter referring to the presumed presence of a laccolith at depth, Dietz et al. concluded that Richat was not of impact origin. Monod and Pomerol (1973) subscribed to an origin by either explosive volcanism or emplacement of a pluton at shallow depth.

This left the alleged finding of coesite in a shattered sandstone reported by Cailleux et al. (1964). Fudali (1969) claimed that the coesite recognized by Cailleux et al. had been a misidentification, as he had found barite in the same breccia. It is not clear, however, whether Fudali re-analysed the same sample that Cailleux et al. had worked on, or whether he only re-sampled the same locality. Master and Karfunkel (2001) argued that even if coesite were genuinely present at Richat, this would not necessarily argue for an impact origin, as coesite could also be the result of lightning strike. As the Sahara was a much wetter environment in the past, lightning strike-related formation of coesite was a real possibility.

Netto et al. (1992) referred to the location of Richat as being at the intersection of two sets of fractures trending in 10–20°NE and 79–80°E directions. Three kinds of igneous rocks had been detected at Richat: a half ring-dike (or sill?) of gabbro in the western part of the outer depression, and two thinner dolerite sills in the eastern part of the central depression; some analcimolite (already referred by Dietz et al., 1969) in the central depression; and calcitic and magnesian carbonatite veins occurring in the central depression and abundantly in the southwest part of the outer depression (for the latter, see also Woolley et al., 1984). Netto et al. obtained fission-track data on 8 apatites from a carbonatite vein that yielded an apparent age from an isochron of 77 ± 2 Ma age, which was corrected to an age of 85 ± 5 Ma for the beginning of track retention (cooling below 130 ± 10 °C). A lower age of 49 ± 10 Ma corresponded to the last cooling below 75 ± 15 °C. Circumstantial evidence of "quasi-explosive" origin for the carbonatite veins was taken by these authors to suggest a common origin for the Richat doming event and carbonatite emplacement, so that they concluded that the corrected age of ca. 85 Ma dated the Richat formation event itself. Poupeau et al. (1996) obtained a 99 ± 5 Ma age for apatite from a carbonatite.

In 2005, Matton et al. advanced a model for the formation of the Richat structure based on a comprehensive re-investigation of the dome (see also Matton, 2008). They concluded the following: The at least 40 km diameter structure results from differential erosion that has created circular cuestas around a central limestone-dolomite shelf that encloses a kilometer-scale occurrence of siliceous breccia. Basaltic ring dikes, kimberlitic intrusions, and alkaline volcanic rocks have intruded the structure. The authors find that the breccia body is of lenticular shape and thins out towards its margins. The breccia is considered the result of karst dissolution and collapse, with sediments filling the resultant cavities. These authors provide evidence for a relation between the hydrothermal infill and magmatism. Doming and hydrothermal fluidization created favorable conditions for dissolution. Argon dating results by Matton et al. (2005) are in excellent agreement with the earlier fission-track dating results and also indicate a Cretaceous age (98.2 ± 2.6 Ma) for the emplacement of carbonatites – and by implication for the formation of the Richat dome.

No evidence remains that could be linked to an origin by impact of the Richat structure.

6.5.16. Semsiyat, Mauritania

A second domal structure, located some 50 km west-southwest of Richat's center on the Chinguetti Plateau at 21°00′N/11°50′W in Mauritania, was referred by Dietz et al. (1969). Its diameter is about 5 km. These authors observed that its tectonic style was the same as that at Richat, so that they regarded Semsiyat the "petit frère" of the larger structure. On satellite (e.g., GoogleEarth) or space shuttle imagery, Semsiyat appears as a three-ring feature, with the innermost structure appearing as a depression of irregular outline. A bull's-eye pattern such as Richat, with multiple, well-defined, and narrow rings, is not exhibited. While already Dietz et al. (1969) noted that Semsiyat could be clearly observed from the air, it apparently does not have a strong ground exposure. The dome was described by them as having a relief of just a few meters occurring on a flat plateau. The innermost circular area and an outer annulus were characterized as shallow depressions. Dietz et al. did not observe any evidence suggesting impact, such as shatter cones, but collected some unshocked chert breccia in the innermost area. In the absence of any evidence for exogenic action, they concluded on the basis of similarity to Richat that Semsiyat was also of endogenous origin.

