Chapter 6

Discussion

6.1 Introduction

Previous structural studies of the collar rocks of the Vredefort Dome focused predominantly on specific areas of the collar (e.g., Lilly 1978, 1981) or specific features, such as pseudotachylitic breccias (e.g., Martini 1991; Gibson et al. 1997) and shatter cones (e.g., Manton 1962; Albat and Mayer 1989; Nicolaysen and Reimold 1999). Many of the studies were carried out in the era leading up to the confirmation of the dome as part of an impact structure and, hence, they did not necessarily examine these structures within the context of an impact-induced evolutionary history. This study aimed to document all structural features over the entire exposed extent of the collar of the dome in an attempt to establish geometric and/or temporal relationships between individual structures and their relation to the impact process (impact-induced versus non-impact-induced deformation).

Furthermore, the temporal relationships between impact-related deformation features were investigated in order to obtain a better understanding of the processes involved in the evolution of the deep central parts of a complex impact structure during the cratering process. More specifically, this study aimed at improving the understanding of the processes that take place during the formation and subsequent collapse of the central uplift in such a crater.

A brief summary of the main observations will be provided at the beginning of each section in order to point out the salient facts, which are necessary for a comprehensive discussion of the results with regard to the formation and collapse of the Vredefort Dome, as well as other impact structures in general.

6.2 The deformation of the collar rocks

Large-scale structural analysis of the Witwatersrand Supergroup rocks in the collar of the Vredefort Dome indicates that, rather than exhibiting a comparatively simple geometry involving rigid polygonal segments separated by radial faults as suggested by Antoine et al. (1990), the collar displays a highly heterogeneous internal structure involving both ductile and brittle strain features. The 2.06 Ga metamorphic and alkali intrusive events (Gibson and Wallmach 1995; Gibson and Jones 2002; De Waal et al. 2006) provide a convenient time marker with which to distinguish pre-impact brittle structures from those generated during or after the impact event. The association of the folds, faults and at least some of the fractures in the collar with voluminous pseudotachylitic breccias - that are themselves overprinted by the impact-related thermal event (e.g., Gibson et al. 1997; Gibson et al. 1998; Gibson and Jones 2002) - supports an impact origin for these structures.

6.2.1 Faults and Folds

Folds in the collar rocks of the Vredefort Dome occur as symmetric and asymmetric structures at metre- to hundreds-of-metre scales. Metre-scale folds are usually found along the limbs of large-scale fold structures and display a curtain-type folding style (see Figs. 2.8 and 2.9). The fold hinges of the large-scale folds are radial to oblique with regard to the dome and plunge with shallow to moderate angles towards the dome centre. The overall orientation of the fold axial planes in the collar of the Vredefort Dome is radial with subvertical to steep dip angles. An impact-origin of these folds is in contrast to the findings of some previous studies (e.g., Van der Merwe 1986; Colliston 1990; see section 1.8). For example, Colliston (1990) proposed a non-impact-related formation of these folds based on a model that involves a SW-trending, SE-dipping subsurface ductile shear zone (see Fig. 1.15a). In this scenario, NW-directed thrusting and decoupling at lower levels were responsible for the overturning of the strata and formation of folds. However, Colliston's (1990) model fails to explain the occurrence of fold structures with subvertical axial planes and radially oriented fold hinges, as observed during this study.

Large-scale faults in the collar rocks have radial, tangential and oblique-radial strike orientations with regard to the dome centre. The radial and oblique faults do not seem to have truncated the entire collar. Radial faults show predominantly left-lateral displacements, but oblique faults show both left- and right-lateral displacements. Subhorizontal striations on slickenside surfaces in the fault zones confirm a general strike-slip movement pattern (see sections 2.6 and 2.7). Some authors postulated a pre-impact origin for at least some of the faults in the collar (e.g., Albat 1988). Although pre-impact faults are known to the north of the collar and have been related to the Ventersdorp-age rifting event (e.g., Pretorius et al. 1986; Tinker et al. 2002; see section 2.2), an extension of a W-E striking pre-impact fault west of the collar, near Viljoenskroon, as postulated by Albat (1988), is not supported by this study and recent works (e.g., Holland et al. 1990; Tinker et al. 2002; see section 2.2). Albat (1988) claimed that scissor-type rotation took place along the fault, in order to explain the variation in shatter cone orientation in this part of the collar (see Fig. 2.4). However, the observations during this study demonstrate that there is more than one fault in the western collar where strata were rotated against each other. In fact, highly variable shatter cone orientations are even found on individual samples (see Fig. 3.17 and discussion in section 6.2.5.4). Holland et al. (1990) showed that the consistency in stratigraphic thickness and continuity of strata in the Central Rand Group indicates little tectonic activity prior to the impact. Furthermore, the consistent geometric relationship between radially oriented fold axial planes and large-scale faults with subhorizontal displacements, especially between oblique sinistral faults and sinistral asymmetric folds (see sections 2.5 and 2.6 and Fig. 6.1), in the collar of the Vredefort Dome suggest that their formation is linked. A syn-doming and, thus, syn-impact scenario also concurs broadly with Lilly's (1978) findings, who proposed that primary faulting was caused by shortening of the collar's circumference during dome formation.

Similar geometric patterns of brittle deformation in the inner parts of complex impact structures, as illustrated during this study (see Fig. 2.21), were also documented in other impact structures, such as Gosses Bluff (Milton et al. 1996) and Haughton (Bischoff and Oskierski 1988; Osinski et al. 2005; Osinski and Spray 2005), as discussed in the following sections (sections 6.2.1.2 and 6.2.2.1).

6.2.1.1 Stress pattern responsible for the formation of impact-related folds and faults

Recent studies (e.g., Ivanov et al. 1996; Dence 2004) have suggested that the rocks underneath a central uplift area move as large, massive "blocks" or "slabs" separated by thin shear zones, following the outward-directed movement during the excavation of the transient cavity. This inward-motion is driven by the rise of the central uplift and further by inward-slumping of material from the crater rim, thus producing an inward-converging particle trajectory (also, Melosh 1979; Kenkmann et al. 2000).

Figure 6.1 illustrates schematically the deformation features that are produced during tangential shortening of the collar strata. Inward movement and tangential compression led to the formation of ductile structures (symmetric and asymmetric folds) and brittle structures (faults). Although definite indicators for inward-directed movement are rare in the Vredefort Dome (see section 6.2.1.2), inward movement of material along the radial to oblique strike-slip faults in the order of hundreds of metres is evident from the evaluation of aerial photographs and Landsat images (see Figs. 2.6 and 2.20). The predominantly sinistral radial faults and both sinistral and dextral oblique faults indicate a conjugate fault pattern that can be reconciled with a formation during tangential shortening of the collar strata (Fig. 6.1). In addition, folds are commonly bounded and displaced by large-scale faults, and the hinge zones of these folds are commonly disrupted and displaced by these faults, indicative for a similar time of formation of both folds and faults (see sections 2.5 and 2.8). This geometry indicates a tangential compressional stress (σ_1) during the formation of the folds and faults (Fig. 6.1). These large-scale structures only represent the two-dimensional stress pattern during the formation of the central uplift, i.e., the lateral

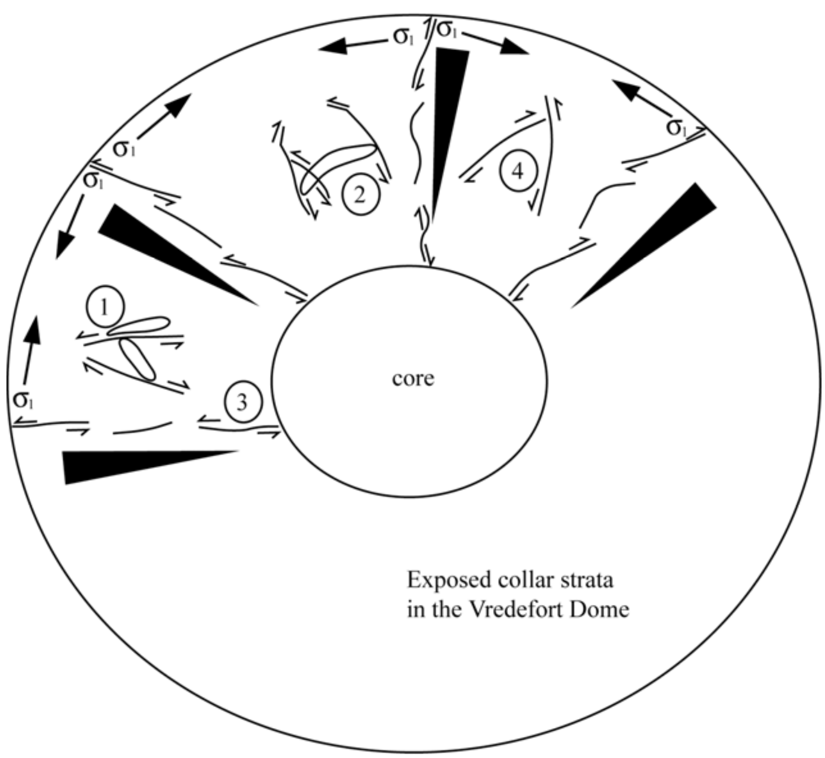

Fig. 6.1: Schematic 2-dimensional illustration of stress and deformation pattern during the formation of the central uplift in the Vredefort Dome. Symmetric (blocks labelled 1) and asymmetric folds (block labelled 2), as well as radial (labelled 3) and conjugate oblique faults (labelled 4), are produced by tangential principal stress (σ_1). After the excavation of the transient cavity, material is driven inwards (thick arrows).

movement as observed in the field. The rise of the central uplift is three-dimensional (vertical movement); however, evidence for this vertical movement, in the form of concentric faults displaying a core-side-up movement, is absent in the collar of the Vredefort Dome (see following section).

6.2.1.2 Kinematic indicators

The radial movements that take place during the rise and subsequent collapse of the central uplift are directed in an opposite sense (inward movement during rise of the central uplift and outward movement during collapse) and change within seconds. The rise of the central uplift is related to upward-directed movement, thus core-side-up movement and development of concentric faults are expected (e.g., Melosh 1989). However, inward-directed movement leads to tangential shortening of material, which produces predominantly tangential and oblique faults (and some radial faults) with predominantly strike-slip displacements (see section 6.2.1.1). Subsequent collapse of the central uplift should then produce movement patterns that display senses that are directed opposite to the previous motion (i.e., downward- and outward-directed). The latter movements may have exploited previous weak zones represented by pre-existing fault surfaces, which provided already a degradation of rock strength. Hence, the initial inward-directed movement is expected to be obliterated to a certain degree (see section 6.2.1.3). Outward-directed collapse structures dominate in the upper parts of central uplifts from other impact sites, such as Gosses Bluff (Milton et al. 1996), Haughton (Bischoff and Oskierski 1988; Osinski et al. 2005; Osinski and Spray 2005) and Upheaval Dome (Kriens et al. 1999; Kenkmann et al. 2005). In Vredefort, the greater depth presently exposed may explain why the main structures appear to be constrictional (this study, see section 6.2.2).

Numerical modelling results by Collins et al. (2004) showing the particle flow and the amounts of total and net strain in complex impact structures (Fig. 6.2) seem to confirm the field observations and the interpretations presented above. The particle flow is greatest in the central uplift area (labelled 1 in Fig. 6.2a) displaying first downward movement during crater excavation and upward movement during the rise of the central uplift. Thus, the total plastic strain, which is defined by Collins et al. (2004) as the strain during initial downward-movement plus the upward-movement, is, as expected, also greatest in the central uplift area. The net strain (Fig. 6.2c) is the resulting strain during both movements, which is less than the total strain owing to the reversal of the sense of motion during the excavation of material and the rise of the central uplift. According to Collins et al. (2004), the particle flow between the central

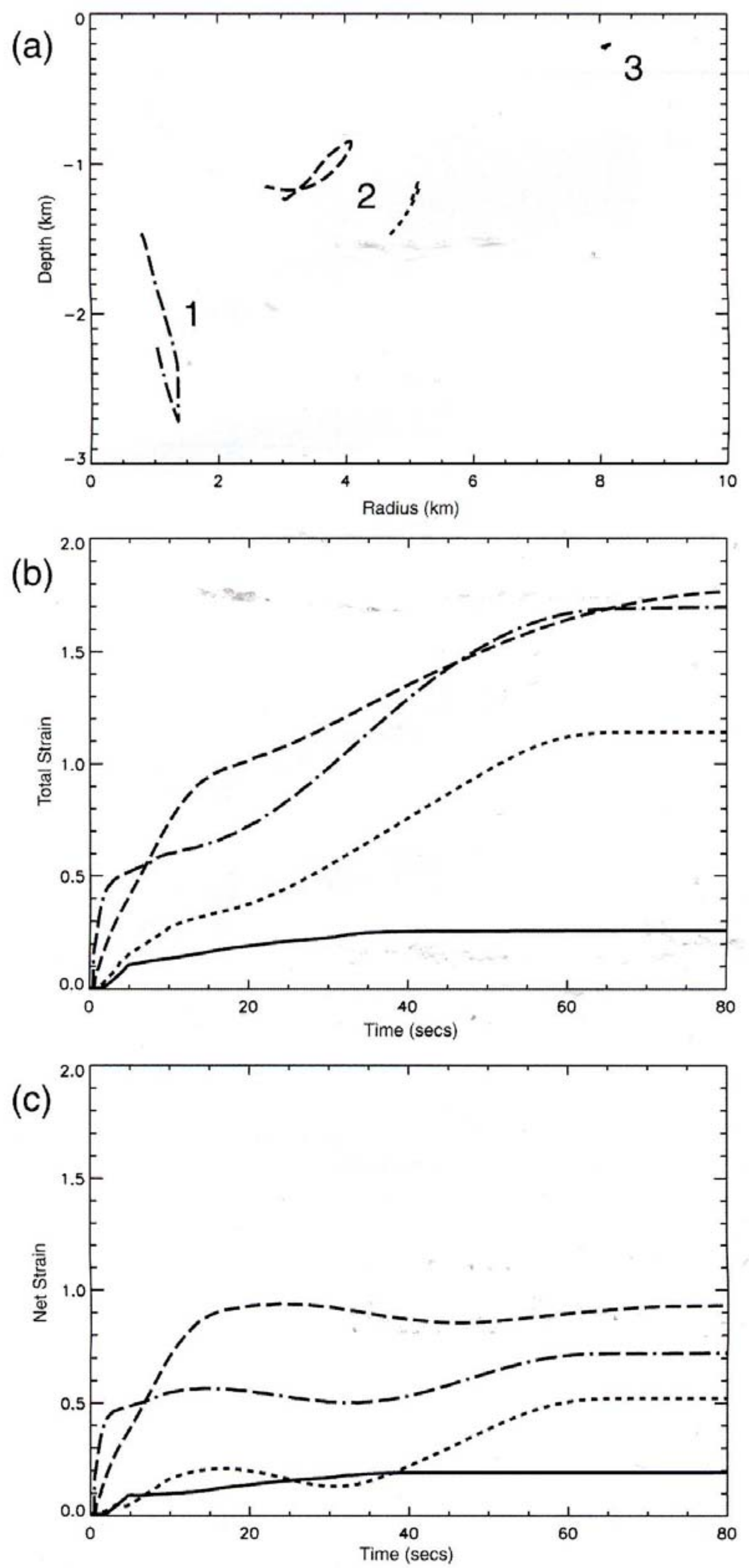

Fig. 6.2: Displacement and strain histories for three different regions of the target: 1) in the central uplift area (stippled line; first downward then upward movement), 2) between central uplift area and crater rim (dashed lines), and 3) beyond the rim (solid line). (a) Particle motions, (b) total plastic strain, (c) net plastic strain. For further detail, see text (Figure from Collins et al. 2004).

