Chapter 1

Introduction

1.1 Purpose of this study

The Vredefort impact structure in South Africa is currently the oldest known impact structure on Earth (e.g., Kamo et al. 1996; Gibson and Reimold 2001) and the largest one (e.g., Henkel and Reimold 1998). In contrast to other large impact structures in the world, it is only in the Vredefort impact structure in South Africa that deep levels of erosion, combined with limited post-impact tectonic disturbance, allow the study of the largely undisturbed (largely undisturbed with regard to post-impact tectonism) roots of a giant, complex impact structure, and investigation of the formation and modification mechanisms of large central uplift structures. In addition, the Vredefort Dome is also a window into the deep structure and composition of the Archaean Kaapvaal Craton, providing information on more than 3 Ga of tectonic evolution and metamorphism (Lana et al. 2003b, 2004). Numerous studies have been done on the Vredefort Structure (see bibliography by Reimold and Coney 2001), but the collar rocks of the Vredefort Dome have remained largely unstudied from a structural point of view. Furthermore, most of the earlier studies of the collar did not recognize the impact origin of the Vredefort Dome (e.g., Nel 1927a-c; Bisschoff 1969; Simpson 1977, 1978; Lilly 1978).

To provide a basis for the discussion of the processes involved during the development of a complex impact structure, such as the Vredefort Structure, a review of studies of impact structures (section 1.2) and how a possible impact crater can be identified (section 1.3) are presented. This is followed by an introduction of the different types of impact structures (section 1.4) and the formation of complex craters (section 1.5). In order to analyse the deformation features in the collar rocks of the Vredefort Dome, the remaining problems regarding the Vredefort Dome are pointed out (section 1.6). To establish the origin of the structures in the collar rocks in a regional context, a review of the geological and tectonic history of the region of the Witwatersrand Basin is necessary (section 1.7). This is followed by the formation of the Vredefort Dome (section 1.8), before its geological and tectonic history are addressed (section 1.9). This includes the presentation of different models for the
evolution of the Witwatersrand basin and controversies regarding the origin of the Vredefort Dome.

Despite intensive research on impact structures around the world (e.g., Melosh 1989; French 1998, and references therein), many questions remain unanswered about the nature and exact timing of individual processes involved in the formation and modification of impact structures. In particular, the processes involved during the rise of the central parts in very large impact structures and their subsequent collapse remain enigmatic. Therefore, in order to understand the process of uplift and to possibly distinguish different structural features with regard to the timing of various stages in the development of an impact structure, a detailed structural analysis of the Vredefort impact structure was carried out, of a part of the innermost collar, with some excursions into the surrounding Witwatersrand Basin. This project was aimed at documenting all structures observed within the study area in an attempt to establish geometric and temporal relationships between these structures. This structural analysis and the possible relationships of structures may help to establish the origin of and kinematics related to individual deformation features (pre-, syn- or post-impact).

The objectives of this project can be expressed by the following critical questions:

- How does a central uplift form in such a large impact structure and what kind of deformation features (brittle – ductile, large-scale – small-scale) are involved?
- What kind of kinematic movements can be expected during initial central uplift formation at such deep levels?
- What type of deformation is associated with the collapse of the central uplift?
- Which movement patterns can be found in the field in the collar rocks and further outwards?
- Can impact-related structures be distinguished from non-impact structures?

The results of the structural analysis are presented in Chapters 2 to 5. The structures observed in the collar rocks of the Vredefort Dome are presented in Chapter 2. Chapter 3 deals with the shatter cone deformation phenomenon and Chapter 4 with
the pseudotachylitic breccias in the collar rocks. The results of the surface mapping in the wider environs of the Vredefort Dome are addressed in Chapter 5. Finally, the results are discussed and compared to observations from other impact sites in Chapter 6.

1.2 Research history of Impact Cratering

The possibility that large extraterrestrial bodies impact violently on the Earth’s surface was rejected for centuries (e.g., Schröter 1791; Herschel 1787, cited in Melosh 1989). Only in the 20th century did geologists begin to recognize impact of large bodies from space, especially during the 1960s and 1970s, when mankind started to explore the Earth’s nearest planetary neighbour – the Moon – and to investigate our Solar System with space missions and space probes. The craters on the Moon had been known for centuries but were long considered of volcanic origin (see review in Melosh 1989 and references therein). However, the closer study of the Moon (its surface and samples) found definite evidence that most of the lunar craters are of impact origin (e.g., Baldwin 1949; Wilhelms et al. 1987; Taylor 1992). Solar System wide exploration, for example by the Pioneer and Voyager space probes, resulted in the remote sensing observations of impact structures on other planets in our Solar System (e.g., Mercury, Venus, Mars). Consequently, and supported by these space missions, scientists concluded that meteorite impacts played – and still play – a significant role in the formation and evolution of all planets by accretion of small bodies (planetesimals) and by shaping the planetary surfaces. This was brought to the attention of people when in July 1994 fragments of a comet impacted into the atmosphere of Jupiter.

Contemporaneously with the acceptance of meteorite impacts on other planets in our Solar System, scientists began to search for evidence of impacts of extraterrestrial bodies on Earth (e.g., Dietz 1947, 1959; Shoemaker et al. 1961). Scientists realized that meteorite impacts have significantly influenced the evolution of our planet as well and that meteorite impacts changed the surface of the Earth and its geological history (see Melosh 1989; French 1998, and references therein). Furthermore, impact events also led to geological structures with some economic value. Large mineral resources are associated with some large impact structures, e.g., the world’s largest nickel-copper sulphide deposits at the Sudbury impact structure,
Canada (e.g., Pye et al. 1984; Dressler et al. 1994; Grieve and Masaitis 1994; Lightfoot and Naldrett 1994; Grieve and Therriault 2000) or the world’s largest gold deposit in the Witwatersrand Basin in South Africa (e.g., Reimold et al. 2005).

Large meteorite impacts also had devastating consequences for life on Earth. The possibility that the mass extinction 65 Ma ago – including the end of most dinosaurs at the Cretaceous/Tertiary boundary – was caused by a catastrophic impact event is now generally accepted (e.g., Hildebrand et al. 1991, Sharpton et al. 1992; for a debate about this topic, see, e.g., Silver and Schultz 1982; Sharpton and Ward 1990; Ryder et al. 1996). The investigation of the relationships between large impacts on Earth and other mass extinctions in the Earth’s history has become an active field of study in recent years (e.g., papers in Koeberl and MacLeod 2002, or in Buffetaut and Koeberl 2002).

While abundant impact structures with different diameters are known from almost every continent (e.g., Grieve 1991, 1998; Melosh 1989; French 1998, 2004), only a small number of structures have so far been identified on ocean floors (see French 1998, 2004, and references therein). This is due to the relatively young age of the ocean floors (Grieve 1998), and because investigation of marine impact structures has received more attention only in recent years (e.g., Gersonde et al. 1997; Smelror et al. 2001; Dypvik and Jansa 2003; Poag et al. 2004).

The first impact crater on Earth (Barringer Crater, Arizona) was recognized in the early stages of the 20th Century (Barringer, 1906, as cited in Melosh 1989) but remained controversial for several more decades. Since then, more than 170 impact structures have been identified on Earth (Grieve, 1998), with new impact structures being found every year. An updated impact structure record can be viewed on the website of the University of New Brunswick (www.unb.ca/passc/ImpactDatabase).

The comparatively huge number of impact structures found on other planets and planetary bodies in the Solar System indicates that Earth also must have been the target of extraterrestrial bodies throughout its history. The number of crater structures on other planets makes it possible to determine the intensity of bombardment, and, as a result, it is possible to estimate the flux of bodies that must have hit the Earth (e.g., papers in Grady et al. 1998). To understand the processes that are involved in a collision, studies of terrestrial impact structures are necessary, as only they can provide hands-on experience of impact-related processes. Numerous well-preserved and well-exposed impact structures are found in a variety of sizes on the Moon and
other planets ranging from microcraters <1 mm across to great ring basins >2000 km in diameter (French 1998) (Figs 1.1 and 1.2); however, they are not directly accessible for geoscientific analysis. The intensity of the bombardment and the ages of impact craters on planetary surfaces could be established by the abundance of well-preserved impact structures on their surfaces with the simple logic that older surfaces accumulate more craters than younger ones (Shoemaker and Hackman 1962; Shoemaker et al. 1963). In the case of the Moon, this method was combined with radiometric dating of recovered samples. It was determined that the most intense bombardment of the Moon occurred at about 4.4 Ga, immediately after the formation of the Moon, decreasing fast until 3.8 Ga, and then decreasing further at a slower rate (Wilhelms et al. 1987; Hörz et al. 1991; Taylor 1992). Although the rate of bombardment was suggested to be fairly constant after 3.8 Ga, a controversial discussion commenced about a variation in projectile flux of ~2x during the Phanerozoic (see French 1998 and references therein). On Earth, the uppermost levels of some, especially of the few known, large and old, impact structures have been removed by erosion and the present surfaces expose deeper levels of the structures. Therefore, the study of impact structures on Earth provides a third dimension that is missing from the largely two-dimensional, imagery-based data sets for other planetary impact crater populations.

Fig. 1.1: Example of an unnamed simple crater on the Moon, with a diameter of about 1.5 km (Source: www.nasa.gov/multimedia). The typical bowl-shape of the crater and the elevated rim are well preserved. Also clearly visible is the ejecta blanket surrounding the crater.
While numerical modelling of impact cratering processes has become a main focus of impact cratering studies (e.g., Ivanov and Deutsch 1999; Melosh and Ivanov 1999; Collins et al. 2004) and provides useful information about the cratering process, data from field observations are still a crucial part of the investigation of terrestrial impact structures and are also required to provide important controls for numerical modelling (Pierazzo and Herrick 2004).

**Fig. 1.2:** Perspective view of the 93 km wide complex Copernicus impact structure on the Moon at the southern edge of the Mare Imbrium. The elevated central region is visible indicated by a peak ring. Furthermore, the terraced walls at the crater rim indicate inward slumping of material off the crater rim, a typical feature of large impact structures (Source: www.nasa.gov/multimedia).

### 1.3 Identification of impact structures

Impact structures that were found in the past centuries but not interpreted as such, were identified due to their circular shapes (actual crater shapes, lakes or rings of hills). They were usually related to endogenic processes, such as volcanic activity, but not to meteorite impact. Where impact melt rock was noted in outcrop, it was taken for magmatic rock and classified as such (e.g., review in Reimold 1980). Other circular structures attracted attention because of the upturned bedding orientation and elevated central regions within a circular structure, similar to features of salt domes, or because of the presence of unusual breccias (French 1998). Some impact structures have been found by accident in the course of fieldwork or geological surveys. Since the interpretation of the first impact crater (Barringer Crater, 1906; see e.g., Melosh 1989; French 1998), many “circular structures” have been investigated. The
recognition of impact structures, however, has been marked by debate and controversy (see e.g. Dietz 1963; Bucher 1936, 1963; French 1968, 1990; Sharpton and Grieve 1990; Nicolaysen and Ferguson 1990).