6.5.17. Shakiso, Ethiopia

Evdokimov (1987) and Evdokimov and Abebe (1987) reported on the presence of a possible impact structure at Shakiso (centered at 05°46′N and 38°54′E) in south-central Ethiopia, about 500 km from Addis Ababa. Shakiso is situated in the fairly well-studied Adola gold field. Rocks outcropping in the area belong to the about 630–680 Ma old Middle Complex in the classification of basement rocks of Ethiopia. These rocks include N–S trending biotite gneiss and muscovite-kyanite schists forming the gneissic terrain that surrounds the Adola greenstone belt. Evdokimov (1987) reported on the existence of "impactites and explosion breccias" as well as Ni–Fe pellets. Geological data supposedly indicated a structure with about 2.5 km diameter centered directly on the town of Shakiso, extending about half way to the Hawata river. He cited an age of about 10 Ma, but without giving any analytical data to support this. Evdokimov and Abebe (1987) revised the diameter to 6 km based on the occurrence of "shatter cones" and gave an age of 65 Ma.

However, detailed field work, petrographic, and geochemical studies on rocks from the area did not show any evidence of an impact structure at that locality (Abate and Koeberl, 1998).

6.5.18. Zimbabwean structures

In September 1993, a group of impact researchers from the University of the Witwatersrand (Johannesburg) and University of Vienna (Austria) toured parts of Zimbabwe to investigate several targets identified on morphological or geophysical grounds (Reimold, 1994; Reimold et al., 1994b). The targets included the Thuli structure on the geological block of the same name in the southern part of the country, west of Beitbridge town, the twin Save structures in the southeast close to the Mozambican border, and the Chegutu geophysical anomaly southeast of the city of Mucheka. A first exploratory visit was paid to Highbury as well, results of which are reported above (section 6.4.11).

Thuli, which had been identified on aerial photographs as a 1100 m diameter, near-circular, crater-like structure, which was also marked on the 1:100,000 Thuli geological map of the Zimbabwe Geological Survey, is located in a rather inaccessible part of the tribal areas some 100 km west of Beitbridge, along the Sentinel Road. Thuli location is situated at 21°55′S/29°12′E. Reimold (1994) suggested that the maze of dirt roads in the Manimani Tribal Area could have been the reason that Thuli had never been investigated by geologists prior to this visit. Thuli crater offers, upon approach, a panorama very similar to that of *Tswaing* impact crater in South Africa. Field work unfortunately quickly convinced the visitors that Thuli crater was of volcanic origin, due to the abundance of basalt, gabbro, and diorite in evidence. The crater-like form was thought to be the result of differential weathering of these different igneous products.

The *East and West Save* twin crater structures are reached via some 100 km of dirt road commencing at the Masvingo-Birchenough Bridge road to Masapas Ranch in the immediate vicinity of the Save structures. The Save East and West crater-like features measure 600 and 800 m in width. For Save West a GPS location at 32°17′25″E and 20°03′25″S was measured (Master and Robertson, 2009). Especially the smaller structure has a pronounced crater shape. On aerial and satellite imagery, these structures closely resemble small impact craters such as Tswaing. The result of field work was that basalt and gabbro occur throughout the structures, the rims of which are formed from indurated, weathering resistant sandstone. Prior to our 1994 visit to the structure, the twin Save structures had been referred as volcanic pipes, including possible locations of kimberlites.

A ground-magnetic investigation of the smaller twin did not yield any further evidence in support of an impact origin; instead the N–S traverse measured was judged (also by Master and Robertson, 2009) to indicate that these structures could possibly represent cylindrical, vertical volcanic bodies that might be at least several hundred meters deep. Volcanism is thought to have taken place during Karoo Supergroup times.