uplift area and the crater rim (labelled 2 in Fig. 6.2a) reflects more lateral than vertical movement and the plastic strain is much less than closer to the centre (Fig. 6.2b and c). Although the Collins et al. (2004) model does not extend to the collapse of the central uplift and the particle flows are only given to a depth of 2 to 3 km, these results suggest that the net strain after the collapse is again reduced by the downward movement of material. This may explain the absence of evidence for vertical movement during the initial formation of the central uplift, but can be reconciled with the dominant presence of structures in the central uplift in the collar rocks of the Vredefort Dome that have accommodated the lateral inward movement (tangential shortening, see previous section).

Colliston and Reimold (1990) speculated that weakly defined tangential shear fabrics at the core-collar contact in the northwestern sector of the dome might have been produced during the doming event; they deduced a dextral strike-slip movement sense. Given that both left- and right-lateral movements were observed during this study along tangential faults, the pattern proposed by these authors may imply complex adjustments in response to tangential shortening. Alternatively, these shear fabrics might be attributed to pre-impact (Ventersdorp-age) faults, the existence of which in the Witwatersrand Basin and the northern part of the collar was proposed by several authors (e.g.; Holland et al. 1990; Myers et al. 1990; Tinker et al. 2002; this study – see sections 2.6.1 and 5.4). Lilly (1978) 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" (that would correspond to the collapse of the central uplift). According to him, the other part of this conjugate fault set comprises tangential, shallowly inward-dipping, normal faults, as documented by Simpson (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 of Transvaal Supergroup strata in the northern sector of the dome and separates the gently dipping rim syncline (Potchefstroom Syncline) from the steeply up- to overturned supracrustal strata of the collar (Simpson 1977, 1978; this study). Similar, inward-dipping, normal dip-slip faults cut the hinges of upright- to weakly-outward-verging circumferential folds in the rim syncline (Simpson 1977, 1978). This was confirmed by field observations during this study in the wider environs of the Vredefort Dome (see sections 5.2, 5.3, 5.4). Domeward-dipping normal faults were found in the Potchefstroom Syncline and were also reported in the goldfields of the Witwatersrand

Basin, where they are associated with occurrences of pseudotachylitic breccias (e.g., Fletcher and Reimold 1989; Killick and Reimold 1990; Killick 1993; Reimold and Gibson 2006 and references therein, see section 4.2). In contrast to these normal faults, evidence for outward-directed movement along thrust ramps during excavation of the transient cavity, as proposed by Brink et al. (1997, 2000a,b), was not found during this study, at least on surface (see section 5.4). Given the numerical modelling results (Collins et al. 2004) of similar particle flows of outward movement during excavation and inward movement during central uplift formation (Fig. 6.2), these thrust faults may have been obliterated by normal faulting during central uplift formation (see section 6.2).

The only mesoscopic kinematic indicators that were found during this study in the collar rocks are striations on slickenside surfaces. Slickensides on the shallowly outward-dipping set of fractures (set IV, see section 2.6) indicate a movement of the hangingwall outwards (see Fig. 2.38), whereas slickensides on bedding planes show a core-side-down movement. However, the occurrence of slickenside surfaces is very limited and both sets of slickensides (on bedding and joint surfaces, see section 2.6) were never observed at the same site, thus making it impossible to establish any temporal relationship for inward- and outward-directed movements. The true slip magnitude remains unknown, but is possibly in the range of centimetres to tens of centimetres, as deduced from the offset of marker horizons in shale units (see Fig. 2.38). Although the slickensides may resemble small-scale manifestations of such structures related to the in- or outward-directed movement mentioned above, the scale of movements required in the cratering process is of the order of kilometres to tens of kilometres. In addition, in many places in the collar of the Vredefort Dome, bedding surfaces are either sealed by recrystallization caused by post-shock heating, or by pseudotachylitic breccia veining - thus making an interpretation of the slickenside surfaces in the context of an impact-related movement pattern difficult. The possibility that these small movements were related to post-impact tectonics can also not be completely ruled out.

Similarly, the origin of thrust zones that were locally observed in the collar and the wider environs of the Vredefort Dome (see sections 2.6.1 and 5.4.3) cannot definitely be linked to the impact event. These thrust zones may represent imbricate bedding structures that are responsible for a duplication of bedding, as proposed by several studies (e.g., Grieve and Therriault 2004, and references therein; this study) to occur during the collapse of complex impact structures (see section 6.2.2.3 for a discussion of these structures). Evidence for such thrust zones in central uplifts of complex structures has recently been reported from the Araguainha crater in Brazil (Lana et al. 2005). Alternatively, however, these features could be related to similar structures that were observed in other parts of the Witwatersrand Basin, and which have been linked to the complex pre-impact history of the basin (McCarthy et al. 1986; 1990; Roering et al. 1990; Myers et al. 1990, 1992; Courtnage 1996; see section 5.2).

6.2.1.3 Tangential shortening

The amounts of shortening calculated in this study – about 20% in the lower West Rand Group and less than 10% in the Central Rand Group (see Table 2.1) – are somewhat higher than those deduced by Manton (1965) using Nel's (1927b) original map of the dome (he estimated 7% shortening in the West Rand Group, and no shortening in the Central Rand Group, see section 2.4.1). They confirm the radial outward decrease in the amount of shortening that was also indicated by Simpson (1977, 1978), who suggested that no discernable shortening occurred in the Transvaal Supergroup rocks in the outer collar. The values obtained during this study fall well within the range predicted by numerical modelling of large central uplifts (e.g., Collins et al. 2004; Ivanov 2005). Collins et al. (2004) showed that the reversal in flow patterns from the initial downward and outward excavation flow to the inward- and upward-directed flow during central uplift formation implies that net plastic strains may be significantly lower than the total plastic strains. The total plastic strain, naturally, is much greater than the net plastic strain, as it is the sum of all movements in different directions (see previous section). According to these modelling results (Collins et al. 2004; Ivanov 2005), these movements, and thus deformation, occur in a zone that is restricted to the upper levels of an impact structure (Fig. 6.3) – the so-called damage zone. This damaged zone of material – and, hence, the amount of movement – is greatest closest to the centre of the impact structure (Collins et al. 2004) and decreases with distance and depth from the centre (Fig. 6.3). Taking into account the deep level of erosion in the Vredefort impact structure, it is likely that the level of presently exposed strata in the collar rocks is beyond the zone of maximum damage and, thus, represents a zone where only limited movement occurred.

Furthermore, as demonstrated above, in the central uplift area the dominant movement is probably vertical, which may have led additionally to the low amount of lateral movement (tangential shortening) documented in the collar rocks of the

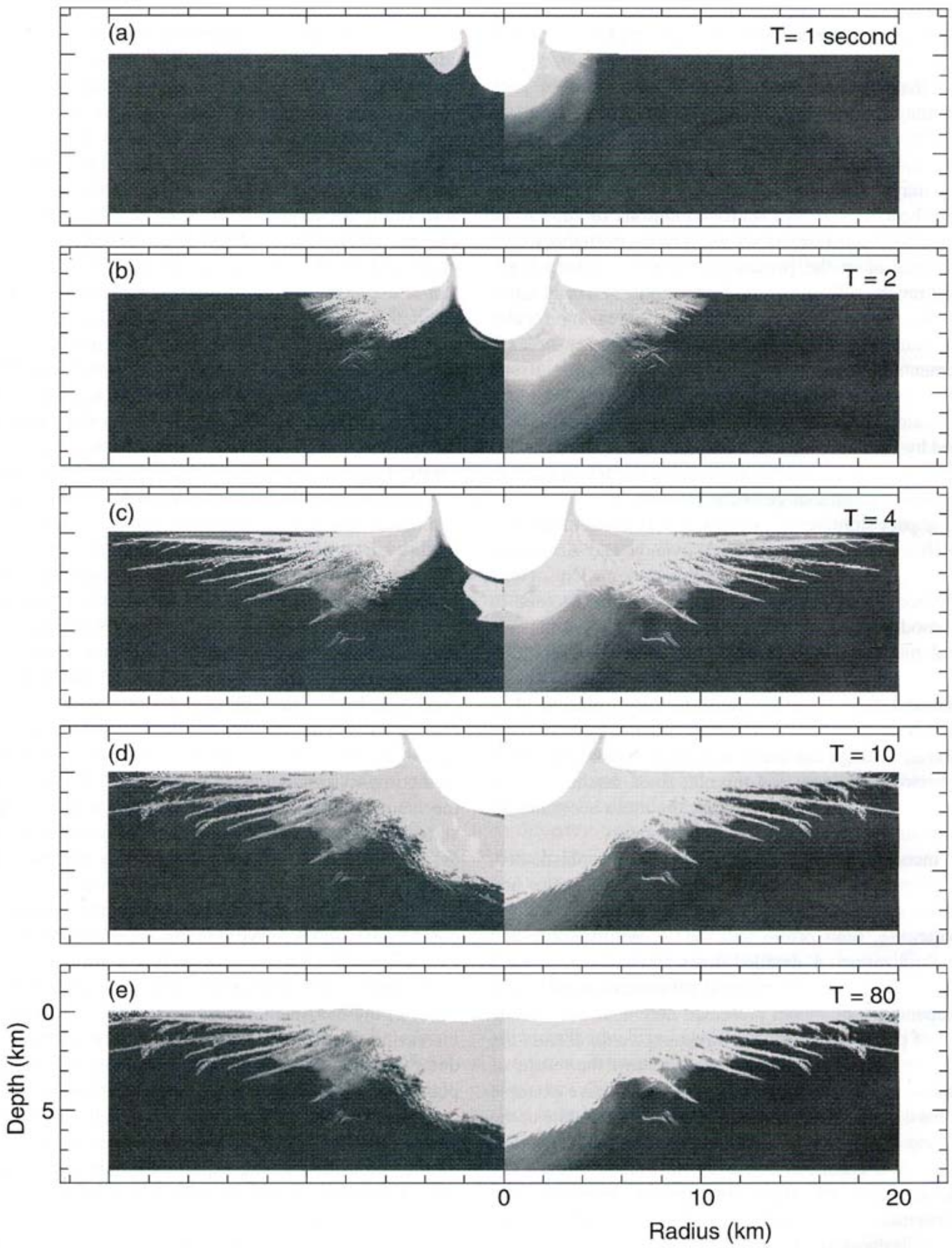

Fig. 6.3: Damage contours for a 10 km-diameter transient crater at several times shortly after the impact event (after Collins et al. 2004). On the left side, the tensile damage is shown, on the right side the combined shear and tensile damage. (a-c) The damaged zone increases with time due to shear failure driven by the passage of the shock wave and the excavation flow and by tensile failure induced by the release wave. (c-d) During the transition between outward excavation flow and the inward collapse flow, further tensile failure occurs as the rise of the central uplift begins. (e) The final damage zone extends well away from the rim and floor, and is highly localized in the distal, brittle region of the target.

large impact structures, such as the Vredefort structure, the inward-sliding material from the walls may not reach the central parts of the structure, in contrast to smaller structures, as proposed by Ivanov (2005). The rise of the central uplift may be followed by an extensional outward collapse of the central uplift, forming a peak ring.

A further possible aspect that is not related to impact tectonics may also have played a role in the determination of the low amount of lateral movement in the collar rocks of the Vredefort Dome. It is known that, prior to the impact event in Ventersdorp times, an extensional stress pattern dominated in the Witwatersrand Basin (e.g., Stanistreet and McCarthy 1986; Clendenin et al. 1988; Myers et al. 1990, 1992; see section 1.7). Similar to the strain pattern and particle flow proposed by Collins et al. (2004; see previous section) for the impact process, the inversion of movements from the extensional regime in pre-impact times to a compressional one during central uplift formation may have partially negated impact-induced shortening and the total amount of lateral movement. However, little is known about how much pre-impact deformation in the Vredefort Dome and its environs might have been obliterated or reactivated during the impact process (see section 2.2).

6.2.2 Implications for central uplift formation and cratering mechanics

6.2.2.1 Formation of the central uplift

The similarity in the stress field responsible for the formation of the folds and faults and their spatial coincidence, together with the similar vergence of faults and asymmetric folds all around the collar of the Vredefort Dome, suggests that formation of the folds and faults overlapped, although ongoing fault-related block rotations may explain the somewhat varied orientation of the folds shown in Figure 2.10 and the generally lesser amounts of overturning of the southwestern limbs of these folds. Whilst fold formation indicates ductile strain, the development of brecciated hinge zones that are commonly pervaded by pseudotachylitic breccias supports a brittle strain component as well, which is consistent with the highly variable strain-rates typical of impact processes. Ductile behavior may have been enhanced by the pre-impact depth setting of the rocks at elevated temperatures and the elevated post-shock temperatures in these rocks, estimated at >500 °C in the inner parts of the collar but decreasing to ~300 °C in the Central Rand Group rocks (e.g., Gibson and Wallmach

1995; Gibson et al. 1998; Gibson and Jones 2002). This radial temperature gradient – the product of differential shock-induced heating and the pre-impact geotherm (Gibson et al. 1998; Gibson and Reimold 2005) – may have played a role in the radial outward decrease in the amount of folding relative to faulting seen in the collar (see section 2.5). However, this pattern may equally be the result of the radial outward decrease in the amount of shortening across the collar (see previous section and Table 2.1), or the change from a mechanically heterogeneous sequence of quartzite and metapelitic units in the inner West Rand Group, which should have favoured a folding response to shortening, to the somewhat more homogeneous and more competent, quartzite-dominated succession in the outer Central Rand Group, which should have favoured faulting.