With both increased interest in this geological field and more advanced technologies, such as remote sensing and geophysical methods (aeromagnetic and gravity surveys), the discovery of possible impact structures dramatically increased in the 1960s to 1980s (for a review, see, e.g., Melosh 1989; French 1998).

1.3.1 Geophysical and macro-deformation signatures

The presence of a distinct circular depression may only help to identify relatively young or well-preserved structures. Where the circular shape may be completely absent, e.g., at older and relatively deeply eroded structures, localized intense deformation (fracturing, faulting) or unusual brecciation (e.g., impact melt or pseudotachylitic breccias) could attract attention. Intense fracturing can extend to several kilometres below the actual crater floor (to depths up to one-third of the diameter of the structure; Melosh 1989; Collins et al. 2004; Dence 2004) and may be detectable by geophysical surveys. The structures may occur as negative gravity anomalies with a general circular shape due to the regular distribution of fractured rocks and breccias of the crater structure, which are less dense than the original target rock. In large, complex structures, however, the rise of denser rocks from beneath the crater floor, forming the central uplift (see section 1.4), may reduce the negative anomaly or even turn it into a positive anomaly in the centralmost parts of the structure (see e.g., Stepto 1990; Pilkington and Grieve 1992; Grieve and Pilkington 1996). Magnetic anomalies of impact structures can be positive or negative and are caused by the impact-induced magnetization of an impact melt and central uplift rocks, or the uplift in the central parts of more magnetic strata (e.g., Grieve and Pilkington 1996; Henkel and Reimold 2002). Seismic profiling has also helped in the past to investigate the structure and deformation at deeper levels of the structures (e.g., Gosses Bluff, Australia, Milton 1977; Milton et al. 1996; Chicxulub, Mexico, Morgan et al. 1997). Furthermore, about one-third of today’s known impact structures were discovered by seismic studies, because they are buried beneath younger sediments (Grieve 1991; Grieve et al. 1995), e.g., the Chicxulub Structure in Mexico and other, hydrocarbon-bearing structures in North America (Grieve 2005).
It is, however, important to emphasize that these techniques alone do not provide unambiguous confirmation of the presence of an impact structure. A circular feature discovered by these methods may have various origins. Verification on the ground is still necessary, in every case. The verification by field geology requires the identification of definitive impact-induced features (shock deformation) and, therefore, the collection of rock samples, or the identification of traces of the meteoritic projectile (e.g., Montanari and Koeberl, 2000).

1.3.2 Shock deformation features

The extreme pressures, temperatures and strain rates (e.g., Spray 1998) created by a shock wave propagating through the target rock immediately after the impact produces deformation effects that include elastic and plastic mineral deformation, mineral transformation, and melting in the target rocks. They include, inter alia, shatter cones (Dietz 1947, 1959; French 1998; Nicolaysen and Reimold 1999), high-pressure polymorphs of quartz and other rock-forming minerals in upper crustal rocks (White 1993; Stöffler and Langenhorst 1994), planar deformation features (PDFs) in important rock-forming minerals (e.g., Stöffler and Langenhorst 1994) and the formation of diaplectic glasses and mineral and rock melts (see French 1998 and references therein). Diamonds may also be formed in carbon-bearing target rocks (e.g., Masaitis et al. 1972; Masaitis 1998; Koeberl et al. 1997). More recently, nanodiamonds were found in suevite breccia (Hough et al. 1995) and in ejecta at the K/T boundary (Carlisle and Braman 1991; Hough et al. 1997).

Only with the establishment of diagnostic impact-deformation features (shock metamorphic features, e.g., French and Short 1968; Stöffler 1972, 1974; Stöffler and Langenhorst, 1994; Grieve et al. 1996) did scientists start to unequivocally confirm the existence of crater structures caused by meteorite impacts. These “shock effects” are crucial for recognizing a meteorite impact, as only a few (about a dozen, see French 1998) impact structures have actually preserved components of the meteoritic projectile associated with them.

The most accepted and widely distributed criterion is microscopic planar deformation features (PDFs) in quartz and other rock-forming minerals. They may occur in clasts of crater-fill breccias and impact melt rocks, as well as in rocks beneath the crater floor or in the central uplift. PDFs are characterized by single or
multiple sets of narrow-spaced planar lamellae, which are not open and can therefore be easily distinguished from cleavage or deformation lamellae produced in non-impact tectonic processes (Carter 1965, 1968; Alexopoulos et al. 1988; Stöffler and Langenhorst 1994; French 1998). PDFs in quartz consist of highly deformed or amorphous quartz (e.g., Müller 1969; Kieffer et al. 1976; Goltrant et al. 1991, 1992) and are parallel to crystallographic planes in the quartz grains, preferentially to the basal plane c (0001), but also to the zones \{10\bar{1}3\}, \{10\bar{1}2\} and \{10\bar{1}1\} (French 1998). Their distinctive appearance can easily be seen in samples from relatively young structures.

PDFs in samples from older and more strongly altered or metamorphosed rocks are different (see French 1998 and references therein). PDFs are still distinctive in appearance but the amorphous material is recrystallized to quartz again, thus leaving planar arrays of small inclusions (French 1998). These so-called decorated PDFs (Robertson et al. 1968; Stöffler and Langenhorst 1994) still show the original PDF orientations. More strongly recrystallized PDFs show small mosaics of quartz crystals (subgrains), predominantly along orientations originally parallel to (0001) (Leroux et al. 1994). Beyond the range of shock pressure in which PDFs are produced (>7-35 GPa, Stöffler and Langenhorst 1994; Huffman and Reimold 1996), an amorphous glass phase is produced, the so-called diaplectic glass (also called thetomorphic glass, Stöffler 1966, 1967, 1972, 1984; Chao 1967). Diaplectic glasses are completely different from normally produced glasses (produced by heating a mineral above its melting point). Diaplectic glasses did not fuse or flow and, therefore, the original textures of the crystals and the original textures of the precursor mineral are still preserved (French 1998).

In addition, quartz in impact structures may be transformed to high-pressure polymorphs and, in places, to lechatelierite (fused quartz) in impact melts (French 1998). The discovery of coesite and stishovite in impact structures (Shoemaker et al. 1961) led to the rejection of the hypothesis for a volcanic origin of those structures, as volcanic eruptions are not able to produce shock pressures high enough to produce the high-pressure polymorphs of quartz at or near the surface of the Earth (Fig. 1.3).

Remnants of the projectile material can also be found, which is expressed by elevated abundances of siderophile elements (Ni, Co) and Cr, or by unusually high PGE (such as iridium or osmium) content or distinctive osmium isotope ratios. Very
low \(^{187}\)Re/\(^{188}\)Os and \(^{187}\)Os/\(^{188}\)Os ratios are characteristic for a meteoritic component in impact breccia (see e.g., Koeberl et al. 1996; Montanari and Koeberl 2000).

Fig. 1.3: Conditions of shock metamorphism: Pressure-temperature diagram showing the conditions for shock metamorphism and normal crustal metamorphism (shaded area to the left). Approximate formation conditions for specific shock effects (labelled) are given by vertical dashed lines. Stability curves for high-pressure minerals (coesite, diamond, stishovite) are shown for static equilibrium conditions (from French 1998).

In the past few years a new technique was introduced to determine a meteoritic component in rocks. This method is based on the \(^{53}\)Cr/\(^{52}\)Cr ratio in samples (Koeberl and Shirey 1997; Shukolyukov and Lugmair 1998). The \(^{53}\)Cr isotope derives from the \(^{53}\)Mn isotope. Terrestrial rocks do not show a variation in the \(^{53}\)Cr/\(^{52}\)Cr ratio, because \(^{53}\)Mn had decayed long before the homogenisation of the Earth. Therefore, a variation in the \(^{53}\)Cr/\(^{52}\)Cr ratio is indicative for an extraterrestrial component. With this new method it is, therefore, not only possible to determine a meteoritic component in possible target rocks, but one is also able to distinguish between different types of meteorites (Shukolyukov and Lugmair 1998). This method, however, is complicated and time consuming. Furthermore, a significant amount of Cr in the sample is necessary to be of extraterrestrial origin in order to make a distinction from the terrestrial Cr isotopic value. Also, this technique is not as sensitive as the Os based
technique. The $^{187}\text{Os}/^{188}\text{Os}$ isotopic ratios are significantly different in meteorites and terrestrial crustal rocks. In terrestrial rocks the $^{187}\text{Os}/^{188}\text{Os}$ ratio increases fast with time, in contrast to the ratio in meteorites where this ratio only shows small changes.

Dietz (1947) deserves the credit for first describing shatter cones as a probable criterion for the recognition of meteorite impact structures. His argumentation was supported by Shoemaker et al. (1961), who produced shatter cones in small-scale impact experiments. When shatter cones were discovered at other potential impact structures (e.g., Sudbury and Vredefort, Dietz 1964), these features became widely accepted as a diagnostic shock-deformation phenomenon (French 1998). Shatter cones are probably not only the most distinct megascopic impact-related deformation feature, but they may also be widely distributed in an impact structure (for a detailed review see Chapter 3).

Another feature observed in impact structures is impact breccias (for nomenclature, see Stöffler and Grieve 1994). These can occur as deposits within the crater (crater-fill breccias) or outside of the crater (ejecta) (e.g., French 1998). In addition, the subcrater floor may also contain brecciated fragments of the target rocks (lithic breccias and breccia dykes). Lithic breccias consist of rock and mineral fragments, whereas breccia dykes are divided into four categories (e.g., Bischoff and Oskierski 1988; Dressler and Sharpton 1997: (1) They can have a composition similar to the lithic breccias; (2) Suevitic dike breccias that comprise fragments of glass or melt as well as rocks and minerals in a clastic matrix; (3) impact-melt breccias consist of rock and mineral fragments in a glassy or crystalline melt matrix; and (4) Impact melt rocks of a glassy or crystalline melt with only a small proportion of inclusions.

