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Structural characteristics of the Sudbury impact structure, Canada: Impact-induced versus orogenic deformation—A review

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Abstract–Orogenic deformation, both preceding and following the impact event at Sudbury, strongly hinders a straightforward assessment of impact-induced geological processes that generated the Sudbury impact structure. Central to understanding these processes is the state of strain of the Sudbury Igneous Complex, the solidified impact melt sheet, its underlying target rocks, overlying impact breccias and post-impact sedimentary rocks. This review addresses (1) major structural, metamorphic and magmatic characteristics of the impact melt sheet and associated dikes, (2) attempts that have been made to constrain the primary geometry of the igneous complex, (3) modes of impact-induced deformation as well as (4) mechanisms of pre- and post-impact orogenic deformation. The latter have important consequences for estimating parameters such as magnitude of structural uplift, tilting of pre-impact (Huronian) strata and displacement on major discontinuities which, collectively, have not yet been considered in impact models. In this regard, a mechanism for the emplacement of Offset Dikes is suggested, that accounts for the geometry of the dikes and magmatic characteristics, as well as the occurrence of sulfides in the dikes. Moreover, re-interpretation of published paleomagnetic data suggests that orogenic folding of the solidified melt sheet commenced shortly after the impact. Uncertainties still exist as to whether the Sudbury impact structure was a peak-ring or a multi-ring basin and the deformation mechanisms of rock flow during transient cavity formation and crater modification.

INTRODUCTION

According to cratering rate statistics, far more impact structures should be present on the Earth's surface than are currently recognized. The low number of known impact structures larger than 150 km in diameter in particular, and the difficulty in identifying older ones may be, in addition to the effects of surface processes such as erosion, due to the modification of impact structures by orogenic processes. The 1.85 Ga Sudbury impact structure in Ontario (Pye et al. 1984), Canada, is the most prominent terrestrial impact structure that underwent a strong shape change during Proterozoic orogenesis. The Sudbury structure is widely regarded as the erosional relic of a deformed multi-ring impact basin with an estimated diameter of up to 250 km or larger (Butler 1994; Spray et al. 2004; Pope et al. 2004).

Similar to Chicxulub and Vredefort, the Sudbury impact structure is one of the largest impact structures known on Earth (Grieve and Theriault 2000), but it is the only one hosting an exposed differentiated impact melt sheet, the 1.85 Ga Sudbury Igneous Complex (SIC: Krogh et al. 1984, 1996). The northern portion of this complex, the North Range, is underlain by Archean granitoid and granulite basement rocks, whereas the southern portion, the South Range, rests on polydeformed Paleoproterozoic metasedimentary rocks of the Huronian Supergroup (Fig. 1). Collectively, these rocks and the 2.2 Ga Nipissing dikes and sills formed the target material and display a variety of impact-induced effects, such as planar deformation features in quartz, feldspar, and zircon, as well as shatter cones and various breccia types, notably clast-rich pseudotachylite bodies (e.g., Dressler 1984a). The Main Mass of the SIC is assumed to have formed entirely by shock-induced fusion of upper crust and this impact melt was ponded in a large crater with a relatively horizontal floor (e.g., Grieve et al. 1991; Grieve 1994; Deutsch et al. 1995; Theriault et al. 2002). Static differentiation of the superheated impact melt sheet into the observed basal so-called gabbroic to noritic Sublayer and overlying norite, gabbro and granophyre sheets (Fig. 2a), respectively, occurred prior to geometric modification of these layers into an asymmetric synform (Milkereit et al. 1992; Boerner and Milkereit 2000) as a consequence of