It does not appear plausible that the steeply radially-plunging folds in the collar of the Vredefort Dome may represent similar structures to those of rotated radial transpression ridges (RTT), which were introduced in the generic model by Kenkmann and von Dalwigk (2000). This is also because the RTT were reported from the outer parts of an impact structure (Kenkmann and von Dalwigk 2000), and it is questionable whether they would be able to extend into the central uplift area. Also, RTTs are a product of major inward movement of wedge-shaped blocks. The formation of the central uplift involves mostly vertical movement. Furthermore, RTTs are shallow features, and, considering the deep level of erosion in the Vredefort Dome, these features would not be present in the currently exposed collar strata.

The predominantly radial to radial-oblique arrangement of structures in the Vredefort Dome, and the dominance of tangential subhorizontal displacements during their formation, contrast with the tangential to oblique-tangential arrangement of centripetally-directed thrusts found in the central uplifts of smaller impact structures such as Gosses Bluff (Milton et al. 1996), Upheaval Dome (Kriens et al. 1999; Kenkmann et al. 2005), Haughton (Bischoff and Oskierski 1988; Osinski et al. 2005; Osisnki and Spray 2005) and Sierra Madera (Wilshire et al. 1972). In the carbonate-hosted Sierra Madera structure (Wilshire et al. 1972, 8 km diameter central uplift) subvertical "curtain-style" folds with radial axial planes around the central uplift indicate a predominantly ductile response to shortening. These vertical folds occur within a zone about one mile wide and indicate constrictional strain; however, no record of any late-stage extensional features has been provided by Wilshire et al. (1972). Results from the Gosses Bluff structure (Milton et al. 1996, 5 km diameter central uplift), where the target comprises a well-layered sandstone succession, indicate a more brittle response during central uplift formation, dominated by concentric thrust and normal faults. Similarly, the Haughton structure (Bischoff and Oskierski 1988; Osinski et al. 2005; Osinski and Spray 2005, 5 km diameter central uplift) displays predominantly brittle deformation, with concentric, but also radial faults, and only minor gentle folding in the limestone and shale strata. Recent observations from field studies from the Haughton impact structure (Osinski et al. 2005; Osinski and Spray 2005) suggest that radial faulting preceded the formation of the concentric listric faults. This temporal relationship could not be confirmed in the Vredefort impact structure during this study, owing to the deep level of erosion. The collar rocks of the Vredefort Dome show an intermediate response, with both ductile and brittle deformation at a variety of scales. The development of ductile structures may also have been enhanced by the longer duration of the modification phase in such a large impact structure (estimated at a maximum of ~15 minutes, Melosh 1989, see section 6.3) and the deep and hotter (shock and residual heat, see above) levels involved in the central uplift (Gibson et al. 1998; Henkel and Reimold 1998; Lana et al. 2003a and b).

The Vredefort Dome exposes the root zone of a complex impact structure (e.g., Gibson and Reimold 2001), whereas these other impact sites compared above represent smaller and more shallowly exhumed structures. Whilst the zone of convergence of the transient crater wall slumps may be preserved in the smaller central uplifts, it is more likely that the centre of the Vredefort impact structure was never marked by such a zone (see Dence 2004; Ivanov 2005). This is because the inward propagation of the faults from the collapsing walls was unlikely to have been sufficiently fast to reach the centre of a very large structure before the floor started to rebound (Dence 2004), or never reaches the central parts (see Ivanov 2005). Deeper levels of exposure also means that the presently exposed strata in the Vredefort Dome reflect more vertical rebound than lateral movement, as discussed for very large impact structures, e.g., by Dence (2004). This suggests that the total amount of centrifugal movement related to central uplift formation and collapse was minor at the present level of exposure in the Vredefort Dome, although, based on numerical simulations, it may have been significantly higher at levels closer to the surface (Ivanov 2005; see Fig. 6.3).

Recent numerical modelling (Collins et al. 2004; Ivanov 2005) confirmed that such structural complexity exists instead in the peak rings of large structures, with the complexity caused by a strong component of centrifugal thrusting, driven by outward collapse of the central uplift, opposing the centripetal motion of the walls. This motion is made possible by the strength reduction of highly fragmented material in the central uplift. A similar approach to explain the movement during the formation of central uplifts and subsequent collapse has already been suggested in the past, involving independently moving megablocks underneath central zones of large impact structures ("acoustic fluidization model", Melosh 1979; Ivanov et al. 1996; see sections 1.5 and 6.2.1.1).

Given the large size of the Vredefort impact structure, the inward-slumping parts of the rims may have converged underneath the rising central parts of the transient cavity, thus enhancing the rebound effect of the innermost parts and leading to a rise of the central uplift above its isostatic stability limit (see Fig. 1.5). As a consequence, the uplifted material collapsed completely without leaving a peak behind in the centre of the structure, as proposed for peak ring structures by, for example, Collins et al. (2004) and Ivanov (2005). Instead, the peak was smoothed or even completely removed by the outward collapse of material (see Fig. 1.2). As a result, the lack of a distinct peak in the Vredefort Dome could be explained by the effects of in- and outward motion in very large impact structures, which is consistent with recent studies (e.g., Dence 2004; Collins et al. 2004; Ivanov 2005) and supported by the steeply up- to overturned collar rocks indicating the outward collapse of the central uplift. However, it has to be pointed out that such interpretations are difficult, given the deep level of erosion in the central parts of the Vredefort Structure.

The presence of large discrete faults in the collar but apparent lack of them in the core of the dome may reflect a combination of increased post-shock temperatures towards the centre of the dome (Gibson et al. 1998; Gibson and Reimold 2005), different rock types (heterogeneous layered collar versus massive crystalline core), and lesser amounts of rotation in the core (plug-like geometry, Lana et al. 2003b). One option is that slip could have been distributed more evenly through the rocks in the core of the dome because of the increase in total pseudotachylitic breccia volumes towards the centre of the dome (Reimold and Colliston 1994; Gibson and Reimold 2005), thereby obviating the need for widely-spaced large-magnitude faults (Lana et al. 2003a, see Fig. 1.17). Melosh (2005) rejected this possibility on the grounds that the melts would have quenched almost immediately after forming, because of the large temperature difference with the host rocks; however, superheated shock melts such as those indicated by Gibson and Reimold (2005) may have been able to survive long enough (all that is required are 2-3 minutes for the formation of the central uplift, Melosh 1989; see section 1.5) given the elevated host-rock temperatures found in the centre of the dome (Gibson et al. 1998; Gibson 2002; see below). This is substantiated by the results of cooling calculations for pseudotachylitic rocks in the Vredefort Dome of this work (see sections 4.4 and 6.2.6). Dence (2004) noted a similar lack of macroscopic fragmentation (i.e., faults) in the central parts of the Charlevoix and Manicouagan central uplifts where shock pressures exceeded 25 GPa, from which he suggested that the amount of brittle deformation in central uplifts may scale inversely with shock pressure. He postulated that this may occur because strength degradation could take place in rocks at such high pressures when rock-forming minerals start to transform into high pressure phases (diaplectic glass, coesite, stishovite).

6.2.2.2 Collapse of the central uplift

Collapse of the Vredefort central uplift is indicated by the radial and tangentially oriented granophyre dykes and by the joints in the collar rocks that developed in a stress field involving simultaneous radial and tangential extension. Impact-melt dyke intrusion is unlikely to have been possible until vertical uplift had ceased, a fact borne out by the lack of any evidence of displacement of the dykes by the folds and faults (e.g., Therriault et al. 1996; Bisschoff 2000; Reimold and Gibson 2006). In all likelihood, the dykes indicate that the central uplift collapsed to the point that it formed a topographic low, allowing at least some of the impact melt to accumulate and be retained above it, from where some could infiltrate along extensional fractures into the central uplift.

Further evidence for outward collapse of the central uplift to form a peak ring may be provided by the increasing amounts of rotation measured from the centre of the dome outwards. Based on structural mapping of the crystalline basement core of the dome, Lana et al. (2003a) concluded that a central region approximately 25 km wide experienced minimal impact-related rotation, and that this is surrounded by an annulus comprising the outer core of the dome, where ~90° of rotation is needed to explain the present orientation of the pre-impact structures in the gneisses. In contrast, the inner collar rocks studied here show an average of ~120° of overturning (see Fig. 2.7 and Bisschoff 2000). There is no support for the suggestion by Lana et al. (2003a) that the overturned dips are asymmetric around the dome in that they reflect only 90° of impact-related rotation superimposed onto a moderately steeply NW-dipping rock sequence. Instead, it is proposed that this increase in rotation of strata reflects the outward collapse of the outer parts of the central uplift in the final stages of crater modification, as demonstrated by recent numerical modelling studies (e.g., Ivanov 2005, Fig. 6.4). The overturning of strata ranges from 90 to 120° in the inner collar of the Vredefort Dome, up to 150° in the outer collar. The shallowly inward-dipping strata in the outer collar would correspond to the almost horizontal grids on the right side of Figure 6.4 up to a distance of 40 to 50 km from the centre.

The cause of the persistent sinistral asymmetry of faults and folds around the collar of the dome is unknown. Pronounced asymmetry in the central uplifts of smaller complex impact structures such as Gosses Bluff and Upheaval Dome has been interpreted as the result of asymmetric mass displacement of rocks close to the surface caused by oblique impact (Milton et al 1996; Kenkmann et al. 2005), but the levels exposed in the Vredefort Dome are far deeper than those studied in these smaller impact structures, making such an option less likely. It is possible that a slight asymmetry in the target rocks – for instance, pre-impact tilting of the supracrustal

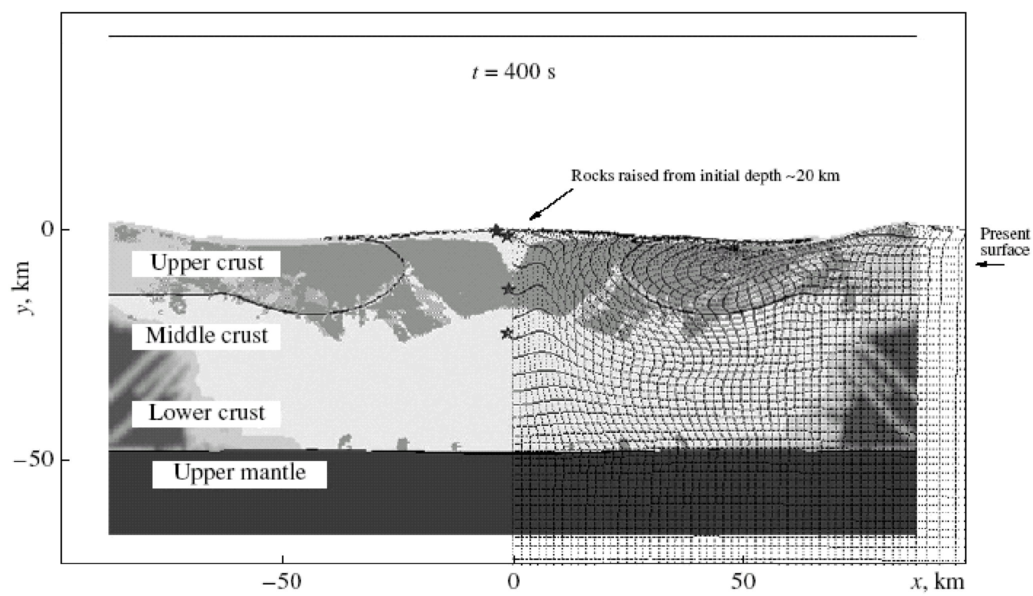

Fig. 6.4: Cross-section through the Vredefort Dome after 400s based on numerical modelling (Ivanov 2005). The dark-grey area below the surface corresponds to the damage zone in the upper level proposed by Collins et al. (2004). The grids on the right hand side demonstrate the increasing overturning of strata with distance from the centre.

succession, but by a much smaller amount than the 30° proposed by Lana et al. (2003a), might induce asymmetry in the impact-related structures. Unfortunately, nearly the entire southern half of the Vredefort Dome is buried beneath younger cover, which hinders further investigation of this problem.