The strongest negative aeromagnetic anomaly with a gradient of −800 gamma between rim and center in Zimbabwe is centered at 18°08′13″S/30°08′09″, the location of the train station at *Chegutu* town in central Zimbabwe. Archean basement, including some pegmatites and breccias, were observed there. The initial hope that the latter could be of possible impact origin was, however, also quenched quickly when the breccias were recognized as volcanic agglomerates. The short visit did not have the potential to solve the origin of this magnetic anomaly, but it remains an enigmatic geophysical feature and interesting exploration target.

7. Outlook

This review has highlighted areas where the investigation of African impact structures has made important inroads and contributed to the general understanding of impact cratering. This includes:

1. Since the review by Koeberl in 1994, when only 14 impact structures had been recognized on the African continent, the African impact record has been increased to 19 confirmed structures plus 1 shatter cone occurrence – which is not a major increase over a 19 year period. To date, not less than 49 sites with possible impact structures have been proposed, only a few of which are actually under direct investigation. However, the morphologies of some of these structures are promising that they eventually might turn out to be *bona fide* impact structures. A further 29 alleged impact structures have been shown to date not to be of impact origin.
2. This state of affairs demonstrates that the days when it was relatively easy to identify new impact structures on Earth are over. The easy targets have been picked and entered into the impact record. What remains to be done is a lot of groundwork – ascecerbated by the readily available remote sensing data sets for the continent. However, the current record (Tables 2a–c) also illustrates vividly that the interest in impact as a fundamental natural process has grown enormously since 1994; the number of researchers responsible for the three parts of the record (confirmed, possible, discarded) speaks for itself. Fig. 16 shows that the structures investigated occur throughout a major part of the African continent. And yet, there are still gaps, particularly in Central and East Africa. Maybe this discussion of the African impact record will stimulate some workers in those countries where no structures of interest are known yet to reconsider the geology of their country.
3. Furthermore, there are still large regions on the African continent that have not been assessed for possible impact heritage – be it for inaccessibility due to rainforest terrain or due to civil unrest and war.
4. On the other hand, those structures that have been confirmed have yielded sometimes amazing information that has contributed extensively towards general knowledge about impact

structure geology and impact cratering processes. A prime example is the fascinating geology of the Vredefort Dome that has provided an extraordinary perspective on the deep parts of the central uplift of a very large impact structure. The relationship between Vredefort Dome and surrounding Witwatersrand Basin has provided a new perspective on the evolution of those unique Witwatersrand gold deposits. In addition to impact cratering-specific gains of knowledge, detailed investigation of some African impact structures has benefitted greatly other geological or socio-economic goals: Again, Vredefort can be cited with regard to the importance of understanding the impact structure in the context of the economic geological importance of the Witwatersrand Basin within its confines. The 3D analysis of the Bosumtwi impact structure has also highlighted the potential of impact structures as reservoirs for paleoclimatic records, besides the need to preserve the food resources of the lake for the local inhabitants.