A further type of breccia in impact structures is pseudotachylitic breccia (see review in Reimold 1995, 1998; Reimold and Gibson 2005). These breccias range in thickness from less than 1 mm to many metres and can exceed strike lengths of hundreds of metres to many kilometres (in the largest impact structures such as Sudbury and Vredefort). Pseudotachylitic breccias commonly contain a fine-grained glassy to crystallised melt matrix and subangular to rounded clasts of various sizes (see Chapter 4). However, as similar breccias can also be produced by normal tectonic processes, they are not definitive impact features. For a review on this issue see Dressler and Reimold (2004), Reimold and Gibson (2005), and Chapter 4.
1.4 Crater morphology

Impacts of hypervelocity bolides leave scars on the Earth’s surface. Small meteorites lose their kinetic energy and velocity when passing through the atmosphere, and disintegrate or lose mass by ablation. Larger bolides do not slow down enough in the atmosphere and result in a low-velocity impact, leaving only a small, shallow crater (French 1998). Impacts of even larger bolides result in wide, relatively shallow, circular structures with diameters of kilometres to hundreds of kilometres (Melosh 1989). A distinction is made between simple craters, complex craters and multiring impact basins (French 1998). Simple craters (Fig. 1.4) show a bowl-shaped morphology of less than 2-4 kilometres in diameter, the size depending on the composition of the target sequence. Simple craters do not show much modification of their transient cavity and are only modified by minor collapse of the crater walls and by redeposition of ejected material into the crater. The final crater may be ~20% larger than the transient cavity, but the original depth remains more or less the same (Melosh 1989) (Fig. 1.4c). The crater floor is covered with shock-melted rock (impact melt) and fallback of ejected material.

The structure of larger impact craters is more complicated, and they are called complex impact structures. They are characterized by a central uplift region, a flat floor, and terraced walls. The terraced walls form due to rocks collapsing inward and due to gravitational adjustment in the final stages of crater development (e.g., Dence 1968; Grieve et al. 1977, 1981; Grieve 1991; French 1998). A variety of different central uplifts are possible, ranging from single peak structures to peak rings. The transient cavity of large impact structures is intensely modified during the formation of the final crater structure (Fig. 1.5). The rapid changes of movement directions and the vertical rise during development of the central uplift are easier to identify in sedimentary sequences with marker horizons than in massive crystalline rocks. A detailed statistical analysis of some complex impact structures (Grieve et al. 1981; Grieve and Pilkington 1996) has revealed that the actual stratigraphic uplift (SU) in impact structures can be approximated as SU = 0.086 D^{0.03} (D = final diameter of the structure) or, more simply, as about one-tenth of D. Therefore, it is suggested that rocks beneath the centre of large impact structures (D = 100 – 200 km) are uplifted by 10-20 km. The rise and subsequent collapse of the central parts of large structures is
Fig. 1.4: Schematic illustration of the evolution of a simple crater. (a) The meteorite collides with the target rocks. (b) Material is ejected out of the crater, thus forming the transient cavity. (c) A simple crater is characterized by a simple bowl-shaped morphology; (d) The final crater shows a similar shape and size to that of the transient cavity (Source: www.nasa.gov/multimedia).
**Fig. 1.5:** Schematic illustration of the formation of a complex crater. The most important difference to simple craters, besides the large size, is the presence of a central uplift. There is no sharp boundary between the different stages, rather a transition from one stage to another. (a) Contact/compression phase: Initiation of the shock wave, material is moved out-and downward (arrows). (b) Excavation phase: Ejection of material and formation of a transient cavity. (c) Modification phase: Rise of the central uplift, and (d) subsequent collapse. (e) Final stages of the modification phase is characterized by inward slumping of material from the crater rim (arrows; source: www.nasa.gov/multimedia).
very rapid and is estimated to take place within a few minutes (Melosh 1989; Henkel and Reimold 1998). In the case of the Vredefort impact structure the maximum
duration for the modification phase was estimated at ~15 minutes by Henkel and
Reimold (1998). However, the exact processes that take place during the formation of
the central uplift are still not well established (e.g., Collins et al. 2004; Dence 2004;
Ivanov 2005).

Field observations from central uplifts mostly stem from the highly
fragmented upper parts of shallowly exhumed structures (e.g., Sierra Madera,
Wilshire et al. 1972; Gosses Bluff, Milton et al. 1996; Upheaval Dome, Kriens et al.
1999; Kenkmann et al. 2005; Haughton, Bischoff and Oskierski 1988; Puchezh-
Katunki, Ivanov 1994). Besides numerical modelling of the deeper levels of central
uplifts in recent years (e.g., Collins et al. 2004; Dence 2004; Ivanov 2005), field
evidence of such processes can only be found in the deeply eroded central parts of the
Vredefort Structure. The results of numerical modelling (e.g., Collins et al. 2004;
Ivanov 2005) indicate that such deep levels, which are presently exposed in the collar
rocks of the Vredefort Dome, are beyond the limit of fragmentation. In general,
extensive movement patterns, especially changes in lateral movements, are not
expected, and, instead, vertical motion might have dominated (for a detailed
discussion on this issue see Chapter 6). As a consequence, the inner parts of the
Vredefort Structure offer a unique opportunity to test the modelling results in the
field. Only limited structural work has been carried out, to date, in the Vredefort
Dome (e.g., Manton 1962, 1965; Lilly 1978, 1981; Lana et al. 2003a-c, 2004), and
most studies dealt with specific features (such as shatter cones, pseudotachylitic
breccias). This demonstrates the need for further detailed structural analysis in this
impact structure, such as this study. For a review of the structural work in the
Vredefort Dome see below (section 1.7) and Chapter 2.

The inner parts of large impact structures, such as the central floor and the
central uplift, are commonly covered by an impact melt sheet and crater fill breccia
(Melosh 1989; French 1998).

The diameter of the transient crater in large impact structures is estimated to
range from 0.5 to 0.7 of the diameter of the final crater (e.g., Melosh 1989; Therriault
et al. 1997). Impact structures with a diameter of more than 100 km are even more
complicated (French 1998). Structures with diameters between a few hundred
kilometres and more than 1000 km are called multiring basins, but have only been
found on other planets (e.g., Moon, Mercury, Mars; e.g., Spudis 1993). However, some authors (e.g., Grieve and Therriault 2000) have discussed that the Vredefort, Sudbury and Chicxulub impact structures could represent terrestrial multi-ring basins.

1.5 Mechanics of formation of complex impact structures

The formation of large impact structures can be divided into different stages (e.g. Melosh 1989; French 1998) (Fig. 1.5): the contact/compression phase, the excavation phase and the modification phase.

The contact and compression phase begins when the meteorite comes into contact with the ground and explodes under its own compression (Melosh 1989), which results in a shock wave that moves down- and outward from the explosion site. This causes compression of the target rocks and downward and outward acceleration of target material from the point of impact. At the time of impact, the kinetic energy of the meteorite is transferred into a shock wave. A shock wave is an intense, transient, high-pressure stress wave (Melosh 1989). This shock wave is triggered at the contact between the meteorite and the target, where it reaches pressures of up to hundreds of GPa or more, and speeds of more than 10 km/s, in the target rocks (Melosh 1989; French 1998). When the shock wave reaches the back end of the impacting body, it is reflected forward again into the body. This wave is called the rarefaction or tensional wave (Melosh 1989; French 1998). The contact/compression phase ends when the rarefaction wave from the rear of the projectile enters the target rocks and, thus, produces release of the shock pressure, which then leads to vaporisation and melting of both the target rocks and the projectile.

The short contact/compression phase is immediately followed by the comparatively longer excavation stage, during which the actual crater is opened up. The shock wave expands hemispherically and weakens as a result of counteraction caused by the material strength of the target materials until the shock wave is turned into a seismic (elastic) wave. The shock wave and its rarefaction wave cause the movement of the target rocks down- and outward; thus, a hemispherical transient cavity is formed and a large amount of material is ejected into the atmosphere and into the surrounding area. In very large impact structures this phase can last up to several minutes (Melosh 1989; French 1998).
The duration of the *modification stage* is similar to that of the preceding excavation phase. It is characterized by a broadening and shallowing of the transient cavity and by slumping and debris sliding back towards the centre of the crater. During this phase the central parts of complex impact structures rise and form the central uplift. However, as the change from the excavation phase to the modification phase is transitional, some authors argue that central uplift formation has already started before the end of the excavation phase (e.g., Henkel and Reimold 1998).

Subsequently, in the largest impact structures, the central uplift collapses. In complex impact structures, terraced walls are formed due to the gravitational instability of target material at the elevated crater rim. As a consequence, the crater walls slump inwards (Melosh 1989; French 1998). This phase is characterized by rapid changes of the stress environment, i.e., first inward movement of material due to the rise of the central uplift, followed by outward movement during the collapse of the central rise and then inward movement of crater wall material. The question whether these movements follow each other in a sequence or overlap is debated (see, e.g., Henkel and Reimold 1998). The phase does not last for more than 5-10 minutes, even in very large structures and is clearly indicative for extremely high strain rates. Impact-related deformation (e.g., seismic activity, fracturing) may continue for much longer, due to gravitational adjustment and cooling of the target rocks.

In structures with a peak ring in their centre, the central uplift has risen beyond its gravitational stability and, subsequently, collapsed, leaving a trough behind in the central parts surrounded by a peak ring. As a consequence, in- and upward movement during central uplift formation is overprinted by the outward movement during collapse of the central uplift in peak ring structures. In contrast, craters with a central peak indicate that the central uplift has not risen beyond this point, and, thus, no peak ring forms (Melosh 1989).

The exact mechanisms that take place during the modification of an impact structure are not completely understood yet. The speed with which the central uplift forms and the subsequent collapse occurs suggests drastic degradation of the strength of the rocks in the subcrater basement during and after the passage of the shock wave (Melosh 1989). Melosh (1989) suggested that strong vibrations initiated by the shock wave in the subcrater basement are responsible for the temporary strength reduction. He termed this process “acoustic fluidization”. Ivanov et al. (1996) suggested that these strong vibrations affected fault-bounded blocks, which were tens to hundreds of
metres in size. This suggestion seems to require the presence of a megabreccia zone in
the upper part of the central uplift volume (e.g., the Puchezh-Katunki structure;
Ivanov 1994). Ivanov et al. (1996) proposed that such blocks are able to oscillate
independently of their neighbours. Studies from other impact structures (e.g., Deep
Bay and Clearwater Lake, Canada, Dence 2004) support this theory where drilling
results revealed that the upper parts in the central uplift area consist of strongly tilted
material that is apparently broken into large blocks. Dence (2004) noted a lack of
fragmentation at a mesoscopic scale in the central parts of these structures, but
suggested a movement of individual, massive blocks separated by shear zones. He
also postulated an increase in size of the individual blocks with increasing crater size.
Results from gravity studies confirm a low fracture porosity in rocks of central peaks
in craters with diameters >30 km (Sweeney 1978). In such large impact structures the
fracture porosity increases with distance from the centre and is highest in the ring
zones and the peripheral trough (Sweeney 1978; Dence 2004). The fracture porosity is
notably negatively correlated to the shock pressure distribution in large structures
(e.g., Dence 2002). Dence (2002) also postulated that rocks in the central peak of
large impact structures (>30 km) dilate as they rise, thus allowing impact-induced
melts to intrude these extensional sites. Furthermore, Dence (2002, 2004) stated that
the rise of the central uplift is triggered by relaxation of the elastically compressed
rocks in the floor beneath the central parts of the impact structure, but is further driven
by inward movement of the collapsed crater rims, which itself is driven by gravity.