The time needed for the rise of central uplifts in very large structures, such as the Vredefort impact structure (see Chapter 1), has been proposed to be in the range of a few minutes (e.g., Melosh 1989; Henkel and Reimold 1998; Turtle and Pierazzo 1998). However, the question of how long an impact structure takes to develop from the moment of contact to the final stages of the modification phase has not been solved yet. The problem is that it is unknown how long gravitational and isostatic adjustment would take in a large impact structure such as Vredefort, and, thus, a clear definition of the termination of the modification phase is lacking. The timeframe for the rise of the central uplift (2-15 minutes, e.g., Melosh 1989; Henkel and Reimold 1998; Turtle and Pierazzo 1998) for the Vredefort Structure and the formation of the main structures (such as folds, faults and, at least some, fractures) may be adequate; however, the modification phase certainly must have lasted much longer than 15

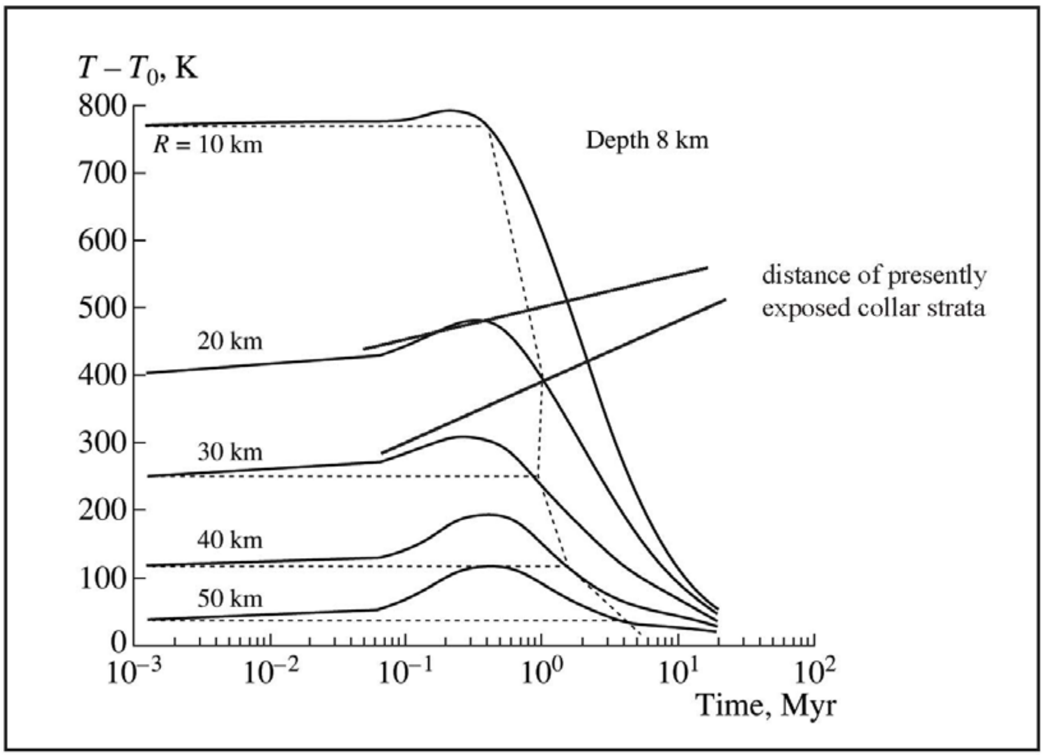

Fig. 6.5: Time variations in the temperature of rocks that were initially at a depth of 8 km after formation of the impact structure (solid lines), as derived from numerical modelling for the Vredefort Dome by Ivanov (2005). The dashed lines correspond to the radial distance from the centre of presently exposed rocks in the Vredefort Dome. The diagram indicates that it took millions of years for the rocks to cool down to the pre-impact thermal gradient.

minutes. Cooling of the post-shock temperatures to pre-impact gradients would have taken several millions of years, as calculations of cooling rates for the Vredefort Dome indicate (Turtle et al. 2003; Ivanov 2005). Ivanov (2005) showed that the rocks at a depth of ca. 8 km (which corresponds to the presently exposed rocks of the collar in the Vredefort Dome) took tens of millions of years to cool down to the temperature of the regional thermal gradient (Fig. 6.5). The small-scale, extensional fractures (see section 2.7 and 6.3.4) also suggest that at least some of these fractures may actually represent possible cooling fractures, or that they were formed due to unloading (drop of pressure) and erosion (see section 6.3.4). As a consequence, the formation of these fractures may also have taken place over the same time span as cooling of the surrounding rocks after the impact.

6.2.2.3 Deformation in the outer parts of the Vredefort impact structure

The existence of map-scale structures that are arranged concentrically to the dome (e.g., major faults, Potchefstroom Syncline, Spitskop Syncline, Rand Anticline) was confirmed during this study and previous works (e.g., Truter 1936; Fletcher and Gay 1972; Simpson 1977, 1978; McCarthy et al. 1990, see section 5.3). At the present level of exposure of the Vredefort Structure, indications of central uplift collapse are provided by the upright to slightly centrifugally-verging folds in the rim syncline that were initially documented by Simpson (1978) and confirmed in this work (Fig. 1.12). However, apart from the folds, some concentric strike-slip faults and a few radial faults, no evidence was documented - on surface - of the major outward-verging thrust zones proposed by Brink et al. (1997, 2000a,b) from reflection seismic data. Brink et al. (1997, 2000a,b) postulated but failed to present evidence on surface and below surface that this alleged outward-directed movement of strata had developed along the limbs of concentrically arranged anticlines. These thrusts were envisaged by Brink et al. (1997, 2000a, b) as having been triggered by the outward movement during the compression phase of impact cratering and turned into extensional faults during the modification stage (see section 5.3). However, the majority of studies in the goldfields and the Potchefstroom Syncline have reported predominantly inward-directed movement along the existing faults (see section 5.3).

This concurs broadly with the field observations of extensional structures obtained during this study from the outer parts of the Vredefort Structure (see section

5.4). In the context of impact-related kinematics in large impact structures (e.g., Dence 2004; Collins et al. 2004; Ivanov 2005), the overriding movement direction is represented by the last movement of material in the cratering process (see section 6.2.1). Thus, inward-directed movement related to the collapse of the crater rim in the final stage of the modification phase is expected to be found in the field. This would mean that the thrust zones proposed by Brink et al. (1997, 2000a,b) related to the early modification phase would then have been inverted to extensional inward-facing normal faults. This scenario is substantiated by field observations of inward-facing (i.e., towards the centre of the dome) faults displaying core-side-down movements in the outer parts of the Vredefort Dome – around the Potchefstroom Syncline (Simpson 1977, 1978) and the Potchefstroom Fault (Van der Merwe 1986; Brink et al. 2000b; this study, see section 5.4.5). Also, occurrences of pseudotachylitic breccias associated with inward-facing faults in the goldfields of the Witwatersrand Basin support the existence of such impact-related structures (e.g., Killick and Reimold 1990; Killick 1990; Killick and Roering 1995, see Chapter 4). Further evidence that the existence of these alleged thrust faults (Brink et al. 1997, 2000a,b) is, at least on surface, questionable, was given indirectly by these authors themselves. They discussed that the Foch Thrust (the second “ring”) joined the Potchefstroom Fault northeast of Potchefstroom. While the Foch Thrust is not exposed, several studies have confirmed the extensional character (i.e., normal fault with east-side-down movement) of the Potchefstroom Fault (e.g., Van der Merwe 1986; Brink et al. 1997; this study: see section 5.4.5). Van der Merwe (1986) postulated a pre-impact formation of this fault. Given the observations of inward-dipping normal faults in the goldfields of the Witwatersrand Basin (e.g., Killick and Reimold 1990; Killick 1990; Killick and Roering 1995) and closer to the collar strata (e.g., Simpson 1977, 1978), a reactivation of pre-impact faults by the impact event is likely. Evidence for this scenario could not be obtained in the field. Taking into account that Brink et al.’s (1997, 2000a,b) interpretation is largely based on data from underground mining activities, boreholes and seismic studies, which are still classified, and that this interpretation is not confirmed by the results of any other studies (e.g., De Kock 1964; Simpson 1977; Engelbrecht et al. 1986), the existence of concentrically arranged, impact-related thrust faults surrounding the Vredefort Dome, must be seriously questioned. Alternatively, these thrust faults, if they exist, could be attributed to tectonic activity in post-impact times (e.g., Friese at al. 1995, see Fig. 1.19).

The general lack of suitable outcrops of structures in the outer parts of the Vredefort Dome does not allow a conclusive assessment of the nature and origin of the observed deformation features. Most of the structural information in the outer parts of the Vredefort Structure stem from underground studies or borehole and geophysical data (e.g., Phillips 1984; Frimmel et al. 1993; Phillips and Law 2000; Jolley et al. 2004). Given the general rarity of surface exposure in the outer parts of the Vredefort Dome and the consistent findings of previous studies (e.g., Lilly 1978; Simpson 1978; Fletcher and Reimold 1989; Killick and Reimold 1990; Killick 1993; Reimold and Colliston 1994; Reimold et al. 1999; Reimold and Gibson 2005, 2006) and this study, the interpretation of outward-directed thrusting by Brink et al. (1997, 2000a, b) seems highly speculative.

In addition, several large vertical faults in the goldfields that strike radial to the dome also appear to have been active, or reactivated, during the impact event (Fletcher and Reimold 1989; Roering et al. 1990; Killick 1993; Reimold and Gibson 2006) and may represent transpressional or transtensional features as postulated by Kenkmann and von Dalwigk (2000) for the outer parts of a complex crater (see section 6.2.2.1). In contrast, possible pre-impact faults of Ventersdorp-age that were identified in the northern collar of the Vredefort Dome do not seem to have been reactivated through the impact event (see section 2.2).

The association of the "chocolate-tablet-type" brecciation (Brink et al. 2000a,b) with these thrust faults cannot be supported by this study. It is confirmed, however, that an unusual type of brecciation occurs within certain chert horizons of the Pretoria Group, as postulated by previous studies (e.g., De Kock 1964; Engelbrecht et al. 1986; Brink et al. 2000a,b). Although these breccias are found, in places, near the traces of normal faults, a regional association of the brecciation with faulting or thrusting in the wider environs of the Vredefort Dome cannot be confirmed due to the lack of suitable surface exposure. This brecciation is also known from carbonate-bearing rocks and was linked to dissolution and/or faulting in extensional tectonic settings in Transvaal Supergroup times (e.g., Berlenbach and Roering 1992; Reimold and Colliston 1994; Reimold and Gibson 2006). The random orientation of clasts within these chert horizons contradicts the formation hypothesis (3-dimensional stress release) of these breccias proposed by Brink et al. (1997, 2000a,b). As a consequence, the terminology "chocolate-tablet-type boudinage", as introduced by

Wegman (1932), is not valid for the occurrences of chert horizons observed during this study (see section 5.4.2).

Lack of clear evidence of impact-related deformation was also found during this study closer to the collar rocks, e.g., along the traverse from Carletonville to Parys (see Fig. 5.1) and around the Losberg Complex (see section 5.4.6). The question why the quartzites of the Magaliesberg Formation around the Losberg Complex do not show any evidence of impact-related deformation, such as shatter cones and pseudotachylitic breccias, remains unsolved, although the fracture pattern is very similar to the fracture pattern in the inner collar rocks. The only difference is that the fracture pattern in the quartzites of the Magaliesberg Formation around the Groot Losberg is less complex than in the collar rocks, regarding density and spacing. This may possibly be explained by the greater distance from the centre of the dome and, consequently, a decrease in intensity of strain, or a change in the stress pattern.

As a conclusion, the question has to be asked whether data from geophysical studies, drill cores and underground studies might solve the controversy about an impact-origin of certain structures in the outer parts of the Vredefort Structure. Given the poor surface exposure of these structures, these additional data are crucial and need to be accessible for scientific work.

6.2.3 Joints and shear fractures

Based on the field observations of joint geometries in the collar rocks of the Vredefort Dome, at least two generations of joints can be distinguished. One set is characterized by closely-spaced sets of fractures that may be, in places, curviplanar, and, generally, do not exceed a few tens of centimetres in length and show generally only submillimetre dilations. No evidence of macroscopically discernable displacement could be found along these fractures, but they show a strong relationship with striated surfaces related to shatter coning (see section 6.2.5). The fact that metapelitic units at the core-collar boundary show less intense small-scale jointing, in contrast to some units in the outer parts of the collar, may be explained by the fact that the impact-induced metamorphism produced higher temperatures in the inner parts, which led to annealing of at least some of these structures, whereas further away from the core, the temperature was lower and these small-scale joints were preserved. This type of jointing is equivalent to the MSJS phenomenon of Nicolaysen and Reimold (1999).

The second generation of fractures typically shows a straight and open pattern and can be followed over metres to decametres. These dilational fractures in the collar rocks of the Vredefort Dome are indicative of a reversal of the stress pattern from a compression, responsible for the formation of folds and faults (see section 6.2.1.1), into an extensional environment. Their character indicates a formation that is consistent with radial and tangential extension. The spacing varies from a cm- to metre-scale, and the fracture sets include the five sets (labelled I to IV plus bedding-parallel set) discussed in section 2.6. All these fractures crosscut but rarely displace each other, suggesting a similar (or identical) time of formation. These fractures are largely planar and lack striations. The fact that these fractures cut through shatter cones and related fractures (see section 2.7) suggests a formation after the shatter coning (see Chapter 3). It seems that vertical and shallowly outward-dipping fractures dominate in the central parts of the exposed quartzite ridges, whereas the conjugate fracture set (labelled II and III, see section 2.7) appears to be predominantly present towards the ends of those ridges, which are typically crossed by large-scale faults (see Fig. 2.33). This might indicate that the conjugate set of fractures formed with the faults during the initial formation of the central uplift. However, their crosscutting relationship with the orthogonal fractures (labelled I and IV, see section 2.7) and their extensional character suggest that they were formed or reactivated during the collapse of the central uplift (see section 6.2.2.2). The radial and shallowly outward-dipping fractures were likely formed during the same phase of the cratering process, considering the movement pattern and orientation of the main stress (σ_1 is vertical) during the collapse of the central uplift (see section 6.2.2.2).

In addition, subvertical and bedding-parallel fractures also crosscut pseudotachylitic breccia veins. Given an impact-related formation for these veins (see Chapter 4), the extensional fractures in the collar rocks of the Vredefort Dome were probably formed between the formation of the large-scale folds and the solidification of the pseudotachylitic breccia veins. The pseudotachylitic breccias with thicknesses >10 cm solidified very late in or even long after the cratering process ended (possibly up to years, see section 4.4.1). Thus, no conclusive answer can be given whether these fractures were formed directly by the impact event - after the formation of folds and faults during the formation of the central uplift and during the subsequent collapse or gravitational adjustment - or in post-impact times due to pressure decrease and unloading by erosion (which might still be related to the modification phase, see discussion in section 6.2.2.2).

6.2.4 Shatter cones

6.2.4.1 Distribution and occurrence of shatter cones in the Vredefort Dome

Striated surfaces are abundant in the collar rocks of the Vredefort Dome. They were found as far as 65 km from the centre of the dome (see Fig. 3.9). Shatter cones are preferentially developed on bedding or small-scale, closely-spaced fracture surfaces, but show a complex pattern with regard to their orientation, morphology and striation geometry.