5. The long duration of geological exploration of the Vredefort Dome that only after some 90 years culminated in the confirmation of the origin of Vredefort by impact has taught many lessons about impact and shock metamorphism, has stimulated much research on shatter cones, the structural geology of impact structures and metamorphic and shock metamorphic patterns of the interior of large central uplift structures, has repeatedly given new impulses to the study of pseudotachylitic breccias and impact melt rock, and it has been a constant source of controversy. Clearly, this is an excellent example how geoscience works – by leaps and bounds, and revealing some detail but causing new confusion (one could also say: generating new questions...) and debate. Those who explored Vredefort would say that they would not want to miss this intriguing aspect of geoscientific detection. And still, after all the work accomplished, there is still a lot more to be learned from the intriguing deep-crustal profile provided by the Vredefort Dome and its overprint of shock and thermal metamorphism.
6. The spherule layers of the Barberton Mountain Land and the Transvaal Supergroup of South Africa have yielded much stimulating knowledge about distal impact ejecta, inherent meteoritic components, and modification of the original meteoritic signatures by later hydrothermal overprint. The fact that several of these spherule beds can be correlated with equivalents on the Australian continent emphasizes the global importance of such marker beds. The role of early impact in Earth's history is exemplified by these layers, and the search for more ancient witnesses of impact in the Archean record will be continued as such ejecta layers will keep forthcoming in the younger stratigraphy.
7. Tswaing and Bosumtwi are prime examples for the importance of crater-sediment records for paleoenvironmental/climatic reconstruction. Bosumtwi has the potential to reveal a 1 Ma equatorial record, and Tswaing has already provided much excellent paleo-environmental data for the Holocene of the southern mid-latitudes.
8. Several African impact structures are already the targets for geoconservation and geotourism efforts (Tswaing, Vredefort – examples of Fig. 59). Roter Kamm has been protected within the Namib-Naukluft Reserve. The potential of Bosumtwi as a geotouristic feature with the need for conservation has also been debated. Clearly outstanding geological sites have touristic value, in general, and impact structures are rare and in many parts of the world extraordinary sites. African geosocieties and legislative bodies ought to consider this possible resource more in future.
9. In many instances, we have emphasized in the above account where further research work is required – and possibly most promising. We have also attempted to suggest where different disciplines could make contributions, including geophysics, structural analysis, mineralogy, or geochronology. Africa's impact crater record has grown over the past two decades, but the large number of still unconfirmed structures emphasizes the need for further analytical efforts – both in the field and in the laboratory. That the number of unconfirmed or disproven structures is growing strongly signifies that the times of easy recovery of an impact structure is past, and that the much improved accessibility of remote sensing data and initial suggestions of new "impact structures" far outpace the rate of confirmation by ground-truthing. We have noted examples of sinkholes, circular anthropogenic features, or volcanic structures that have been investigated based on initial targeting as impact structures.

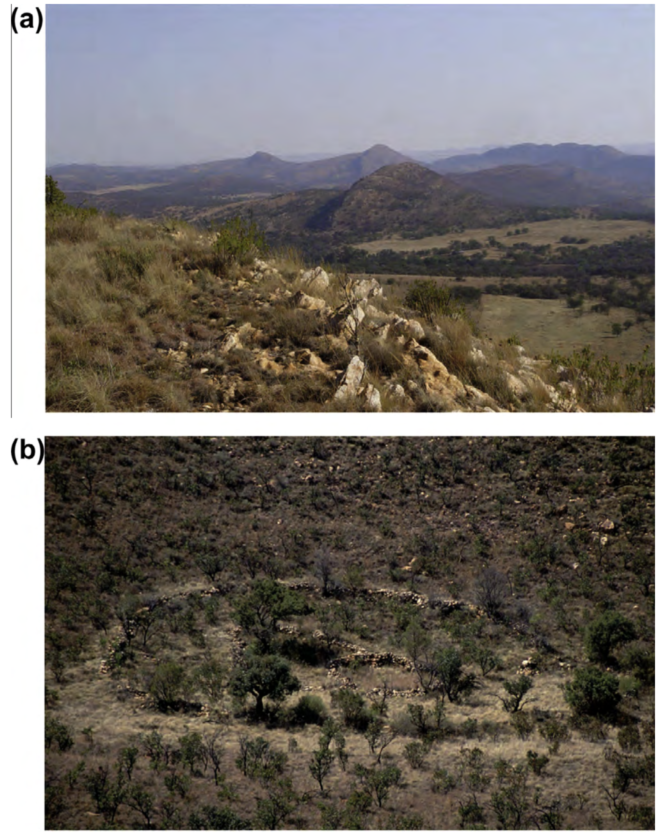


Fig. 59. Two examples of geotouristic aspects of the Vredefort Dome, South Africa. (a) Scenic view from Steenkampsberg, western collar, towards the north across the Vredefort Mountain Land, the rugged landscape of impact-deformed strata that forms the heart of the Vredefort World Heritage Area. (b) Besides the 3.5 Ga of geological evolution of the Kaapvaal Craton and the exceptional deformation phenomena related to the mega-impact event, the Vredefort Dome also offers historical and archeological phenomena of interest: here, the remnants of an iron-age settlement on Thwane, western collar of the Vredefort Dome. For more information on Vredefort geology and geotouristic attractions, refer to Reimold and Gibson (2010).