In the case of the Vredefort Dome, recent studies (Lana et al. 2003a-c, 2004)
demonstrated an absence of such megablocks in the core. These authors suggested
that slip could have been accommodated by extensive microfracture or
pseudotachylitic breccia networks instead. Based on the rotation of pre-impact
gneissic fabrics in the outer parts of the core of the dome and their horizontal
orientation in the inner parts of the core, Lana et al. (2003a) suggested a plug-like
generation for the core of the Vredefort Dome (see section 1.7). This would be
consistent with the theory of rotation of strata during the outwards collapse of the
central uplift.
1.6 Remaining problems regarding the Vredefort Dome

An impact origin of the Vredefort Dome (or Ring, as it was known in the past) was first discussed in the mid-1930s (Boon and Albritton 1936), but remained controversial for decades. Supporters of an endogenic (cryptoexplosion) origin and of an impact origin argued for decades (e.g., papers in Nicolaysen and Reimold 1987; reviews by Reimold 1993, Reimold and Gibson 1996 and Gibson and Reimold 2001), until the impact origin of the Vredefort Dome was finally established through the confirmation of *bona fide* shock deformation features, such as high-pressure polymorphs of quartz (stishovite and coesite, Martini 1978), planar deformation features in quartz and planar fractures in zircon (e.g., Carter 1965; Lilly 1981; Grieve et al. 1990; Reimold 1990; Leroux et al. 1994; Kamo et al. 1996), and even the existence of a small meteoritic component in Vredefort impact melt rock (Koeberl et al. 1996).

Whilst an impact origin of these microdeformation features, and, thus, for the Vredefort Dome, has been established, the origin of meso- to macroscopic structures (such as folds, faults, fractures) in the Vredefort Dome and in the outer parts of the impact structure have remained strongly debated (see section 1.9). Therefore, a detailed analysis of the deformation in the collar rocks was needed in order to find evidence for or against an impact origin of these structures. As a consequence, this study was carried out to investigate geometric and temporal relationships between these structures (folds, faults and fractures) in the collar rocks and the established impact-related deformation features (shatter cones, pseudotachylitic breccia). The results of this study may also contribute to a better understanding of the processes involved in the development of the central parts of very large impact structures, which are still largely enigmatic due to a general lack of field evidence.

The origin of pseudotachylitic breccias in the Vredefort Dome, by shock and/or friction melting, remains unresolved (for recent reviews, see Dressler and Reimold 2004; Reimold and Gibson 2005, 2006 and Chapter 4). Thus, this study included an investigation of the pseudotachylitic breccias and their relationships (spatial, geometric) to meso- and macroscopic structures (folds, faults, fractures) that are present in the collar rocks of the Vredefort Dome. In addition, in an attempt to contribute to a better understanding of the development of the pseudotachylitic breccias in the context of the rapidly-changing strain patterns in an impact structure,
and the heavily debated issue of shock- versus friction-generated melts, cooling calculations for melt veins of different composition and different thickness were conducted (see Chapter 4).

Even less is known about the structures in the outer parts of the Vredefort Structure (see Fig. 1.6a,b). Some authors (e.g., Brink et al. 1997, 2000a,b) proposed the presence of concentric faults around the Vredefort Dome and related them to the remnants of ring structures that were formed by the inward slumping impact-crater walls. Whilst locally inward-dipping faults were reported from underground geological work in the goldfields of the Witwatersrand Basin (e.g., Fletcher and Reimold 1989; Killick and Reimold 1990; Killick 1993; Phillips and Law 1994) and from a detailed study of the Potchefstroom Synclinorium (Simpson 1977, 1978), very poor surface exposure of structures in the outer parts of the Vredefort Structure hampers the confirmation of such faults on a regional scale. Other authors (e.g., McCarthy et al. 1986, 1990; Courtnage 1996; Gibson et al. 1999) suggested that impact-related deformation could be traced as far as the northern margin of the Witwatersrand Basin. Small-scale thrusts and associated folds have been reported from around Johannesburg, ca. 150 km from the crater centre, and were linked to the Vredefort event (McCarthy et al. 1986, 1990). McCarthy et al. (1986, 1990) also stated that post-Transvaal cleavages and folds on the northern edge of the Witwatersrand Basin showed a radial outward vergence away from the Vredefort Dome. Some authors (e.g., Friese et al. 1995) have, however, related the concentric faults to northwesterly-directed thrusting in post-impact times linked to the Kibaran (Grenville) Orogeny at ~ 1.25 – 1.0 Ga (see Chapter 5 for further details).

Field mapping was carried out in the area between the towns of Potchefstroom, Ventersdorp and Carletonville (see Chapter 5 and Fig. 5.1), in order to investigate the possible presence of evidence for or against the argumentation that concentric faults existed which were related to the impact event and represented outer rings of the Vredefort Structure (Brink et al. 1997, 2000a,b; Friese et al. 1995). Given the complex pre-impact tectonic history of the Witwatersrand Basin, a good knowledge of the tectonic events in the entire Witwatersrand Basin in general is essential for an analysis and comparison of structures in the area in and around the Vredefort Dome. Thus, the tectonic history of the Witwatersrand Basin prior to the impact event is outlined in the following section.
1.7 Geology and evolution of the Witwatersrand Basin

The Kaapvaal Craton is considered one of the largest pristine Archaean crustal fragments on Earth (besides the Pilbara craton in Australia) (e.g., De Wit 1998; Schmitz and Bowring 2003). Geochronological data suggest that several crustal fragments were accreted from ~2.9 Ga until ~ 2.6 Ga (e.g., Kimberley block, Amalia-Kraaipan granite-greenstone terrane, Poujol et al. 2000, 2003; Schmitz et al. 2004).

Based predominantly on surface geology and gravimetric and magnetic data, De Wit et al. (1992) divided the Kaapvaal Craton into a number of sub-domains separated by tectonic boundaries (see also Schmitz et al. 2004).

1.7.1 Geological Setting of the Witwatersrand Basin

The Witwatersrand Basin lies in the central parts of the Kaapvaal Craton and has an extent of ~350 km from the southwest to the northeast and about 200 km in a northwest-southeast direction (Corner and Wilsher 1989) (Fig. 1.6a).

The mining activities in the Witwatersrand Basin of past decades (Fig. 1.6b) have resulted in a much-improved understanding of the structural setting of the basin. This has been supported by studies on the surface. Based on these recent studies, at least four major deformation phases have been identified to have affected the Witwatersrand Basin (e.g., McCarthy et al. 1986; Myers et al. 1990, 1992; Roering et al. 1990; Friese et al. 2003).

These deformation events are characterized by episodic changes of the deformational environment, from extensional to compressional regimes and back, over a time span of nearly 1000 million years (Roering et al. 1990; Friese et al. 2003). Several models for the structural evolution of the Witwatersrand Basin have been described in the past. Pretorius (1975, 1981) presented a model for a rift setting in which the Witwatersrand sediments had been deposited in a structurally controlled basin or half-graben that was bounded by faults to the northwest but gently downwarped towards the southeastern margin. A slightly different rift model was proposed by Minter (1978), who suggested epeirogenic tilting and warping of the strata associated with granite doming around the basin as the reason for a closure of the Witwatersrand Basin. Other models explaining the evolution of the Witwatersrand
Fig. 1.6: (a) Schematic map of southern Africa, showing the extent of the Kaapvaal Craton and the location of the Witwatersrand Basin, and the Vredefort Dome in its central parts (modified after Lana et al. 2004). (b) Simplified geological map of the Witwatersrand Basin showing the stratigraphy and distribution of lithologies (post-Witwatersrand rocks are stripped away). The main goldfields in this area are also indicated (after Robb and Robb 1998). (c) Simplified map of the Vredefort Structure showing the stratigraphy and the distribution of lithologies within and around the structure (after Lana et al. 2003a). Also shown is the rim syncline surrounding the structure (Potchefstroom Syncline). Note that the Karoo Supergroup cover has been removed.
Basin during the sedimentation of the Witwatersrand Supergroup sediments included a taphrogenic basin model (Pretorius, 1981) and an intracratonic model (e.g., Vos, 1975). Pretorius’ (1981) model was based on an earlier idea by Brock and Pretorius (1964) that suggested interference of vertical cross-folds having been responsible for developing and closing of the basin. These folds were produced by vertical tectonics (Pretorius 1981) that formed synclines and anticlines (Fig. 1.7a). The resulting basin was then further affected by updoming events that influenced the palaeocurrent patterns of fluvial systems in the basin (Pretorius 1981). Vos (1975) supported this intracratonic setting scenario but introduced a tectono-stratigraphic concept in which sedimentary cycles were reworked.

A plate-tectonic setting for the Witwatersrand Basin was proposed by several authors (e.g., Van Biljon 1980; Bickle and Eriksson 1982; Burke et al. 1985). Van Biljon (1980) suggested that the basin was formed along a suture zone (continent-continent collision), whereby the basin represents a remnant embayment on the edge of one of the continents (Fig. 1.7b). Bickle and Eriksson (1982) proposed a cratonic-shelf setting for the Witwatersrand Basin and a subsequent deposition of a cratonic, rift-controlled sequence during Ventersdorp sedimentation.

Burke et al. (1985) concluded the basin was a foreland basin, based on the observation that the rifting was developed transverse to a collision zone between the Kaapvaal and Zimbabwe cratons (Fig. 1.7c). Recent studies support a foreland basin tectonic setting for the Witwatersrand Basin (e.g., Burke et al. 1986; De la Rey Winter 1987; Stanistreet and McCarthy 1991) but the structural systems that alternated over a period of over one thousand million years are still under debate. What is widely accepted, however, are roughly north-south and east-west striking fault systems, dividing the Witwatersrand Basin into a mosaic-like block pattern (e.g., Myers et al. 1990, 1992, McCarthy et al. 1990).