The grain size in the shatter cone-affected lithologies is obviously a major factor in shatter cone formation, as already noted by previous workers (e.g., Dietz 1959; Manton 1962, 1965). The abundance of shatter cones is clearly enhanced in finer-grained rocks. In contrast, the size of shatter cones does not seemingly depend on the grain size of the host lithology: shatter cones with lengths of tens of centimetres were observed both in fine-grained shale as well as in exposures of medium- to coarse-grained quartzite, and even the largest cones, of 80-100 cm diameter, were formed in shale as well as quartzite. Another parameter that may play a role regarding the size of cones could be layer thickness in the propagation direction of the shock wave. Thinner layers are more likely to produce closely-spaced MSJ in a manner analogous to the controls on spacing of normal joints (see Davis and Reynolds 1996). In addition, narrow-spaced cleavage/foliation/cross-bedding within the layers can also cause scattering/reflection of the shock wave and, thus, may have some effect on local abundance and size of shatter cones. Such a cleavage exists, for example, at the Booysens Shale location on farm Rooderand 26 (location 2 in Fig. 3.9).

6.2.4.2 Shatter cone morphology

The variation in shatter cone forms observed in this study also confirms the findings of some earlier studies (e.g., Manton 1965; Milton et al. 1996; Nicolaysen and Reimold 1999;) and seemingly is a function of the fracture surface (curved or planar) from which the striations develop. Nicolaysen and Reimold (1999) postulated that striations in fact represent the traces of crosscutting MSJ. Both Manton (1965) and Nicolaysen and Reimold (1999) described diverging striations as well as subparallel trends from fracture-related occurrences. According to Nicolaysen and Reimold (1999), the curvature observed on the apex or on basal planes of a cone feature corresponds to a combination of striation segments belonging to several sets of MSJ. They proposed that there are multiple, closely-spaced fracture sets with distinct orientations that intersect, and that the striations represent intersections of linked segments of such fractures. This is in agreement with the observations made during this study of locally curved closely-spaced fractures. Therefore, it is proposed that striations that range from divergent on the top of a segment to a curved pattern at the bottom (Fig. 3.2), either represent a combination of two MSJ surfaces or may be related to interaction of curved fractures belonging to more than one fracture set.

In the case of paraboloid surfaces (which can be approximated by a cone in some cases), the apparent apical angle seems to be relatively consistent at a single site, confirming the observations of Manton (1965) and Albat (1988) from the Vredefort Dome and of Osinski and Spray (2005) from the Haughton Structure. Apical angles range from 90 to 120° from site to site (see Manton 1965; Albat 1988), and only rare exceptions display larger angles. Baratoux and Melosh (2003) suggested that the apical angles of shatter cones, and, thus, the form of a shatter cone surface, are controlled by: (1) the intersection of the magnitude of the extensional and spherical scattered wave and (2) the properties of the heterogeneity, in particular the sound velocity ratio between this heterogeneity and the surrounding material. Taking into account that the target sequence exposed in the collar of the Vredefort Dome is lithologically very heterogeneous and, thus, would have provided a range of possible heterogeneities of different size and nature (lithological boundaries, fractures with variable orientations, grain size, mineral cleavage or lineation, etc.), a highly variable shape of the reflected/scattered wave is expected, which could also explain the variable apical angles observed.

6.2.4.3 Striation geometries

The “typically” diverging pattern of striations on a shatter cone surface has been reported by many authors (e.g. French 1998, and references therein). Sagy et al.

(2004) termed this appearance a branching fracture network and postulated that it is a typical characteristic for shatter cones. However, the current study has shown that their postulation cannot generally be applied to the shatter cones of the Vredefort Dome: diverging striations are, indeed, common, occurring on typically partial conical fracture surfaces and branching off the apex. Locally, however, on almost flat striated surfaces, the striations are subparallel to parallel to each other, making it difficult to impossible to determine the orientation of a cone apex.

The results of the statistical evaluation of striation angles on a single shatter cone segment are in strong contrast to the conclusions of Sagy et al. (2002, 2004), with regard to an increasing mean angle with radial distance from the centre. Their results, however, show a range of striation angles from about 20° to almost 45º on individual shatter cone samples, similar to the spread of values for individual samples determined in this study (see Fig. 3.7). Owing to the irregular distribution of heterogeneities mentioned above, it is proposed that a statistical evaluation cannot be applied to determine a mean striation angle, as all angles on a single segment have to be considered. However, even if applied, the results of this investigation demonstrate that the (meaningless) mean values of striation angles from samples of one rock type (in this case quartzite units of the Witwatersrand Supergroup) do not increase with distance from the centre of the Vredefort impact structure (Fig. 3.24). In contrast, they show a highly irregular trend for sites located throughout the collar, regardless of the distance from the centre. This may again be explained by the fact that the widths of striation angles do not depend on the intensity of the front wave (Sagy et al. 2002, 2004), but mostly on the characteristics of heterogeneities in the host rock (Baratoux and Melosh 2003).

A variable shape of the reflected/scattered wave as proposed by Baratoux and Melosh (2003) may also contribute to the variety of striation angles observed on shatter cone surfaces. Observations made during this study demonstrate that the ridges that form these striation angles do not represent single striations, but are formed by clusters of striations; each ridge represents a small-scale shatter cone itself (see Fig. 3.22).

The statistical evaluation of angle widths during this study shows that the mean angle seems to be narrower in very-fine grained rocks, such as the shales at location 2 (see Fig. 3.23c), and fairly consistent in rocks with similar grain size (e.g., greenstones and fine-grained quartzites, see Figs. 3.23a and b).

6.2.4.4 Shatter cone orientation

Many authors (e.g., Dietz 1961; Hargraves 1961; Manton 1962, 1965; Albat 1988) postulated a more uniform orientation pattern of shatter cone apices on a local scale in the Vredefort Dome than was observed during this study. The idea of simple back-rotation of strata around their strike axes to a pre-impact position, as proposed by these authors, does not result in a focal point-source for the shatter cones evaluated in this study. For instance, shatter cones with apex orientations parallel to the strike of the bedding would not rotate towards the centre of the dome by simple back-rotation into likely pre-impact orientations, considering that in most places the bedding strikes tangentially to the collar.

Prime examples are the famous Schoemansdrift location in the western part of the collar (location 680, see Appendix No. 6), and the Booysens Shale location on farm Rooderand (location 2, see Fig. 3.9) in the northwestern sector, where most (but not all) apex orientations show a westerly or easterly trend, but the bedding dips to the southeast. A non-horizontal pre-impact bedding orientation does also not resolve the problem, as the range of apex orientations at a single site means that a uniform pre-overturning apex is still not attainable. Complex pre-impact structure cannot be invoked either, as simple back-rotation should still show a consistent behaviour for measurements from a given site (at least over an outcrop extent of several metres). Complex post-impact faulting/folding deformation on a macro-scale can also be excluded, for the same reason.

It can be confirmed that, at many sites, the most prominent direction of cone apices in the collar strata is approximately normal to the strike of the bedding, with apices lying on (or parallel to) the bedding surface. However, apex orientation varies, and within small areas apices pointing both up and down can be encountered. If applying back-rotation, this would result in an inward or outward orientation of the apices. These observations do not surprise, as bedding plane-parallel discontinuities (fractures, lithological boundaries, graded bedding, cross-bedding, etc.) most likely provided the most important pre-impact heterogeneities in the sedimentary target rocks of the Vredefort collar that could be exploited by the shock wave during cone formation. However, it has to be emphasized that there are also other orientations, parallel and oblique to the strike of the bedding, which have been largely disregarded in the past (e.g., Fig. 3.19a-c), with a few exceptions (e.g., Manton 1965, and Nicolaysen and Reimold 1999).

Albat and Mayer (1989) concluded that at least some parts of the collar strata had been differentially rotated across major faults prior to the impact event and that, as a consequence, the angle between cone axes and bedding varied across these faults. These parts lie in the western and southwestern sectors of the collar. Field observations during this study confirm that megablocks (hundreds of metres up to kilometres in size) with differential rotations do exist in the collar of the dome (see Fig. 2.24). Some evidence of pre-impact normal faulting has been found, but in most cases the related rotation appears to have occurred during formation and subsequent collapse of the central uplift, i.e., after the shatter cone formation (see section 6.3). The diverse orientation pattern of shatter cone apices, however, can be found at a scale of a few metres, or locally even within a few centimetres, and can, therefore, not be linked to the rotation of these megablocks (see Figs. 3.17a,b). Small-scale (at the metre-scale or less) brecciation/fragmentation has only been observed at a handful of sites, and the detailed shatter cone observations during this study were made at locations where such concerns are seemingly irrelevant. Given the regional history, a pre-impact origin of the faults along these megablocks cannot be completely ruled out (see section 2.3 and 6.2.1). Some of them could have existed and would have been reactivated by the impact event, although a significant thermo-metamorphic event at ~2.06 Ga (Bushveld magmatic event, Walraven et al. 1990) may have sealed some of these faults so that they were unlikely to reactivate.

The observations during this study of fracture-related striated surfaces indicate that the variety of shatter cone orientations can best be reconciled with the different orientations of pre-existing heterogeneities in the target rocks, from which a shock wave is scattered and/or reflected. Scattering/reflection of the shock wave at heterogeneities with different orientations (bedding and fracture surfaces) may explain the range of different shatter cone orientations on individual samples of more than 90° to each other (compare Figs. 3.17a and b) and at an entire outcrop of up to 180° to each other (compare Figs. 3.19a – c). This range of orientations of shatter cone apices concurs with the findings from other impact sites, such as, e.g., Sierra Madera (Wilshire et al. 1972) and Haughton (Osinski and Spray 2005).

The data of apex orientations presented in section 3.4.4 do not support the traditional “master cone” concept, as they obviously do not lie on individual small circles on the stereonets for a specific locality (Fig. 3.19a, b). A crude circular pattern incorporating the bulk of the data in Fig. 3.19b, with a steeply southward-plunging “master cone”, might be discernable; however, the apical angle of this “master cone” would not fit the 90-110° range claimed to be the norm by previous workers (e.g., Manton 1965; Albat 1988). Combining the striation orientations in Fig. 3.19a would suggest a crude master-cone with an apex orientation that is roughly vertical and at more than 45° to the bedding, and an apical angle that would approach 180°. The data presented in this study form a representative set of measurements for this site. The results, thus, concur with Nicolaysen and Reimold (1999) that, as previous studies did not use all the available data from a given locality, the “master cone” concept must be questioned.

What remains to be tested in the field, however, is whether the existence of master cones may depend on the presence or absence of rock layering. In the very heterogeneous sedimentary layers of the Witwatersrand Supergroup, where most of the shatter cones were observed, the master cone hypothesis is clearly not valid. In the more massive layering of the crystalline basement and the Ventersdorp lavas in the outer parts of the collar shatter cones are only rarely present, and, thus, could not be included in the investigation of the existence of master cones.

6.2.4.5 Relationships of shatter cones to other deformation features

Since the first recognition of conical striated fractures nearly a century ago, shatter cones have been extended to include a wide range of striated fracture geometries, ranging from partial cones to almost planar surfaces. There has been much debate in recent years about the relationship between shatter cones and a variety of other fracture surfaces believed to result from impact, such as the “shatter cleavage” from the Gosses Bluff structure (Milton et al. 1996) and the “S-fractures” (Albat and Mayer 1989) and the multipli-striated joints (MSJ; Nicolaysen and Reimold 1987, 1999) from the Vredefort structure.

This study confirms the intriguing relationship between striated surfaces and certain types of fractures of locally variable, planar to curviplanar geometry. The surfaces occur in a parallel to subparallel pattern at spacings typically on a millimetre- to centimetre scale (see Fig. 3.3). These pervasive fractures can best be reconciled with the MSJS described by Nicolaysen and Reimold (1999), as shown in section 3.4

(see Fig. 3.3). This parallel, narrowly-spaced pattern does not appear directly comparable to the so-called "shatter cleavage" from the Gosses Bluff structure, which is said to show a more rhomboidal geometry (Milton et al. 1996). Milton et al. (1996) also stated that the most typical appearance involves segments of cones, but that flat surfaces with almost parallel striations are also observed, and the shatter cones seem to be transitional with the so-called "shatter cleavage". The "S-fractures" resemble a conjugate set of fractures (Albat and Mayer 1989) that were regarded by Nicolaysen and Reimold (1999) as just two sub-sets of the full bundle of MSJ orientations found at any given site. These two sets of closely-spaced, small-scale fractures typically show an orthogonal orientation with regard to the bedding and are limited to individual stratigraphic layers. Furthermore, the lack of dilation along these fractures might indicate a compressional environment. Given the postulation of an impact-timing for these fractures by several authors (e.g., Nicolaysen and Reimold 1999, see above), these fractures most likely formed by the propagation of the shock wave through the target rocks.

Some information regarding relationships between shatter cones and other deformation features, such as fractures, faulting and folding, and pseudotachylitic breccias, was reported from the Vredefort structure previously (by e.g., Simpson 1981; Albat and Mayer 1989; Reimold and Colliston 1994; Nicolaysen and Reimold 1999). Simpson (1981) stated that striated surfaces were especially concentrated on joint planes or faults and suggested that shatter cones formed after faulting and folding. In the course of this study, however, no such relationship of shatter cones could be found; no shatter cones were found concentrated on or near a major fault or fold, and no relationship to pseudotachylitic breccia was observed either. Based on the temporal relationship between these structures, established in Chapters 2 and 3, it is possible that later tectonic activities during the cratering process (central uplift formation, collapse of central uplift) might have obliterated or overprinted striated surfaces. Alternatively, Simpson (1981) may have mistaken some striated surfaces for shatter cones. Striations similar to those belonging to the shatter cone phenomenon are known from normal tectonic processes, such as percussion marks, slickensides etc., and could be mistaken for shatter cone striations (a detailed comparison of striated surfaces – impact-related and non-impact-related – was provided by e.g., Reimold and Minnitt 1996 and Lugli et al. 2005). Whilst the small-scale, narrowly-spaced sets of fractures are closely related to the shatter cone occurrences (see above), another type of fracture with a penetrative pattern and lengths of tens of centimetres to metres cross-cuts the closely-spaced fracture sets and shatter cones. This study (see section 3.4) showed that these penetrative fractures are present throughout the collar and that they represent dilational joints (see section 6.2.4). As demonstrated in the section on shatter cone orientations, the cross-cutting tensile fractures suggest a formation *after* the shatter cone formation, either in the modification phase, late in the impact cratering process, during the collapse of the central uplift, or in post-impact times due to unloading of rocks (see section 6.3.4). The curviplanar, closely-spaced fracture systems (or MSJS) have striations on the fracture surfaces and, accordingly, are considered to be related to the conical shatter cone phenomenon. In contrast, the large-scale pervasive tensile fractures show a lack of such striations on their surfaces.