And yet, significant discoveries remain to be made. Large countries such as Sudan and South Sudan, Central African Republic, Angola, Mozambique, and the countries on the Horn of Africa, do not have any impact crater record yet. The easy access to GoogleEarth and other satellite data based software allows students and other individuals to carry out first-order high-resolution searches for circular features. Baratoux et al. (2012) reported on a student class project, whereby the surface of Morocco was apportioned into areas small enough to be manageable by individual students. Their detailed search revealed not less than 30 circular (or near-circular) prospects, which are now followed up with detailed research on local geology and geophysics. The best candidates after this second phase will have to be investigated on the ground. Even if this project remains unsuccessful in terms of confirmation of impact craters, it has enormous intrinsic educational benefit: it combines remote sensing with multidisciplinary geoscientific research, and in the process acquaints the students with one of the fundamental planetary processes – impact cratering. Tuition about impact cratering also places Earth into the planetary context of the wider solar system. At the same time, students are learning to carefully interpret local remote sensing data with regard to many aspects: geography including land use, environmental science, geology, geophysics, local vs. regional geology, etc.

Africa still poses significant challenges to ground-truthing efforts. Unstable political conditions, especially in central African and Saharan countries, pose a severe difficulty for any field effort. On the other hand, still today the geological tuition at many African universities does not incorporate teaching about the basics of impact cratering – despite the strong economic importance of many impact structures and their ore resources. Lack of financing for even basic teaching materials is obviously counterproductive as well – and high-resolution remote sensing data which are absolutely mandatory in support of basic fieldwork in remote regions are still extremely expensive and not readily available to teaching departments and geological surveys. On the other hand, there are many countries in Africa where detailed geophysical surveys have been performed as part of regional exploration enterprises – by geological surveys and mining companies, and even with the support of the World Bank. These data sets might provide many more hints at the possible existence of impact structures – and we can only hope that this review will stimulate some geologists on this continent and abroad who to date have not been conscientious of the possibility of impact structure discoveries. We also hope that this compilation of information will provide a widely used entry into the literature and, thus, represent a teaching aid as well as support for further in-depth research. Already proven impact structures may have great potential as field-based multidisciplinary laboratories, where a wide range of geological and geophysical subjects (stratigraphy, structure, sedimentology, metamorphism, hydrothermal overprint, magmatic processes, geoconservation, hydrology, training in geophysical field methods, etc.) can be taught.

A major problem regarding the African – but also the global – impact cratering record concerns the concomitant age record (Jourdan et al., 2009, 2012). This involves both the direct dating of impact lithologies and chronostratigraphic analysis. Here, too, African geologists can make a strong contribution through their field efforts, and in collaboration with laboratory researchers. This review may provide information about which groups are currently focusing on which aspect of impact cratering research. We hope that networking with them will be supported by this review effort as well.

From a global perspective, the need to further improve our understanding of the impact record, and the catastrophic as well as beneficial aspects of impact cratering, remains high. This is not only a matter of academic interest – that is to gain further planetary insight from the study of the impact records of the terrestrial planets and their satellites, but it is also a matter of survival. The impact threat to Earth's inhabitants is real (Durda, 2013). NASA's Near-Earth Object Program has been focusing in the past two decades on identifying 90% of Earth-threatening NEOs larger than 1 km, with significant success. Such bodies would impact with a considerable lethality and at least regionally disrupt our involved civilization's skeleton of human and communication networks. With this goal now essentially achieved, the focus is on the estimated 100,000 NEOs larger than 140 m. Such a survey will also detect a number of smaller possible impactors, but considering the total number of objects in this size category being some 80 million – one must assume that a Chelyabinsk-sized incident will occur again....perhaps soon.

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