Fig. 1.7: Next page: Different models for the evolution of the Witwatersrand Basin were proposed in the past by several authors. (a) A basin-and-dome structural model with interfering orthogonal sets of folds (labelled A and B, modified after Pretorius 1981), where the synclines of folds A and B are superimposed on each other, thus resulting in a thicker deposition of sediments in the respective synclines. (b) An intra-sutural basin model, modified after van Biljon (1980). The basin developed along the suture zone of a continent-continent collision (numbers 1 to 3). Rifting occurred during the emplacement of the Bushveld Complex (number 5). (c) A plate-tectonic model (modified after Burke et al. 1985) in which the Witwatersrand Basin developed in a foreland basin during rifting between the Kaapvaal and the Zimbabwe cratons. For further details see text.
The Witwatersrand Basin rocks are underlain by 3-3.2 Ga granitoids and older greenstone remnants of the Kaapvaal Craton (Lowe 1999; Poujol et al. 1999, 2001) that consist mainly of polydeformed Archaean migmatitic gneisses, with subsidiary mafic and metasedimentary xenoliths. These rocks experienced mid-crustal, upper amphibolite- to granulite-facies metamorphism at ~3.1 Ga (Stepto 1979, 1990; Hart et al. 1999; Moser et al. 2001; Lana et al. 2003a; Armstrong et al. 2006). The granitic basement is exposed in the central parts of the basin, in the Vredefort Dome (Fig. 1.6c), in the northeastern part in the Devon Dome, in the north in the Johannesburg Dome, and along the northwestern margins of the Witwatersrand Basin (e.g., Corner and Wilsher 1989) (Fig. 1.6b).

The rocks of the Witwatersrand Basin comprise a metamorphosed sequence of sedimentary and volcanic rocks deposited between 3.07 Ga and ~2.7 Ga in a succession of basins on the Kaapvaal Craton (Clendenin et al. 1988; Walraven and Martini 1995) (Fig. 1.8).

1.7.1.1 Dominion Group

The oldest sedimentary rocks in the Witwatersrand Basin belong to the Dominion Group with an age of ~3.074 ± 0.009 Ga (Armstrong et al. 1991). This group includes basaltic andesites as well as felsic lavas and rift-related clastic sediments, with an estimated maximum thickness of 2.5 km in the Klerksdorp area. In the Vredefort Dome, only andesitic metalava of amphibolite metamorphic grade is exposed at a maximum thickness of ~400 m (Jackson 1994).

The first identified deformation phase in the Witwatersrand Basin took place in an extensional tectonic environment, during the early-Dominion-related rifting (~ 3.074 ± 0.009 Ga) resulting in NE-SW trending faults (e.g., Frimmel and Minter 2002; Lana et al. 2006) (Table 1.1).

1.7.1.2 Witwatersrand Supergroup

The Witwatersrand Supergroup comprises ca. 7 km of predominantly clastic metasediments that were deposited between ~3.0 and 2.7 Ga (Robb et al. 1997, Robb and Robb 1998) (Fig. 1.9). The Witwatersrand Supergroup is divided into the West
Fig. 1.8: Schematic stratigraphic column for the Witwatersrand Basin and the ages for the deposition of the strata (modified after Brink et al. 1997).
Rand Group (lower Witwatersrand Supergroup) and the Central Rand Group (upper Witwatersrand Supergroup) (Fig. 1.9). The West Rand Group consists mainly of shallow marine to subtidal argillaceous-arenaceous sediments and is subdivided into the Hospital Hill, Government and Je pp estown Subgroups. The sediments of the West Rand Group show depositional cycles, related to fluctuations between fluvial and marine shelf environments, indicative of eustatic sea-level changes (Stanistreet and McCarthy 1991; Frimmel and Minter 2002; Frimmel 2004; Frimmel et al. 2005). One volcanic event is documented (Crown Formation lava) related to the Jeppestown Subgroup and has been dated by U-Pb zircon age determination at 2.974 ± 0.008 Ga (Armstrong et al. 1991). The maximum thickness of West Rand Group sediments is up to 5 150 m in the Klerksdorp area (western part of the Witwatersrand Basin), but thins out towards the northeast where it reaches only 830 m (Frimmel and Minter 2002; Frimmel et al. 2005).

The Central Rand Group lies unconformably on top of the West Rand Group and consists of coarse-grained fluviatile to subtidal arenaceous-rudaceous sediments. This sequence is divided into the Johannesburg and the Turffontein subgroups and reaches a maximum thickness of almost 3000 m (Frimmel and Minter 2002) (Fig. 1.9). Although certain formations in the West Rand Group are also mined (e.g., the upper Bonanza Formation, Frimmel and Minter 2002), the majority of gold mining activity takes place within the horizons of the Central Rand Group. Gold mineralization in the wider Witwatersrand Basin is predominantly found in the conglomerates of the Central Rand Group (Fig. 1.9). Five horizons with gold mineralization are known in the Johannesburg Subgroup (e.g., Frimmel and Minter 2002; Frimmel et al. 2005). The names of each horizon are different from one goldfield to the other. The five horizons are: the Ada May and Beisa reefs; the North, Main, South and Carbon Leader reefs; the Livingstone reef; the Basal, Steyn, Vaal, Saaiplas, and Leader reefs (see Frimmel and Minter 2002; Frimmel 2004; Frimmel et al. 2005). In the area of the Vredefort Dome one goldfield has also attracted attention and is situated near the town of Venterskroon in the northwestern collar (see Fig. 1.6b).

When the deposition of the Witwatersrand succession commenced, a change of the tectonic setting towards a compressional regime took place. This syn-depositional deformation is characterized by variation of stratigraphic thickness and the occurrence of syn-sedimentary folding and faulting (Roering et al. 1990; Myers et
Fig. 1.9: Detailed stratigraphic column of the Witwatersrand Supergroup showing the formation names as well as the main lithologies (after SACS 1986).
al. 1992). Myers et al. (1992) suggested that the Witwatersrand Basin was penetrated by intersecting east-west, left-lateral and north-south, right-lateral fault systems that produce a mosaic pattern of independently moving fault blocks, which controlled both the style of sedimentation and stratigraphic thickness (Fig. 1.10, Table 1.1).

Fig. 1.10: Sketch of the Witwatersrand Basin showing the major synsedimentary faults and associated folds during the deposition of the Witwatersrand sediments (from Myers et al. 1992). Numbers indicate individual fault-bounded blocks as discussed in Myers et al. (1992). North-south and east-west striking faults divide the Witwatersrand Basin into a mosaic-like pattern. The resulting fault blocks controlled the sedimentation within the basin.

1.7.1.3 Ventersdorp Supergroup

The end of sedimentation of the Witwatersrand Supergroup is marked by an up to 3 km thick sequence of tholeiitic flood basalt of the Ventersdorp Supergroup, followed by the deposition of up to 2 km of rift sediments and felsic volcanics. The latter volcanics were dated by U-Pb age determination on zircons by Armstrong et al. (1991) at 2.714 ± 0.008 Ga, an age that is regarded as the end of Witwatersrand sedimentation.
During mid-Ventersdorp times the deformation regime changed back to extensional. In this tectonic environment, already existing reverse (thrust) faults were reactivated and associated half-graben developed (Clendenin et al. 1988; Stanistreet and McCarthy 1991; Myers et al. 1992; Table 1.1). Tinker et al. (2002) suggested that major N and NE trending faults and grabens developed in the Witwatersrand Basin during the Vendersdorp-age rifting. McCarthy et al. (1986) proposed that at least two major faults associated with the Ventersdorp rifting event also affected the strata in the northern collar of the Vredefort Dome (Bank and West Rand Faults; also Fletcher and Reimold 1989). Thickening of the Ventersdorp Supergroup to the west of the Bank Fault suggests that a major component of slip on this fault was probably down-to-the-west normal dip-slip during Ventersdorp rifting. T.S. McCarthy (University of the Witwatersrand, pers. commun. 2004) has also referred to rift-related felsic volcanic rocks in the western collar of the dome. The exact location could not be given, but, according to him, it was where “the Ventersdorp rocks stick up above the Karoo” cover sequence on the southwestern margin of the Vredefort Dome. On all geological maps, these rocks are incorporated into the Klipriviersberg Group (see Fig. 1.9) - the Platberg Group was never differentiated. The significance of T.S. McCarthy’s observation would be that Platberg-type grabens existed in the area of the Vredefort Dome prior to overturning of the strata.

1.7.1.4 Transvaal Supergroup

Deposition of Transvaal Supergroup rocks took place in a shallow sea environment commencing at about 2.6 Ga. Dolomites and subsidiary iron formations of the Chuniespoort Group (Walraven and Martini 1995) were deposited at ~2.557 Ga (Jahn et al. 1990). At the base of this group, a conglomerate bed is present, the Black Reef Quartzite Formation, that contains minor gold mineralization (e.g., Frimmel and Minter 2002). A minimum age of 2.642 ± 0.002 Ga is given for this horizon (Eriksson et al. 1995). Following this, at ~2.35–2.1 Ga, an ~3-km-thick succession of argillaceous-arenaceous sediments of the Pretoria Group (Walraven and Martini 1995) was deposited. Stromatolites are found, in places, in the carbonate rocks of the Transvaal Supergroup, also in the area around the outskirts of the Vredefort Dome.

After the accretion of the Limpopo fragment to the Kaapvaal Craton at ~ 2.6 – 2.55 Ga (Barton and van Reenen 1992), rift basins developed due to the erosion of the
Chapter 1

Limpopo belt and gravitational collapse, thus forming the Transvaal basin (Clendenin et al. 1988; Eriksson and Clendenin 1990; Stanistreet and McCarthy 1991). Erosion over much of the Kaapvaal Craton was responsible for extensive peneplanation (Robb et al. 1997; Frimmel and Minter 2002). Thermal subsidence formed an epeiric sea over the Kaapvaal Craton leading to the thin but widespread sedimentation of the Black Reef Quartzite Formation and the deposition of the Chuniespoort Group sediments (Tankard et al. 1982; Clendenin et al. 1988; Martin et al. 1990; Robb et al., 1997). An erosional unconformity marks a change in the depositional environment with the deposition of the Pretoria Group sequence. Eriksson et al. (1995) suggested that half-graben or fault-controlled basins were associated with the Pretoria Group sedimentation (Table 1.1). They also proposed a closed basin setting with limited marine influence.