A crosscutting relationship between shatter cones and pseudotachylitic breccia veins, such as the case reported by Reimold and Colliston (1994), was not observed in this study. However, many pseudotachylitic breccia veins display striated margins, and in some cases, pseudotachylitic breccia veins in basement granitoid are cut by centimetre-spaced fractures with striated walls (see Chapter 3). These could well be less densely spaced equivalents to MSJ or even shear fractures that were formed late in the cratering process (modification phase). It was noted in thin sections that closely-spaced fractures meander around grains. Consequently, they can be much closer spaced in fine-grained rocks (shales) than in coarser-grained arenites. No distinct relationship between Martini's (1991) thin (his so-called A-type pseudotachylite) melt breccia veinlets from the Kromellenboog locality in the northeastern part of the collar and shatter cones could be determined in the course of this field work. It can be confirmed, though, that some shatter cone surfaces carry a thin surface veneer of glassy melt, as already described by e.g., Gay (1976), Gay et al. (1978), Martini (1991), Gibson and Spray (1998) and Nicolaysen and Reimold (1999). These melt films were only observed on some samples, and exclusively from quartzite horizons in the collar. According to Gibson and Spray (1998) melt coats and spherules on shatter cones from the Sudbury Structure consist predominantly of silicate material. They concluded that the spherules were produced by "*localized melting and/or vaporization of the shatter cone walls by a combination of friction and shock*" (p. 333, Gibson and Spray 1998). Friction involvement was supported by Nicolaysen and Reimold (1999), who reported melt at intersections of MSJ, together with micro-displacements. Micro-displacements along these fractures can be substantiated by this study (compare Figs. 3.16a and b). Thus, the thin melt films on shatter cone surfaces (e.g., Martini 1991; Gibson and Spray 1998; Nicolaysen and Reimold 1999; this study) indicate that slip was involved during shatter cone formation.

6.2.4.6 Genesis of shatter cones

The formation of shatter cones is a complex issue and the hypotheses for their formation presented in the past seem to explain only certain aspects of the origin of shatter cones.

Important aspects of the shatter cone phenomenon include the orientation of shatter cone apices parallel, normal and oblique to the bedding, the geometry of the characteristic striations (subparallel to divergent) and variable forms of shatter cone surfaces (planar to paraboloid). The distribution and occurrence of shatter cones seem to depend also on the rock type (massive vs. heterogeneously layered). Also important in the context of the formation of shatter cones is the relationship of striated surfaces to closely-spaced sets of fractures (MSJS), as well as micro-displacements along shatter cone surfaces described by previous workers and the presence of a thin film of melt on at least some surfaces.

Sagy et al. (2002, 2004) and Baratoux and Melosh (2003) followed the idea of reflection or scattering of a rapidly moving wave at inhomogeneities in the target rock (such as textural or structural [joints, bedding planes] heterogeneities) or change in lithology (e.g., grain size, degree of recrystallization, or mineral content). Whilst Sagy et al. (2002, 2004) suggested a constant intensity of the front wave at a given distance from the centre of the structure, and, consequently, consistent apical angles and mean striation angles over this distance, Baratoux and Melosh (2003) proposed that the main wave is often reflected and scattered due to rock heterogeneities and that the intensity of the reflected wave depends on the speed difference of this wave between the heterogeneity and the surrounding material. Apical angles from 90° to 110° for shatter cones in the Vredefort Dome were reported previously (e.g., Manton, 1962, 1965; Albat 1988; this study) - with rare exceptions (see section 3.4.2). The 20° variation may be well explained by a difference in the wave speed as proposed by Baratoux and Melosh (2003).

The observations of variable striation angles ("V-angles", Sagy et al. 2004) in the collar rocks of the Vredefort Dome during this study demonstrate that these striation angles can be explained by the front wave and the regular decay of the main shock front. In contrast, the range of striation angles from individual samples and *in situ* shatter cone segments measured during this study (from tight to open, even on the same shatter cone segment, see Fig. 3.22), as well as the irregular pattern of mean striation angles with distance from the centre of the dome (see Fig. 3.24), imply necessarily that the intensity of the stress wave varies locally, as expressed in the front wave hypothesis. These variations could simply result from the existence of a reflected and scattered wave as postulated by Baratoux and Melosh (2003). The direction in which this wave is reflected/scattered at these heterogeneities cannot be predicted, as it also depends on the shape and nature of these inhomogeneites (see above). This may explain the different orientations of shatter cones at one specific location.

Whilst the Sagy et al. (2002, 2004) postulation of a decrease of mean striation angles with distance from the centre of the structure sounds plausible for the cratering mechanism during an impact event (e.g., Melosh 1989), this postulation is in contrast to field observations during this study (see Fig. 3.24) and some previous studies (e.g., Nicolaysen and Reimold 1999), and can only be applied in combination with a dependency of this reflected wave on the characteristics of heterogeneities in the target rocks (Baratoux and Melosh 2003, see above).

The reflection of a wave at heterogeneities can also explain why shatter cones are best developed in fine-grained rocks. The finer the grain size, the more heterogeneities are present in the rocks. This includes, naturally, more grains than in more massive rocks, but also the existence of possible cleavages and foliations.

Both hypotheses (Sagy et al. 2002, 2004; Baratoux and Melosh 2003) and the MSJS/shatter cone relationship postulated by Nicolaysen and Reimold (1999) have in common that heterogeneities in the target rock are required in order for the rapidly moving wave to be reflected/scattered. The different shapes of these heterogeneities can explain the range of shatter cone surfaces, from planar to paraboloid/hyperboloid, as observed by many workers (e.g., Nicolaysen and Reimold 1999; Sagy et al. 2004; this study). The association of striated surfaces with closely-spaced sets of fractures with variable orientations is strongly supported by this study. This strong relationship implies that the shatter cones must have formed coevally with, or after, the formation of the structural inhomogeneites (fractures), in order for them to be exploited by a rapidly moving wave. The striations might in fact represent fracture traces, as proposed by Nicolaysen and Reimold (1999).

Some workers (e.g., Sagy et al. 2002, 2004; this study, see Chapter 3) have suggested that shatter cones represent tensile fractures formed in an environment of decompression, best reconciled with the time immediately after the passage of the shock wave (e.g., Baratoux and Melosh 2003). This is substantiated by the presence of thin coats of melt on some shatter cone surfaces (e.g., Gay 1976; Gay et al. 1978; Simpson 1981; Gibson and Spray 1998; Nicolaysen and Reimold 1999) which - at least partially – could have formed by a drop in pressure rather than by friction. This is in contrast with most of the previous models which postulated that shatter cones are formed during shock compression (e.g., French 1998 and references therein). Notably Nicolaysen and Reimold (1999) also observed displacements along some MSJ, at some intersections of which they also found glass. This seems to imply that frictional melting must have contributed to the formation of melt on shatter cone surfaces. During recent field work in the Haughton impact structure, Osinski and Spray (2005) observed shatter cones in crater-fill breccia deposits. These authors concluded that shatter cones form in the very early stages of the cratering process (contact/compression phase), without specifying whether their observations can be reconciled with a formation during shock compression or shock decompression.

As a result, the new observations of shatter cones in the collar rocks of the Vredefort Dome by this study demonstrate the complex nature of shatter cone formation. None of the current hypotheses can fully explain the variety of shatter cone orientation, surface morphology, striation geometry and striation angles. Therefore, a combination of aspects of the recent works seems to provide the best solution to explain most of the characteristics of shatter cones. These aspects, which are based on field observations during this study and supported by previous studies (e.g., Manton 1962, 1965; Milton et al. 1996; Nicolaysen and Reimold 1999), include:

1. Shatter cones are tensile fractures and are preferentially associated with closely-spaced sets of fractures (MSJS). This would mean that closely-spaced fractures and shatter cones were not formed during shock compression, as widely postulated in the past, but immediately after the passage of the shock wave, by the interference of the scattered

elastic wave and the tensional hoop stress that develops behind the shock front (see Baratoux and Melosh 2003).

2. Scattering/reflection of the shock wave at heterogeneities in the target rock is able to explain the variety of shatter cone shapes, the range of striation geometries and striation angles, as well as the diversity of apex orientations observed during this study and by other workers (e.g., Manton 1965; Wilshire et al. 1972; Nicolaysen and Reimold 1999; Osinski and Spray 2005).

On the other hand, these observations contradict certain concepts that have been proposed in the past. This includes the “master cone” concept (e.g., by Albat and Mayer 1989) and a relationship between the size of striation angles and the distance of the shatter cone location from the centre of the impact structure (Sagy et al. 2002, 2004). Whilst the database concerning some aspects of shatter cone formation (such as apex orientation, surface morphology and striation geometry) has been substantially increased in recent years, unsolved problems (such as the reflection/scattering of a fast propagating shock wave and the question of when and under what conditions during crater formation and evolution shatter cones form) require further attention.

6.2.5 Pseudotachylitic breccias

Pseudotachylitic breccias are found throughout the collar of the Vredefort Dome, with their occurrence and volumes decreasing with increasing distance from the centre of the dome. The fact that massive pseudotachylitic breccia occurrences are present in the core of the Vredefort Dome (e.g., Lana et al. 2003a; Dressler and Reimold 2004) but not in the collar (see section 4.3) may potentially be explained by a combination of the different target rocks (massive crystalline basement in the core vs. heterogeneous sedimentary layers in the collar), different pre-impact temperature, the different average shock pressures (up to >45 GPa in the core vs. ~ 10 GPa in the inner collar, Gibson and Reimold 2005) and different post-shock temperatures (up to 700-1400 °C in the core vs. 500-300 °C in the collar, Gibson et al. 1998; Gibson 2002; Gibson and Reimold 2005).

The pseudotachylitic breccias in the collar occur as veins with parallel and oblique orientation to the bedding, typically up to a few centimetres in width, and as more voluminous veins and pods in fold hinges and in the vicinity of large-scale faults. Although the veins in the latter structures do not show preferred orientations, they show a consistent spatial relationship with these large-scale structures, where they seem to have filled extensional fractures (compare Figs. 4.5, 4.6, 4.7).

The exact mechanism(s) by which the breccias formed remain(s) problematic (see review in Reimold and Gibson 2005). Three mechanisms have been proposed: (a) localised shock melting caused by extreme fluctuations in shock pressure, (b) friction melting along slip surfaces triggered during the modification stage of cratering, or (c) a combination of shock and friction melting during the shock stage.

Considering a shock origin of all pseudotachylitic breccia melts in the Vredefort Dome, the cooling calculations suggest that most of the melts crystallized prior to the onset of the modification phase (central uplift formation) and that only melts with more than 500-600°C of superheating *and* exceeding widths of 1 cm may have been able to survive in a molten state into the modification phase (see section 4.4). Given the liquidus temperatures of quartzite and metapelitic melts, this scenario would be more likely for melts with metapelitic compositions.

Evidence exists for localised enhancement of shock deformation against fractures hosting narrow melt breccia veins in both the core (Gibson and Reimold 2005) and collar (Martini 1991) of the dome, but these fractures also invariably show small amounts of displacement. Given their small volumes, it is likely that these melt breccias crystallized or quenched virtually instantaneously (see section 4.4), indicating a syn-shock formation of these fractures in a shock scenario for these melts. 1 mm wide melt breccias with less superheating may have crystallized at the same time as the fractures, superheated veins of the same thickness may have stayed molten a few milliseconds longer, however, crystallization most likely took place directly after formation of these fractures. Small-scale displacements along thin bedding-parallel veins are quite common in the collar rocks and indicate that at least some friction was involved during their formation or, alternatively, that melt along fractures may have lubricated them, thus, allowing later frictionless slip during the modification phase. Melosh (2005) showed that mm-scale displacements are sufficient to produce thin melts, and, given the cooling results, a friction generation of, at least some, thin superheated melts cannot be completely ruled out. Alternatively, some of the thin veinlets may have originally been wider and lost their volume by discharge of the melt (see below) or the slip postdates the melting.

Although the high degree of superheating can be achieved by shock melting, the question is whether these high initial melt temperatures are reasonable for melts with metapelitic compositions. Unfortunately, this questions remains unsolved. Shock melting, however, is a function of the shock pressure which is itself probably linked to the background shock pressure. The problem is that the background shock pressure in the collar rocks is in the order of 2-10 GPa (e.g., Gibson and Reimold 2005) and even doubling the pressure would not achieve the pressure needed to melt the rocks in the collar (in contrast to the shock pressure in the core of the Vredefort Dome; see Gibson and Reimold 2005).

If the assumption is made that all melts were formed by friction, then it is likely that melts with thicknesses > 1cm and superheating of several hundred degrees could survive from the shock stage to intrude dilational sites during the modification phase. However, thinner veins or veins with less superheating would have to have formed during the modification stage, in order to intrude sites linked to central uplift formation.

The question in a friction scenario is whether frictional melting is able to produce the required heat for the high degrees of superheating as calculated in section 4.4 for veins exceeding widths of 1 cm. This is because frictional melting may be self-limiting with regard to the maximum melt temperature, first as the melting of individual minerals will buffer the internal temperature. Second, the melting of minerals via this mechanism will also lead to lubrication of the fault surface until such time as the melt can be expelled; such lubrication obviously prevents further frictional sliding.

However, it has to be pointed out again that all these calculations were based on several assumptions, some of which may influence cooling rates dramatically, e.g., clast content and further heat production by continuous friction. The latter might become important in the scenario of removal of a melt into zones of lower strain rate (Melosh 2005, see above). If enough melt is extracted, then the width of the fracture zone is small enough to support further friction and, hence, further heat and melt production. Therefore, it is possible that some of the thinner (1 mm, 1 cm) veins in the Vredefort Dome may have originally been much wider than presently observed due to this migration effect, and, consequently, the results presented above are only estimates and the numbers for the cooling calculations only represent maxima.