1.7.1.5 Post-Transvaal sedimentation and tectonic activity

The Transvaal sedimentation was terminated by the emplacement of - or may have predated by ~ 100 Ma - the 2.06 Ga Bushveld Complex (Walraven et al. 1990), with massive intrusions of ultramafic and mafic magmas. Associated with this magmatism are also voluminous granites and silicic volcanics. At 2.061 ± 0.002 Ga (Buick et al. 2001) felsic volcanic rocks of the Rooiberg Group erupted at almost the same time as a mafic to ultramafic succession of the Rustenberg Layered Sequence intruded the sedimentary rocks of the Witwatersrand Supergroup. Plutons of the Lebowa Granite Suite intruded this sequence at ~ 2.054 ± 0.002 Ga (Walraven and Hattingh 1993). The main remnants of the Bushveld Complex are to the north and northeast of the Witwatersrand Basin. However, it is believed that the Bushveld Complex may not have reached as far as the present-day extent of the Witwatersrand Basin (e.g., Frimmel and Minter 2002). The widespread magmatic intrusion into the lower to middle crust may have caused an increased crustal geotherm, as is evidenced by alkali granite and mafic intrusions into the rocks of the Witwatersrand, Venterdorp and Transvaal supergroups in the area of the Vredefort Dome (Gibson and Wallmach 1995; Gibson and Reimold 2000; Frimmel and Minter 2002; Gibson and Jones 2002). Greenschist to mid-amphibolite facies metamorphism is recorded in the rocks of the Witwatersrand Supergroup in the dome and was dated at ~ 2.078 ± 0.012 Ga (Gibson et al. 2000). However, the Losberg igneous complex, a presumed
Fig. 1.11: Schematic map of the Vredefort Dome showing the intrusive complexes in the area and the distribution and exposure of the Vredefort Granophyre (modified after Gibson and Reimold 2001).
outlier of the Bushveld Igneous Complex, is situated within the Potchefstroom Syncline, just north of the Vredefort Dome, and has been dated at 2.041 ± 0.041 Ga (Coetzee and Kruger, 1989). Other post-Transvaal alkali granite and dioritic plutons are also found in the area of the Vredefort Dome (Fig. 1.11), for which ages similar to the Bushveld Complex have been determined (De Waal et al. 2006; Gibson et al. 2000).

The last major deformation event that affected the Witwatersrand Basin was the impact of a large extraterrestrial body at 2.02 Ga, which resulted in the Vredefort impact structure (Kamo et al. 1996; Reimold and Gibson 1996; Gibson and Reimold 2001).

Table 1.1: Pre-Vredefort deformation

<table>
<thead>
<tr>
<th>Timing</th>
<th>Tectonic environment</th>
<th>Deformation</th>
</tr>
</thead>
<tbody>
<tr>
<td>~ 3.074 Ga (Dominion Group)</td>
<td>extensional, intra-continental rift basin</td>
<td>N to NE-trending faults</td>
</tr>
<tr>
<td>~ 3.0-2.8 Ga (Witwatersrand)</td>
<td>compressional</td>
<td>syn-sedimentary N-S and E-W trending oblique-slip thrust faults, folding</td>
</tr>
<tr>
<td>~2.714 Ga (mid-Ventersdorp)</td>
<td>extensional</td>
<td>reactivation of thrust faults to extensional faults, half-graben development</td>
</tr>
<tr>
<td>~2.6-2.55 Ga (Transvaal)</td>
<td>compressional</td>
<td>SE-verging recumbent folds and associated thrust faults with northerly vergence</td>
</tr>
</tbody>
</table>

1.8 The Vredefort impact structure

The 2.02 Ga (Kamo et al. 1996) Vredefort impact structure is one of the three largest impact structures known on Earth (besides Sudbury and Chicxulub; Grieve and Therriault 2000; for a detailed review see Gibson and Reimold 2001). Its centre, the Vredefort Dome (Figs. 1.6c and 1.12), lies in the centre of the Witwatersrand Basin, itself centred at about 120 km southwest of Johannesburg.

The original size of the impact structure is not well constrained by first-order data. This uncertainty is due to the absence of a well-defined crater rim, which is a result of deep level erosion. However, scaling studies by Therriault et al. (1997), using macroscopic and microscopic shock deformation features occurring in the dome and its wider environs, and geophysical modelling by Henkel and Reimold (1998) have indicated a probable original diameter of ~250-300 km. The argument for the
size is based principally on the distribution of shock deformation features, such as PDFs and shatter cones (e.g., Therriault et al. 1997), and on the original pre-impact thickness of cover strata (e.g., Henkel and Reimold 1998). In contrast, numerical modelling of the Vredefort impact structure (Turtle and Pierazzo 1998; Turtle et al. 2003) resulted in an estimate of the crater diameter of 120-160 km (for further discussion, see also Turtle et al. 2003; Ivanov 2005). This discrepancy is due to the fact that the numerical modelling did not extend to the end of the crater modification phase and that the estimates for the original pre-impact thickness of the cover strata exposed in the collar of the Vredefort Dome are not well constrained. The scaling of the diameter of the Vredefort Structure is also complicated by the oval-shaped, NE-SW elongated shape of the Witwatersrand basin (Fig. 1.6a,b) and the slightly NW-SE elongated shape of the Vredefort Dome. This may be either explained by originally (pre-impact) non-horizontal stratigraphy (Lana et al. 2003a), or by post-impact tectonic activity (McCarthy et al. 1990; Roering et al. 1990; Friese et al. 1995, 2003; Henkel and Reimold 1998).

Between 7 and 10 km of erosion (McCarthy et al. 1990; Gibson et al. 1998) have led to the removal of the actual crater form and its entire impact breccia fill, and have exposed the deeper levels of the structure.

The Vredefort impact structure has a central uplift (the Vredefort Dome) that is surrounded by a rim syncline – the Potchefstroom syncline (Fig. 1.12), which was first described in detail by Simpson (1977, 1978) - and fault zones that are alleged to occur in a concentric pattern (Brink et al. 1997, 2000a,b; Figs. 1.13 and 5.2). The formation of this syncline around the central uplift is believed to be largely responsible for the preservation of the gold-bearing Witwatersrand Supergroup conglomerates that have been mined around the periphery of the basin (McCarthy et al. 1990) for more than 100 years and that have generated some 45% of all gold ever mined in the world (Handley 2004).

The ~80-90 km wide Vredefort Dome is the deeply eroded root zone of the central uplift (Gibson and Reimold 2001). It consists of an ~ 40 km wide core of Archaean basement gneisses enclosed by the so-called collar of subvertical to overturned Late Archaean to Palaeoproterozoic supracrustal strata belonging to the Dominion Group, as well as to the Witwatersrand, Ventersdorp and Transvaal Supergroups. The collar is about 20 km wide and is, in turn, surrounded by the ~ 50
Fig. 1.12: Landsat TM image of the Vredefort impact structure. The central parts of the structure consist of Archaean crystalline basement, and the surrounding collar of supracrustal strata. Also shown are the rim syncline (Potchefstroom Syncline) and the Rand Anticline (modified after Gibson and Reimold 2001). Inset shows the location of the Losberg complex (see section 5.4.5).
km wide Potchefstroom Synclinorium (Figs. 1.12 and 1.14). The dome is well exposed in its northern and northwestern sectors, but is largely hidden beneath shales and dolerite sills of the Phanerozoic (~250 to 180 Ma) Karoo Supergroup in the south and southeast, with the exception of a few inliers of Witwatersrand strata. The Witwatersrand strata, in turn, are surrounded by less well-exposed rocks of the Ventersdorp and Transvaal supergroups (Fig. 1.11).

The collar rocks are exposed at a radial distance of ~20–40 km from the centre of the dome. The Dominion Group strata have been uplifted by at least 15 km relative to the deepest part of the surrounding syncline (Henkel and Reimold 1998).

Fig. 1.13: Interpretative structural map of the area around the Vredefort Dome (after Brink et al. 2000a), showing the main structural features and locations referred to by Brink et al. (2000a) and in the text. Explanation (according to Brink et al. 2000a): (A) Katdoornbosch Witpoortjie Thrust, (D) Foch Sole Thrust (Second Ring), (E) Wuma Sole Thrust, (F) Potchefstroom Synclinorium, (H) Potchefstroom Master Bedding Plane Fault.
The collar rocks are further intruded by a number of felsic (Vredefort Granophyre) and mafic intrusions (Fig. 1.11). The mafic intrusions can be divided into three groups (for a review see Gibson and Reimold 2001): 1) A poorly exposed intrusion with a wehrlite composition has been observed in the central parts of the core, but has not been studied out in detail. 2) Several subhorizontal intrusions of tholeiitic composition (e.g., Anna’s Rust sheet) crop out in the northeastern and central parts of the Vredefort Dome. They are believed to be related to a widespread magmatic event at ~ 1.05 Ga (Pybus 1995; Reimold et al. 2000). 3) Several deformed or undeformed dykes from various parts of the dome with granitic, aplitic or dioritic compositions (Gibson and Reimold 2001), the timing of which is still not well constrained; however, the mafic intrusions show no evidence of impact-related deformation and are therefore inferred to be post-impact.

Fig. 1.14: Schematic map of the Vredefort Dome and environs showing the distribution of the lithologies as well as the main structural features in this area (after McCarthy et al., 1990). The map shows the concentric synclinal structures and the Rand Anticline around the dome as well as identified thrust faults to the north of the dome. The Karoo cover was backstripped.
1.9 Previous work on the Vredefort impact structure

1.9.1 Models for the formation of the Vredefort Dome

The first detailed geological description of the Vredefort area was provided by Molengraaff (1903, 1904, 1905) and Molengraaff and Hall (1924). The first geological map of the Vredefort Dome was made by Nel (1927a and b); this initial mapping (see also Nel 1927a-c; and the related work by Hall and Molengraaff 1925) focused mostly on lithological aspects, although many of the large-scale deformation features (faults, folds) were already identified on Nel’s geological map. The only other geological map available of the region was published by Bisschoff (2000). At the 1:50 000 scale, it is also largely lithological, with barely any new structural data added since the original mapping by Nel (1927a-c). While large (kilometre-scale) radial faults displacing the collar strata are recorded, movement directions along these faults and temporal relationships between the faults and other structures, such as folds, are not presented or discussed in Bisschoff (2000).