As a result, more voluminous superheated melts may have been able to retain a molten matrix for several hours, if not days (see section 4.4). In contrast to the narrow veinlets, which typically show a close correspondence between their matrix composition and that of their wallrocks (Dressler and Reimold 2004), the larger breccias show abundant evidence of mixing of melt from a variety of sources, as well as exotic clasts (Reimold and Colliston 1994), indicating sufficient time for melts and clasts to move distances of at least metres to tens of metres. In the context of the cratering process, it is, thus, plausible that the thick (10 cm) superheated melts could all have formed simultaneously during the shock stage by either shock melting or shock + friction melting, but that crystallisation and final emplacement straddled the subsequent stages of central uplift formation and collapse, depending on the degree of superheating of the melts, the host rock temperature and the volume of melt present. Equally, however, some of the melts could have formed during the modification phase in response to high-strain-rate slip events along large faults. Although pseudotachylitic breccias that are associated with large-scale faults are known from the goldfields in the Witwatersrand Basin (Killick and Reimold 1990; Killick 1993; Phillips and Law 1994), large-scale fault structures that are required for such slip magnitudes (in the order of several kilometres) have not been found yet in the Witwatersrand Basin and are also absent in the collar (see section 6.2.1) and the core of the Vredefort Dome (Lana et al. 2003a). Furthermore, in the case of friction-generated melts, recent studies (Melosh 2005) proposed that, the thicker the melt, the more likely it becomes that the friction sliding process shuts down, as the two sliding surfaces are not in contact anymore. Although the cooling calculations (see section 4.4) confirm that relatively thick superheated melts (≥10 cm) can stay molten long enough to migrate for hours, it seems unlikely that these melts, which require very large faults (hundreds of metres to kilometres) for their formation, can escape from the sliding surface fast enough in order to prevent shutting down the friction process. This may depend primarily on the viscosity of the melts, but, even if the melts are liquid enough, it is questionable whether they would be able to discharge in time, given the extremely short time for such movements in the cratering process (probably in the order of seconds, Melosh 1989; Turtle and Pierazzo 1998). As a consequence, a friction origin for thick (≥10 cm) superheated melts in the collar and in the core of the Vredefort Dome is seemingly not supported. Instead, the observations seem to support a shock-origin for the bulk of the thicker pseudotachylitic breccias in the Vredefort

Dome. Alternatively, however, a combination of both shock and friction melting is also possible, as postulated by several authors and deduced from shock experiments (e.g., Kenkmann et al. 2000; Langenhorst et al. 2002). These authors suggested that, at low shock pressures (~10 GPa), the shock heat is not sufficient to produce melts, but high enough for frictional shear heating and they called melts formed by this process "shock-induced frictional melts".

It is worth noting that thinner breccias in the collar of the dome are more likely displaced by fractures than thicker breccias (compare Figs. 4.10b and 4.13) and that a melt capable of surviving for even a few minutes within such a complex structural environment may now reside in a very different structural context from the one in which it originated. Thus, the preferential occurrence of voluminous pseudotachylitic breccias in zones of structural complexity (such as fold hinges and faults, see above) may indicate that shock-induced pseudotachylitic breccias were injected into these sites during the formation of the central uplift where they then functioned as lubricants and enhanced slip along the radial and oblique faults (see displacements of fold hinges and the collar strata, sections 2.5 and 2.6). The results of the cooling calculations for thick superheated melts (see section 4.4) indicate that it is plausible for thick veins to have outlasted all impact-induced deformation in a molten state during the cratering process, thus allowing the melts to intrude dilational sites during and after the latter stages of the impact event.

As a conclusion, the results suggest that low amounts of superheating would not enhance melt mobility and longevity and that melts were not mobile enough to survive into the modification phase. High superheating means that all veins could be either shock-induced and survive into the modification phase or are friction-generated.

A combination of shock-induced and friction-generated melts is also possible, as suggested by some recent studies (Kenkmann et al. 2000; Langenhorst et al. 2002; see section 6.2.6 for detailed discussion). The key point for the calculations is the degree of superheating of the melts with different compositions. Unfortunately, the absolute melt temperatures for veins in the Vredefort Dome are not known. Although the issue of shock- versus friction-generated melts remains inconclusive, the first-order cooling calculation performed during this study give an idea of the problematics of melt formation and cooling times and might hopefully encourage further studies.

6.3 Chronology of impact-related deformation

Based on the geometric and temporal relationships of and between individual structures in the collar of the Vredefort Dome, a chronology for the development of these structures during the formation of the Vredefort impact structure can be given, which is summarized in Figure 6.6. This figure illustrates the sequence of impact-related deformation based on the observations of this study in comparison with findings from other impact sites (e.g., Sierra Madera, Wilshire et al. 1972; Gosses Bluff, Milton et al. 1996; Araguainha, Bischoff and Prinz 1994, Lana et al. 2005; Upheaval Dome, Kriens et al. 1999; Kenkmann et al. 2005). The sequence proposed ranges from the shock compression stage (Fig. 6.6a) to the modification stage, which is divided into the rise of the central uplift (Fig. 6.6b), collapse of the central uplift (Fig. 6.6c), and final gravitational adjustment (Fig. 6.6d). As already stated in Chapter 1, it is important to note that there is no sharp boundary between these stages and overlap or transition is very likely.

6.3.1 Contact/compression phase

The oldest mesoscopic impact-related structures seen in the collar rocks of the Vredefort Dome are closely-spaced sets of fractures (MSJS) and shatter cones (section 2.7 and Chapter 3) that, together with at least some of the pseudotachylitic breccia (Martini 1991; Gibson and Reimold 2005), formed during the shock pulse (Fig. 6.7). However, based on the geometric relationship between these fractures and the shatter cones (Nicolaysen and Reimold 1999; this study), the "compression stage" must be divided into a compression (shock wave) and decompression phase (rarefaction wave). Shatter cones reflect aspects of tensile fractures (Sagy et al. 2004). Together with the observed micro-displacements along these fractures (Nicolaysen and Reimold 1999; this study) and the presence of thin melt coats on shatter cone surfaces (Gay 1976; Gay et al. 1978; Simpson 1981; Gibson and Spray 1998; Nicolaysen and Reimold 1999), the observations of shatter cones during this study suggest a formation in a decompressional environment, preferentially along the abovementioned fractures (compare Figs. 3.3 and 3.19a-c). The formation of shock-induced melt breccias is substantiated by this study and concurs with findings of

Fig. 6.6: Chronology of impact-related deformation based on the observations during this study and on the findings from other impact structures (see text for further details). (a) Shock compression stage preceding the excavation phase: development of closely-spaced sets of fractures (MSJS) and shatter cones and shock-induced pseudotachylitic breccias. (b) Formation of the central uplift: formation of folds and large-scale faults. The movement pattern is shown by arrows. (c) Collapse of the central uplift: formation of, at least some, shear fractures and joints; the resulting movements are given by arrows. And (d), gravitational adjustment: formation of inward-dipping, listric faults in the outer parts of the structure, and of extensional fractures. The origin of pseudotachylitic breccias (shock-related or friction melt) cannot be linked to a specific stage of crater development. Furthermore, joints related to the final stage may also have been reactivated in post-impact times.

previous workers (Martini 1978, 1991; Gibson et al. 2002; Dressler and Reimold 2004; Gibson and Reimold 2005).

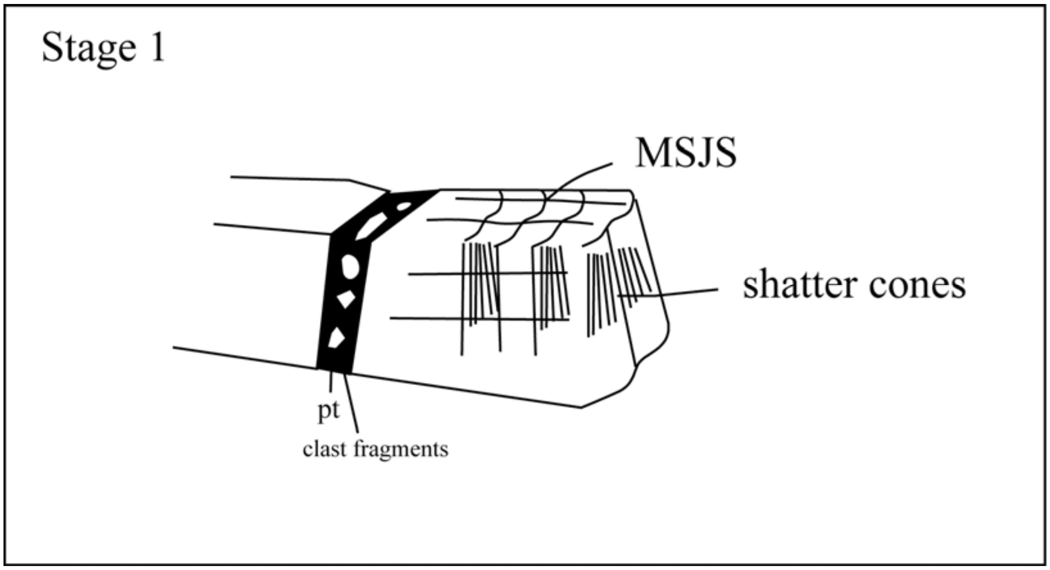

Fig. 6.7: Simplified illustration of the impact-related deformation during the shock stage of crater development. During this stage shock-related deformation features, such as shock fractures (MSJS) and shatter cones were formed. Formation of pseudotachylitic breccias (shock melts, labelled pt) during this stage is also proposed (see text).

6.3.2 Formation of central uplift

The next impact-related structures to form were the folds with/without faults. Recent modelling by Collins et al. (2004) suggests that rocks in the central uplift follow a trajectory in which they are initially displaced downward and radially outward under shock compression, followed by strong vertical uplift. Only towards the end of this do they move closer to the centre of the impact structure than the position from which they started (see discussion in section 6.2.1 and Fig. 6.2). This suggests that the folds most likely developed close to the end of the rise of the central uplift, although the different bedding orientations on the limbs of some folds (see section 2.5) may reflect differential rotation of fault-bounded blocks during the subsequent outward collapse of the central uplift. Continuing tangential shortening was further accommodated by asymmetric to conjugate strike-slip faulting (see section 2.6). Pre-existing faults do not appear to have been particularly reactivated (see section 2.2), possibly because almost all pre-impact faults are likely to have been annealed by the pre-impact metamorphism (e.g., Gibson and Wallmach 1995; section 2.2) or because of unfavourable orientations of pre-impact faults (Holland et al. 1990;

Myers et al. 1990; 1992; Tinker et al. 2002). The rise of the central uplift may also have been accommodated by bedding parallel slip. Although evidence for such movement was not found during this study in the collar rocks, some authors have in the past suggested a core-side-up movement along bedding-parallel surfaces (Lilly 1978, 1981).

It is possible that some pseudotachylitic breccias could have formed at this stage through frictional melting along the faults; however, it is more likely that sufficiently voluminous melts created during the shock pulse could have survived long enough to be driven into extensional sites opening within these younger structures**,** where they either quenched or crystallized (Fig. 6.8). Given that the central uplift probably formed within only minutes (e.g., Ivanov and Deutsch 1999), strain-rates for the tangential shortening were probably of the order of at least 10^{-3} to 10^{0}.

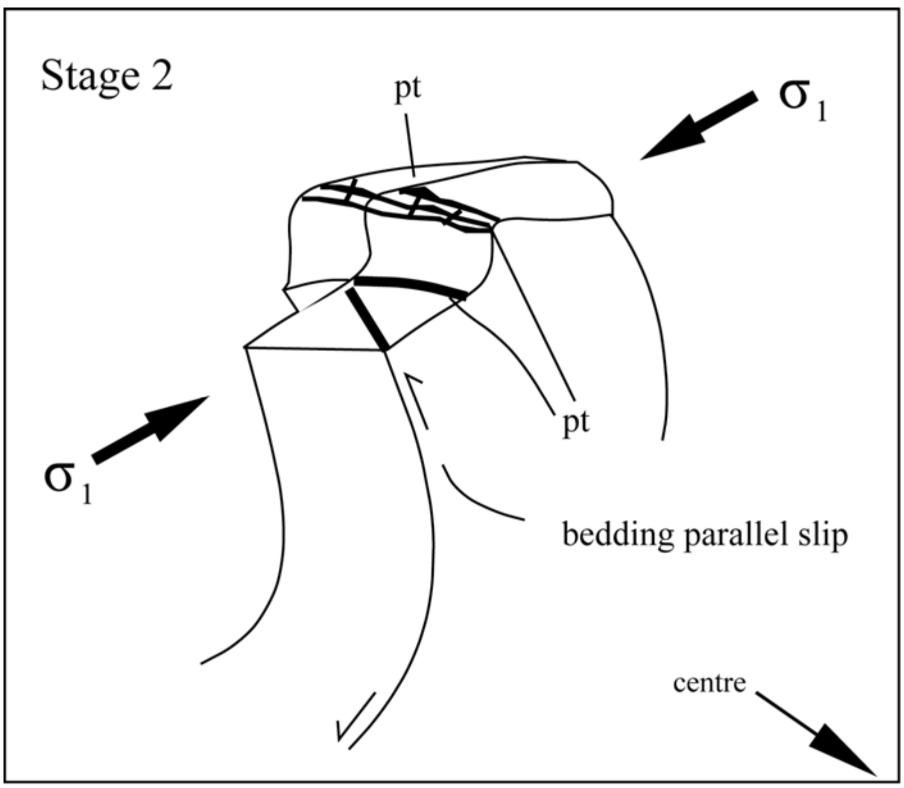

Fig. 6.8: Simplified illustration of impact-induced deformation related to the rise of the central uplift (stage 2). First, slip along bedding-parallel surfaces occurred, then folds were formed due to the accommodation problems during tangential shortening. Coevally or directly after the folding, large-scale radial and oblique strike-slip faults were formed, bounding the folds and disrupting the fold hinges. Formation of pseudotachylitic breccias along these structures and intrusion of melts into dilational sites, such as hinge zones, is likely (labelled pt, see text). The strata are up- to overturned and movement occurs along bedding surfaces (bedding-parallel slip indicated by arrows). The principal stress (σ_1) is tangential to the central uplift.

Possibly overlapping the latter stages of the contractional phase of central uplift formation, the collar rocks may have undergone some tangential faulting and block rotation around horizontal tangential axes (see Fig. 2.24 and different dip direction of bedding on fold limbs as described above and in section 2.5). Equally, this could be attributed to the collapse of the central uplift.

6.3.3 Collapse of the central uplift

The granophyre dykes and ubiquitous joints indicate radial and tangential extension (see section 2.7), with joint geometries consistent with late-stage collapse of the central uplift (stress-release) as discussed in section 6.2.2.2 (Fig. 6.9). The principal compressive stress (σ_1) is now oriented vertically downward. The extensional joints indicate a brittle-dominated regime, which is consistent with regard to results of recent studies (see Dence 2004, and discussion in section 6.2.1). It is possible that earlier-formed strike-slip faults may have undergone reactivation or become dilated during this phase. The injection of melts along km-long extensional fractures is indicated by the presence of the granophyre dykes and pseudotachylitic breccia veins (Fig. 6.9).

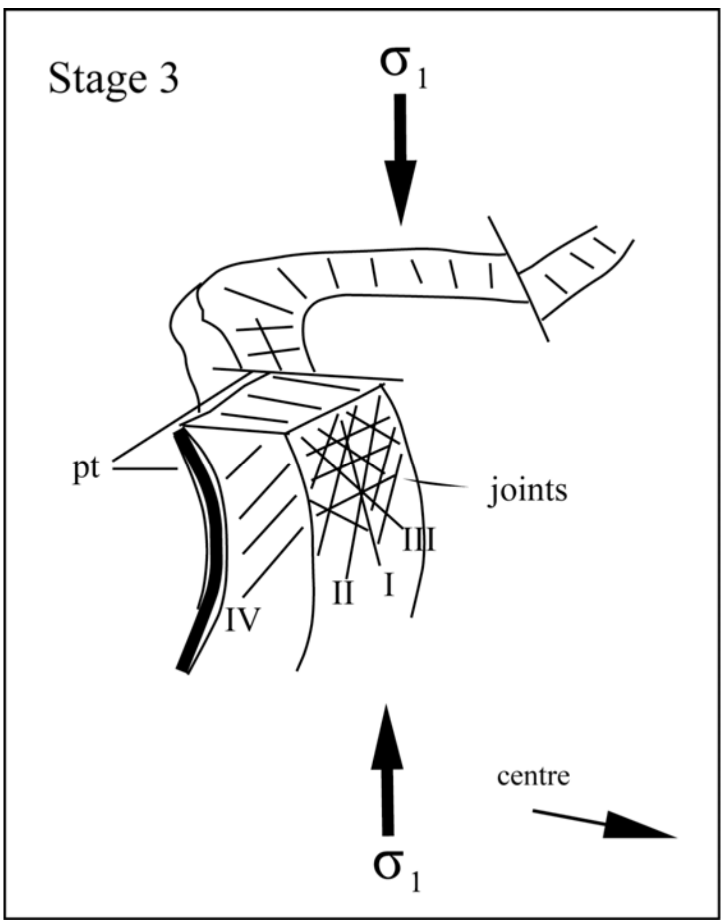

Fig. 6.9: The third stage of impact cratering is linked to the collapse of the central uplift, which produced at least four sets of extensional joints/shear fractures (labelled I to IV). Their geometries are consistent with radial and tangential extension. Thin, bedding-parallel pseudotachylitic breccia veins are cut and, in places, displaced by these shear fractures. The direction of the principal compressional stress (σ_1) is vertical.

6.3.4 Gravitational adjustment

Recent numerical modelling studies (e.g., Collins et al. 2004; Ivanov 2005) suggest that the net strain in a central uplift at the end of the cratering process is much lower than the total strain, due to opposite-directed movement patterns (see discussion in section 6.2.2). Thus, the inward slumping of the crater walls in the outer parts of a large impact crater due to gravitational adjustment may be the dominant feature that is observed in the outer parts of the Vredefort Dome. Listric normal faults, observed in the goldfields of the Witwatersrand Basin (e.g., Killick 1990; Killick and Reimold 1990; Killick and Roering 1995; Reimold et al. 1999), may be the manifestation of such inward slumping in the Vredefort structure and are in agreement with findings from other impact sites (e.g., Haughton, Bischoff and Oskierski 1988; Upheaval Dome, Kriens et al. 1999; Kenkmann et al. 2005; Gosses Bluff, Milton et al. 1996; Araguainha, Bischoff and Prinz 1994; Lana et al. 2005, see section 6.2.1).

Besides the listric faults, some of the extensional joints may also have formed during this stage. Although evidence was presented in Chapter 2 for a general impact origin of these fractures, a reactivation or a formation of some of these extensional joints in post-impact times due to unloading by erosion or post-impact tectonic events cannot be ruled out. This is because there is still some uncertainty of when the modification phase actually terminates (see section 6.2.2.2), which makes it impossible to differentiate between the formation of extensional joints in the final stages of the modification phase or by “true” post-impact tectonic activities.

6.4 Further work

This study has provided the results of a detailed investigation of brittle and ductile deformation in the collar rocks of the Vredefort Dome. Apart from a general documentation of structures and deformation of the collar strata in an attempt to establish temporal relationships between these structures, detailed structural analysis of areas of structural complexity, such as major folds, faults and zones of block rotation was conducted and interpreted in the light of predicted impact-related deformation in the central uplift of a large impact structure. A temporal sequence of impact-related deformation could be established, based on the spatial relationships between structures, and, thus, a step was made towards better understanding of the formation of - and the mechanisms during central uplift formation. However, there are still some problematics and inconsistencies. In particular, the origin of pseudotachylitic breccias (shock versus friction melt) in impact structures needs further attention. Although this study suggests that it may be possible for shock melts to stay molten until the final stages of the modification phase, the existence of friction-generated melts in the Vredefort Dome cannot be ruled out. Thus, the true origin of pseudotachylitic breccias in the Vredefort Dome and other impact structures remains problematic, and further, especially small-scale, field and laboratory work is required. Similarly important, the origin and time of formation of shatter cones in the Vredefort Dome and in other impact structures must be further addressed. The observations made during this study have contributed to a better understanding of the genesis of shatter cones and have complemented previous field observations and numerical modelling on this fracture phenomenon. However, this study demonstrated that there are still deficiencies in the existing models for shatter cone formation. In particular, the precise time of formation of shatter cones, during the initial stage of cratering, remains enigmatic.

Geophysical studies, such as seismic and magnetic profiling, of the Vredefort Dome might be useful to investigate the subsurface orientations of structures in the collar rocks (see discussion in section 5.5). As demonstrated in this study, the large-scale faults are generally covered by soil and vegetation on surface and, based on observations of similar structures in other impact structures, could turn from vertical strike-slip faults into listric faults below surface (see section 6.2.1). Thus, further local geophysical observations might essentially contribute to the results of the detailed surface mapping during this study.

The deep level of erosion of the Vredefort Dome makes it difficult to apply the results of impact-related deformation obtained during this study to other, less eroded, impact structures. Recent modelling studies of very large impact structures (e.g., Collins et al. 2004; Dence 2004; Ivanov 2005) provided a new view on the cratering process in deeper levels of such craters and are in agreement with the findings of this study. In order to understand the complex mechanisms of the cratering process in such a large structure, future numerical modelling needs to consider additional parameters, such as a heterogeneous target sequence (as present in the collar rocks of the Vredefort Dome), pre-impact deformation (as provided by the complex tectonic history of the Witwatersrand Basin), and the possible reactivation of pre-impact structures, and possible strength degradation of already fragmented target rocks. The Vredefort Dome is the only impact structure on Earth where it is possible to study the deep root zone of a giant impact structure and where the results of such numerical modelling can be (and must be) validated by field observations.

6.5 Summary and conclusions

This detailed structural investigation of the collar rocks of the Vredefort Dome and the region to the northwest of the collar is predominantly based on surface mapping. The project covered the exposed meta-sedimentary strata of the Witwatersrand, and to some degree, the Ventersdorp and Transvaal supergroups in the northeastern, northern, northwestern, western and southeastern parts of the collar of the Vredefort Dome and the outer parts of the Vredefort impact structure. Field mapping was further supported by the analysis of Landsat-TM images and aerial photograph and orthophoto analysis. All structures, such as folds, faults, fractures, shatter cones, pseudotachylitic breccias and lineations (shatter cones, slickenside striations, etc.) were documented.

In contrast to the massive crystalline core and poorly-exposed outer parts of the Vredefort Dome, the well-layered siliciclastic rocks of the Witwatersrand Supergroup in the inner collar of the dome preserve a variety of folds, faults, fractures and melt breccias that can be related to the 2.02 Ga impact event. Symmetric and asymmetric folds with wavelengths of up to a few kilometres are arranged radially with respect to the centre of the dome, generally displaying inward-plunging fold hinges and subvertically oriented axial planes, although oblique orientations are found locally that are attributed to subsequent fault block rotation. In addition, metre-scale “curtain-type” folds are present on the limbs of such large-scale folds. Large-scale faults range in length from a few hundred metres to several kilometres and display consistent, conjugate geometries with radial and oblique orientations. Strike-slip displacements along these faults are supported by subhorizontal linear grooves in the fault zones. The faults bound the large-scale folds and disrupt and displace the fold hinges. The bulk of the fold and fault structures relate to tangential shortening of the strata, which was calculated at ~ 20% in strata closer to the centre of the dome (~20 km from the centre) and only ~10% further out (~25 km from the centre), which is consistent with recent results of numerical modelling studies (e.g., Collins et al. 2004;

Ivanov 2005). Shear fractures are ubiquitous in the collar rocks and may be filled by pseudotachylitic breccia veins but also cut and displace them in places. Five main sets of extensional joints were identified in the collar rocks, with bedding-parallel and radial strike orientations. The overall geometry of these joint sets indicates a triaxial stress release, but with a vertically oriented principal stress (σ_1).

The shatter cones in the collar rocks of the Vredefort Dome demonstrate the complex aspects of shatter cone formation. The variation in shatter cone orientation, surface morphology and striation geometry concur broadly with previous studies (e.g., Manton 1962, 1965; Milton et al. 1996; Nicolaysen and Reimold 1999). The "master cone" concept (e.g., Albat and Mayer 1989) and variation of striation angles with the distance of the shatter cone location from the centre of the structure (Sagy et al. 2002, 2004) were not substantiated by field observations during this study. The abundance of striated surfaces along closely-spaced sets of fractures (MSJS) observed in this study can be reconciled with reflection/scattering of a fast propagating wave at heterogeneities in the target rocks, as proposed by Sagy et al. (2002, 2004) and Baratoux and Melosh (2003). None of the existing hypotheses can fully explain the variation of shatter cone orientation, surface morphology, striation geometry and striation angles that are present in the collar rocks of the Vredefort Dome. A combination of aspects of the recent hypotheses seems presently the best solution to explain most of the characteristics of shatter cones that were observed during this study.

In contrast to the massive occurrences of pseudotachylitic breccias in the core of the Vredefort Dome, pseudotachylitic breccias in the collar rocks are less voluminous. They typically occur as up to 1 cm wide veins with orientations parallel and oblique to bedding. More voluminous pods and veins (up to tens of cm in width) are preferentially, but not exclusively, found in sites of structural complexity, such as fold hinges and fault traces where they have intruded extensional sites within these structures. Pseudotachylitic breccias in the collar strata are, in places, cut by extensional joints and shear fractures. Simplified cooling calculations that were performed for veins with different widths (1 mm, 1 cm, 10 cm) indicate that thin (1 mm) veins probably quenched almost instantaneously, whereas thick (10 cm) veins might have stayed molten for hours and even days, thus allowing the melt to migrate into impact-induced sites of structural complexity (such as folds, faults and fractures).

Based on spatial and geometric relationships between these structures, a temporal sequence of deformation in the Vredefort Dome can be established. The narrowly-spaced fractures (MSJS) are believed to have formed either during shock compression or decompression. Given the close spatial association of shatter cone surfaces with these types of fractures and recent numerical modelling (e.g., Baratoux and Melosh 2003), it is postulated that the shatter cones formed after the passage of the main shock pulse and are a product of stress release. At least some shock-induced, pseudotachylitic breccia is believed to have also formed in this phase; however, evidence for friction-generated melts later in the cratering process may be provided by the presence of pseudotachylitic breccias in the vicinity of large-scale faults.

The next deformation phase in the collar of the Vredefort Dome is represented by the large-scale radial folds and large-scale radial and oblique faults. These structures reflect the accommodation problems experienced during the inward movement of material during the rise of the central uplift. However, the total amount of shortening and total amount of displacement during the formation of the central uplift is relatively minor, as demonstrated by the low amounts of tangential shortening determined during this study. Friction-generated melts may also have formed during this phase. Tangential shortening was followed by radial and tangential extension related to collapse of the central uplift that produced ubiquitous jointing and rarer faults, and provided the opportunity for downward intrusion of dykes from the impact melt sheet (Vredefort Granophyre). Crosscutting relationships of joints and shear fractures with pseudotachylitic breccias confirm an impact origin of these small-scale structures late in the modification phase. However, a reactivation of such structures, or a formation of at least some of them, in post-impact times cannot be ruled out.

This study has shown the complexity of impact-related deformation in the collar rocks of the Vredefort Dome. The unique opportunity to study the root zone of a giant impact structure revealed some differences to brittle and ductile deformation found in the central uplift zones of more shallowly exhumed and smaller impact structures (e.g., Sierra Madera, Wilshire et al. 1972; Upheaval Dome, Kriens et al. 1999; Kenkmann et al. 2005; Araguainha, Bischoff and Prinz 1994; Lana et al. 2005; Gosses Bluff, Milton et al. 1996). However, the results are in agreement with findings of recent modelling studies of large impact structures (e.g., Collins et al. 2004; Dence 2004; Ivanov 2005).

Although the findings of this study contributed essentially to a better understanding of the processes and mechanisms involved in the cratering process of very large impact structures, some problems remain unsolved:

- The origin and timing of crystallization of pseudotachylitic breccias within the Vredefort impact event remains problematic.
- The timing and genesis of shatter cone formation need further attention.
- Further geophysical studies and numerical modelling on the Vredefort Dome are necessary to complement the results of surface mapping, especially with regard to large-scale structures, such as faults, and their sub-surface orientation.