Several hypotheses for the origin of the Vredefort structure, both endogenic and exogenic ones, were postulated in the past. While the controversy between a cryptoexplosion (see, e.g., Nicolaysen and Ferguson 1990) and an extraterrestrial origin (e.g., Gibson and Reimold 2001, and references therein) was heavily debated, some other models were proposed but attracted less attention. One of these theories was proposed by Du Toit (1954), who suggested that northerly-directed thrust faults were responsible for the Vredefort Dome, the steep to overturned dips of supracrustal rocks in the northern sectors, and the asymmetric dips of strata in the southeastern sector of the dome (see section 1.9.4). Further theories included the diapiric doming hypothesis of Brock and Pretorius (1964) for the Vredefort Dome basement (see section 1.7). According to these authors, this dome is similar to the other domes in and around the Witwatersrand Basin. A similar theory was introduced by Ramberg (1967, 1981), who suggested that the syncline around and the gravitational high at the centre of the structure related to a marginal sink (syncline) from which a diapiric dome developed. Colliston (1990) presented a model of a NW-verging, SE-dipping, subsurface, ductile shear zone, with its lower boundary at the crust-mantle contact. This northward-directed thrust zone decoupled at lower levels and formed a dome-like, flat-topped, antiformal fold structure (Fig. 1.15a). Furthermore, uplift of this
Fig. 1.15: Endogenic fault-related models for the Vredefort Dome (a) model of compressional tectonics proposed by Colliston (1990). Decoupling at lower crustal levels result in an updoming of the overlying strata, thus, producing an antiformal fold structure (stages 1 and 2). Further rise of the fold structure would then expose different crustal levels as presently visible in the Vredefort Dome (stage 3). (b) Push-up structure model after Coward et al. (1995) that was formed by northwesterly-directed thrusts which are confined to a large strike-slip zone. The interpretation is solely based on seismic data.
structure would then expose the various strata, in the distribution obvious today in the Vredefort Dome. Based on seismic interpretation only, Coward et al. (1995) suggested that the Vredefort Dome was largely a thrust-controlled structure, that formed as a push-up structure on a compressional bend to a large thrust zone (Fig. 1.15b).

1.9.2 Petrographic work

Several studies exist of the shock deformation effects in the Vredefort Dome. Much of this work has been petrographic in nature and has focused on mineral microdeformation features (e.g., Lilly 1978; Martini 1978, 1991; Fricke et al. 1990; Grieve et al. 1990; Reimold 1990; Leroux et al. 1994; Buchanan and Reimold 2002; Gibson and Reimold 2005). Considerable interest has also been shown in the striated cone-shaped fractures that occur primarily in the collar rocks and that are known as shatter cones (Hargraves 1961; Manton 1962, 1965; Albat 1988; Albat and Mayer 1989; Nicolaysen and Reimold 1999). Attention has also focused on the pseudotachylitic breccias in the collar (e.g., Reimold and Colliston 1994; Dressler and Reimold 2004; Reimold and Gibson 2005), but not to the same degree as the work on pseudotachylitic breccias in the core of the dome (e.g., Reimold 1995; Dressler and Reimold 2004). Strong debate remains regarding the timing of the formation of pseudotachylitic breccias within the impact process (Reimold et al. 1992; Reimold 1996; Reimold and Gibson 2005).

The petrographic studies indicate that planar deformation features (PDFs) exist in quartz up to ~35 km from the crater centre (Grieve et al. 1990; Reimold 1990). Although the shock pressures inferred from statistical analysis of these PDFs do not show any consistent increase towards the centre, possibly because of preferential annealing of those PDFs formed at higher shock pressures (e.g., Grieve et al. 1990), recent analysis of deformation textures in the Vredefort Dome (Gibson and Reimold 2005) suggests that pressures in the central parts exceeded those required to produce diaplectic glasses and even feldspar mineral melts, and appear to have reached at least 30-45 GPa. The recognition of coesite and stishovite associated with thin melt breccia veins in upper Witwatersrand Supergroup quartzites in the northern and northeastern parts of the dome led Martini (1978, 1991) to suggest that shock pressure distribution
was heterogeneous on a small-scale. This is supported by Gibson and Reimold (2005), who noted localized enhancement of shock effects in minerals adjacent to thin melt veins in the core of the dome. Based on this evidence, they suggested that the bulk of the pseudotachylitic breccia found in the dome could be shock melt; thus, for now, the non-genetic name “pseudotachylitic breccia” is retained to describe the Vredefort Dome melt rocks, because this origin as shock melt has only been confirmed for a few samples so far, probably because of the extensive annealing that followed the shock pulse. The average shock pressures in the Witwatersrand Supergroup strata were probably in the order of ≤10 GPa based on observations of pre-impact fabrics and textures close to the core/collar contact and the background shock pressures in the currently exposed collar strata (Grieve et al. 1990; Gibson and Reimold 2005).

1.9.3 Metamorphic studies

An additional factor in the investigation of pre-impact tectonics in the Vredefort Dome is that the rocks of the Witwatersrand Supergroup were metamorphosed to lower greenschist- to mid-amphibolite-facies grades only shortly before the impact as a result of elevated geothermal gradients caused by the Bushveld intrusive event (see Gibson and Wallmach 1995 and section 1.7). Another metamorphic event was triggered by the Vredefort impact event. Post-impact metamorphic assemblages suggest that shock heating produced a temperature increase of ~50°C in the outer parts of the dome and of up to 900°C in the core (Gibson and Jones 2002; Gibson and Reimold 2005), which concurs with the shock pressure distribution in the dome based on the shock metamorphic effects (see previous section). Together with the pre-impact geotherm of 15°C/km (Gibson and Jones 2002), the temperature variation and post-shock metamorphic recrystallization are, thus, likely to have obscured many of the older brittle features, and any small displacements or rotations associated with these events are unlikely to be discernable, particularly if they were bedding-parallel (see section 6.2.2.1). Some evidence exists for syn-metamorphic small-scale folding and foliation development in the metapelitic units in the collar (Gibson 1993; Gibson and Reimold 2001, p. 48), and similar features have been described associated with bedding-parallel thrusting from the Witwatersrand goldfields (Phillips and Law 1994; Jolley et al. 2004). Possibly associated with the pre-impact metamorphic event was the emplacement of the alkali granite plutons in the dome (e.g., Frimmel and Minter...
Fig. 1.16: Simplified geological map of the Vredefort Dome showing the main lithologies and structures in the dome, and the eccentric distribution of the pre-impact metamorphic isograd (after Lana et al. 2003a).
Ballooning effects related to this emplacement appear to have triggered significant differential rotation, and possibly even faulting, of the Witwatersrand Supergroup wallrocks in the vicinity of these alkali granites (e.g., Albat 1988, for detailed reviews see Chapters 2 and 3). The anomalous dips of the exposed collar strata in the western part of the collar may also indicate rotations of the strata related to these faults (Pretorius et al. 1986; Albat 1988), although an impact timing of this rotation cannot be completely ruled out. Lana et al. (2003a) proposed that the overall asymmetry of the dips in the collar strata, with normal, outward dips in the southeast and strongly overturned dips in the northwest (Fig. 1.16), together with the eccentric distribution of pre-impact metamorphic isograds, indicates a regional down-to-the-NW tilt prior to the 2.06 Ga Bushveld event, which is regarded as the cause of the metamorphism. These pre-impact metamorphic isograds are eccentric with respect to the collar strata, as they affect progressively younger strata towards the northwestern part of the collar (Bisschoff 1982; Gibson and Wallmach 1995; Lana et al. 2003a; see Figs. 1.16 and 1.17).

Fig. 1.17: (a) Schematic diagram showing Lana et al.’s (2003a) explanation for the eccentric distribution of the metamorphic isograd and the rotated S2 fabric, as discussed in the text, and the NW-tilted core-collar contact in the Vredefort Dome prior to the impact event. (b) Schematic section through the Vredefort Dome demonstrating the asymmetric dip directions in the collar strata after a uniform 90° rotation resulting in overturned strata in the NW and right-way-up in the SE sector) and the symmetric geometry of the metamorphic isograd (after Lana et al. 2003a). See text for further details.
1.9.4 Structural studies in and around the Vredefort impact structure

Ever since extensive gold resources were found in the region of the Vredefort impact structure in the late 19th Century, the main regional geological attention has focused on exploration and mining of the mineral resources, and – thus – geological analysis of the so-called “Golden Arc” of goldfields in the wider environs of the Vredefort Dome (e.g., McCarthy et al. 1990; see Fig. 1.6b). Whilst extensive geological information has been generated by all mining houses operating in the Witwatersrand Basin, also including the occasional excursion into the Vredefort Dome and some geophysical analysis of the entire region (e.g., Durrheim 1986; Durrheim et al. 1986; Antoine et al. 1990), the bulk of these data has remained confidential.

Seismic, magnetic and gravity studies of the central Kaapvaal Craton have produced several interpretations of the regional structural setting of the Vredefort Dome, suggesting that it lies on intersecting NW-SE and NE-SW crustal-scale structures in the Archaean crystalline basement (e.g., Durrheim 1986; Antoine et al. 1990; Corner et al. 1990; Green and Chetty 1990; Nicolaysen 1990). The extent to which these structures are present in the overlying supracrustal sequence exposed in the Vredefort Dome is debatable. Geophysical and borehole data from beneath the Karoo Supergroup cover rocks to the southeast of the dome suggest kilometre-scale displacement of the core-collar contact along at least two faults with northerly to northwesterly trends (Fletcher and Reimold 1989, see Chapter 2), and suboutcrop patterns of the collar strata appear to be highly complex, suggesting either pre-doming erosional unconformities or pre- or post-doming faulting (e.g., Pretorius 1981; Martini 1992; Friese et al. 1995, 2003).

Antoine et al. (1990) emphasized the polygonal shape of the Vredefort Dome and divided the structure into six segments (see Fig. 2.7a). Further geophysical work was done by Durrheim (1986), Durrheim et al. (1986), Hart et al. (1981, 1995), and interpretative work by Friese et al. (1995). Several authors (e.g., Hart et al. 1981, 1995) suggested that during the formation of the Vredefort Dome the entire section of the Archaean crust was exposed (“crust-on-edge model”) and that two major Archaean tectonic boundaries with a NE trend (the so-called “Vredefort Discontinuity” and the “Southeastern Boundary Fault”) separated terranes with different tectonometamorphic histories (Hart and Andreoli 1987; Hart et al. 1987,
1990) that were amalgamated between 3.1 and 2.5 Ga and turned on-end during the formation of the Vredefort Dome (Hart et al. 1990; De Wit et al. 1992).

Previous structural studies in and around the Vredefort Dome were restricted to a few detailed studies of small areas, including the very detailed analysis of parts of the Potchefstroom Synclinorium around the dome by Simpson (1977, 1978; see Chapter 5 and below), and, most recently, the structural analysis of the Archaean Basement Complex in the core of the dome by Lana et al. (2003a-c, 2004). The latter authors dismissed the theory of a Vredefort discontinuity proposed by Hart et al. (1990) and preferred a polyphase tectonometamorphic history for the Archaean basement complex (Lana et al., 2003a). They documented four Meso-Archaean deformation events with the earliest fabric ($S_1$, gneissic foliation) largely transposed by amphibolite- to granulite-facies migmatitic layering ($S_2$ and $S_3$). The $S_4$ fabric is restricted to mylonitic shear zones in the northern and southern parts of the dome. Lana et al. (2003a,b) documented that the pre-impact $S_2$ gneissic fabric was differentially rotated, with the angle of rotation increasing sharply at a distance of ~13–19 km from the centre with subhorizontal $S_2$ orientations in the centre of the core and almost vertical tangential orientations along the core-collar boundary (Fig. 1.16). Based on this orientation pattern, Lana et al. (2003a-c, 2004) proposed a plug-like geometry for the structural features in the core of the Vredefort Dome (Fig. 1.17b). They also reported a lack of distinct megablocks or large-slip-magnitude faults like those described from other central uplifts (e.g., Puchezh-Katunki, Ivanov 1994; Gosses Bluff, Milton et al. 1996) and suggested an accommodation of deformation during the central uplift by small-scale shear and/or rotation along pervasive pseudotachylytic breccia vein-fracture networks.

The meta-sedimentary and meta-volcanic collar strata of the Vredefort Dome have received limited study from a large- to micro-scale structural perspective. Simpson’s (1977, 1978) work on the Potchefstroom Synclinorium did not involve the rocks of the core and the collar of the dome, per se, and, to a large degree, focused on large-scale structural elements such as folds and faults in Transvaal Supergroup strata, which she related to the dome formation (without recognizing an impact origin for the Vredefort Dome) and centrifugal thrusting (see Chapter 5). Fractures in the collar rocks have been studied by Lilly (1978, 1981), Nicolaysen and Reimold (1987, 1999), Albat (1988) and Albat and Mayer (1989). Lilly (1978, 1981) proposed that reverse (core-side-up) dip-slip movement on steep, inward-dipping faults formed part of a
conjugate fault set related to the final stages of settling of the dome. According to him, the other part of this set comprises tangential, shallowly inward-dipping, normal faults, as documented by Simpson (1977, 1978) in Transvaal Supergroup strata along the outer edge of the dome. This latter fault system displays up to 2.3 km of vertical displacement in the northern sector of the dome (see Chapter 5). Similar, inward-dipping, normal dip-slip faults cut the hinges of upright to weakly outward-verging circumferential folds in the syncline (Simpson, 1977, 1978). Apart from the folds, some concentric sinistral strike-slip faults and a few radial faults (Simpson, 1977, 1978), no evidence was recorded on surface of the major outward-verging thrusts proposed by Brink et al. (1997) from reflection seismic data.

Nicolaysen and Reimold (1987), Albat (1988), Albat and Mayer (1989), Martini (1992) and Nicolaysen and Reimold (1999) related some structures observed in the collar of the Vredefort Dome to the formation of shatter cones. Some of this work, as well as that by Manton (1962, 1965), dealt with the relationship of shatter cones to bedding and joints in the collar; in particular, the relationship of shatter cone surfaces and the so-called multipli-striated joint sets (MSJS) was investigated by Nicolayson and Reimold (1999). Detail of this aspect will be reviewed in Chapter 3 dealing with the origin of shatter cones.

Abundant and, in part, massive occurrences of pseudotachylytic breccias along bedding-parallel fault zones in the outer parts of the Vredefort impact structure (especially in goldfield occurrences of the wider Witwatersrand Basin) have been described and discussed by a number of authors (Fletcher and Reimold 1989; Killick and Reimold 1990; Berlenbach and Roering 1992; Killick 1993; Reimold and Colliston 1994; Reimold et al. 1999; Friese et al. 2003). They are principally recognized in the geologically well-studied underground exposures of the goldfields of the Witwatersrand Basin and along some normal faults displaying mainly north-south trends. A number of these occurrences of pseudotachylytic breccia have been dated (Trieloff et al. 1994; Friese et al. 2003) and the results taken to indicate that most of these fault breccias in the wider Witwatersrand Basin are related to the Vredefort event. In addition, however, Berlenbach and Roering (1992) discussed the existence of some pre-Vredefort pseudotachylytic breccias in this region, and Reimold and Colliston (1994) observed several such occurrences in the area of the dome as well. For a detailed review of pseudotachylytic breccias in the Vredefort Dome see Reimold and Gibson (2006).
Another type of melt is also observed in the Vredefort Dome, occurring as vertical dykes, mostly in a concentric pattern at the core-collar contact, but also in four radially trending dykes within the core (Fig. 1.18). Owing to their very homogeneous and unusual composition (Therriault et al. 1997 and references therein; Gibson and Reimold 2005) and because clasts found in the dykes belong to the Witwatersrand Supergroup, these dykes of Vredefort Granophyre are believed to represent gravitationally emplaced fracture fills from the main impact melt body. A recent study by Buchanan and Reimold (2002) found PDFs in quartz grains of a few clasts in the Vredefort Granophyre. Modelling of the Vredefort Granophyre chemistry suggests formation by melting of the supracrustal sequence and the upper crustal granitic basement in the Vredefort Dome (e.g., Reimold et al. 1990; Therriault et al. 1997). Koeberl et al. (1996) found elevated Os contents in the granophyre compared to that of the rocks from which the granophyre derived. Furthermore, an extremely low \(^{187}\text{Re}/^{188}\text{Os}\) ratio confirmed that the granophyre does contain a small (up to 0.2\%) meteoritic component (Koeberl et al. 1996). For a recent review on this issue see Reimold and Gibson (2006).

![Fig. 1.18: Photograph of the Vredefort Granophyre, which mostly occurs as vertical dykes, oriented radially and concentrically to the dome along the core-collar contact, as well as within the core. Width of block, ca. 40 cm (courtesy of R.L. Gibson).](image)
Other structural work includes some macro-structural discussion, largely based on unpublished reflection seismic information, by Brink et al. (1997, 1999, 2000a,b). These authors stated that the impact structure was surrounded by concentrically arranged fold zones and imbricate thrust faults (Foch Thrust Zone and Ensels Fault Zone) (see Fig. 1.13). They postulated that the outward movement caused by the impact went over pre-existing barriers (“ramp structures”, see Brink et al. 2000a, and Figs. 5.3, 5.4). The surface along which this movement occurred followed the subhorizontal Black Reef Décollement Zone (BRDZ), a discontinuity that had been first postulated by Fletcher and Reimold (1989). Thrust faults further outwards would detach from the BRDZ (e.g., the Katdoornbosch Fault, see Brink et al. 2000a) and from concentric folds produced by the impact (Fig. 1.13).

A cross-section through the entire Witwatersrand Basin was presented by Friese et al. (1995), showing inferred concentric thrusts around the Vredefort Dome (see Chapters 2 and 5), similar to those proposed by Brink et al. (2000a,b). However, in contrast to Brink et al. (2000a, b), Friese et al. (1995) related these structures to the post-impact Kibaran event (Fig. 1.19). Furthermore, the nature, displacements or, extent of these alleged large-scale structures beneath the surface are again solely based on reflection seismic data and borehole information that were not presented.

A detailed summary of regional geological events is presented in Chapter 5.

1.10 Methodology

For this project, Landsat and aerial photographic datasets were obtained, which provided a basis for ground-based structural analysis. Remote sensing data sets were also used to determine large-scale structures within the collar rocks, such as folds and major faults. The Landsat images (for details and flight information see Appendix No. 1a) of the area of the Vredefort Dome stem from an aerial survey of the DeBeers Mining Company and had to be georeferenced first (Fig. 1.20). Different colour combinations were tested, and the colour combination TM 321 was found to be the most suitable for identifying large-scale structures and provided the best results for supporting the surface mapping.

Aerial photographs of the Vredefort area at a scale of 1:25 000 and orthophotos at a scale of 1:10 000 were available (for details and flight information see Appendix No. 1b).
Fig. 1.19: Interpretative cross-section through the Witwatersrand Basin from the NW to the SE inferred from seismic data (after Friese et al. 1995), indicating the regional geology and the discussed thrust faults around the collar of the Vredefort Dome. Note that the faults in the SE of the dome also verge to the northwest, indicating that they are not related to the centrifugal displacements caused by the formation of the Vredefort Dome.
The aerial photographs cover mainly the northern parts of the collar, but were crucial to study large-scale structures, as most of them are hidden by dense vegetation and soil and cannot be traced by surface mapping.

The present study involves a structural geological analysis of the northern and western sectors of the collar, mainly in the rocks of the 2.98 - 2.71 Ga (Armstrong et al. 1991) Witwatersrand Supergroup, because of the generally good exposure of these rocks compared to the poorly exposed Ventersdorp and Transvaal Supergroup strata in the outer parts of the collar of the dome.

Initially, the author mapped along three radial traverses across the collar to identify different structures, such as joints, folds, bedding orientation, relationships between shatter cones and pseudotachylitic breccias with other structures (such as joints, folds and faults), and two-dimensional features (e.g., lineations) across different lithological units. The detailed surface mapping was then extended to the whole extent of the Witwatersrand Supergroup in the inner collar in the northeastern, northern and western parts of the dome (Fig. 1.21). Only a few outcrops are available in the southeastern sector, which were also included in the study. Access to these various areas was granted by most of the landowners, but, in a few cases, farmers denied access to their property, and therefore, certain areas could not be mapped. Locally, the dense vegetation and soil cover hampered the continuous mapping, especially – and most unfortunately - along the traces of large-scale faults.

In addition to the investigation around the collar, the areas of the syncline and anticlines in the outer parts of the impact structure between the towns of Potchefstroom, Ventersdorp and Carletonville were also investigated. This was undertaken to compare the nature and origins of impact-related structures in the collar of the Vredefort Dome and the structures in the outer parts. In addition, an attempt was undertaken to find surface evidence for the northerly-directed concentric thrust faults, alleged by Brink et al. (1997, 2000a,b).

The techniques for surface mapping involved measurements of dip directions and dip angles of various structures with a Freiberger geological compass. All the measurements from a single location were then plotted into a stereonet.

Representative samples were taken from many sites and thin sections were made of some of the samples for further measurements of small-scale structures (fracture orientation, lineations, etc.) and microstructural analysis. The total period of
Fig. 1.20: Landsat image with a TM 321 true colour combination of the Vredefort Dome showing the northern parts of the collar and the core. This Landsat image was used additionally to the detailed surface mapping to support and complement the dataset of large-scale structures in the collar rocks of the Vredefort Dome (courtesy of DeBeers Mining Corporation). For details and flight information see Appendix No. 1a.
Fig. 1.21: Schematic map of the exposed collar strata in the Vredefort Dome showing the main lithologies and major faults. The numbered insets (1 to 5) cover the studied areas and are presented in detail with locations in Appendix No. 2 (based on the map by Bisschoff 2000).
time spent in the field was about 10 months stretched over two and a half years from 2002 to 2004.