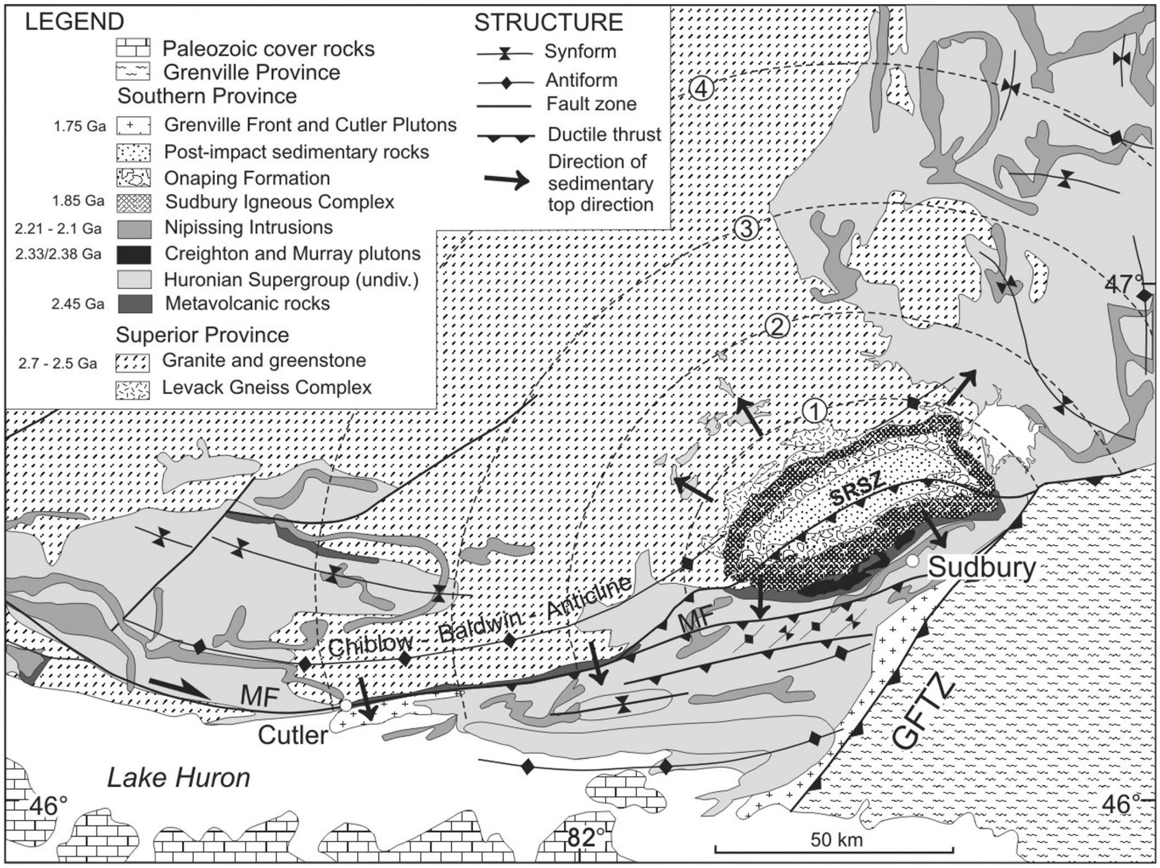

Fig. 1. Simplified tectonic map, modified from Card et al. (1984), showing the relics of the deformed Sudbury impact structure as part of the Eastern Penokean Orogen in the Canadian Shield. The Sudbury Igneous Complex (SIC) rests on Archean rocks, notably granulites of the 2.7 Ga Levack Gneiss Complex in the north, and Paleoproterozoic rocks of the Huronian Supergroup in the south. Note isolated exposure of Huronian rocks at a distance of about 15 km NW of the SIC, the direction of stratigraphic younging indicated by arrows in circumferential Huronian strata around the SIC and evident to a distance of 70 km west of the SIC. Stippled half circles correspond to the structural rings identified by Butler (1994), i.e., ring 1, and those proposed by Spray et al. (2004), i.e., rings 2–4. SRSZ: South Range Shear Zone, GFTZ: Grenville Front Tectonic Zone, MF: Murray Fault.

Proterozoic orogenic deformation (Shanks and Schwerdtner 1991a; Deutsch and Grieve 1994).

Orogenic deformation, both preceding and following impact at Sudbury, strongly hinders a straightforward assessment of the impact-induced geological processes that generated the Sudbury impact structure. Geochemical and petrological data suggest that layer interfaces within the Main Mass of the SIC formed by static magmatic differentiation of a coherent impact melt sheet (e.g., Grieve et al. 1991; Deutsch et al. 1995; Ostermann 1996; Ariskin et al. 1999; Dickin et al. 1999). Thus, the layer interfaces record paleo-horizontal surfaces that were generated following collapse of the transient cavity. However, the mechanism and relative importance of impact-induced and orogenic components of deformation that modified the shape of the SIC are unknown. In addition to erosion, it is largely because of these structural uncertainties that the location of the original center of the impact structure, its size, and its proposed multi-ring nature are still debated. All of these structural variables are paramount in establishing ground truth information to better understand differential crustal motions, deformation mechanisms, mobility of impact melts, and associated generation of ore deposits induced by a large meteorite impact.

This review aims at providing a base for developing further working strategies directed towards understanding the relationships between the effects of deformation imparted to rocks by orogenic processes and complex large meteorite impact. More specifically, major structural, metamorphic, and magmatic characteristics of the SIC and its underlying target and post-impact sedimentary rocks are assessed in terms of their importance for comprehending processes of large meteorite impact at Sudbury. Following a discussion of major structural and lithological characteristics of the impact structure, attempts to constrain the primary geometry of the SIC and pre-impact as well as post-impact deformation, impact-induced deformation along with an account on the formation of quartz-dioritic dikes emanating from the Main Mass is provided. In cases, an interpretation of geological observations or processes based on published data is offered.

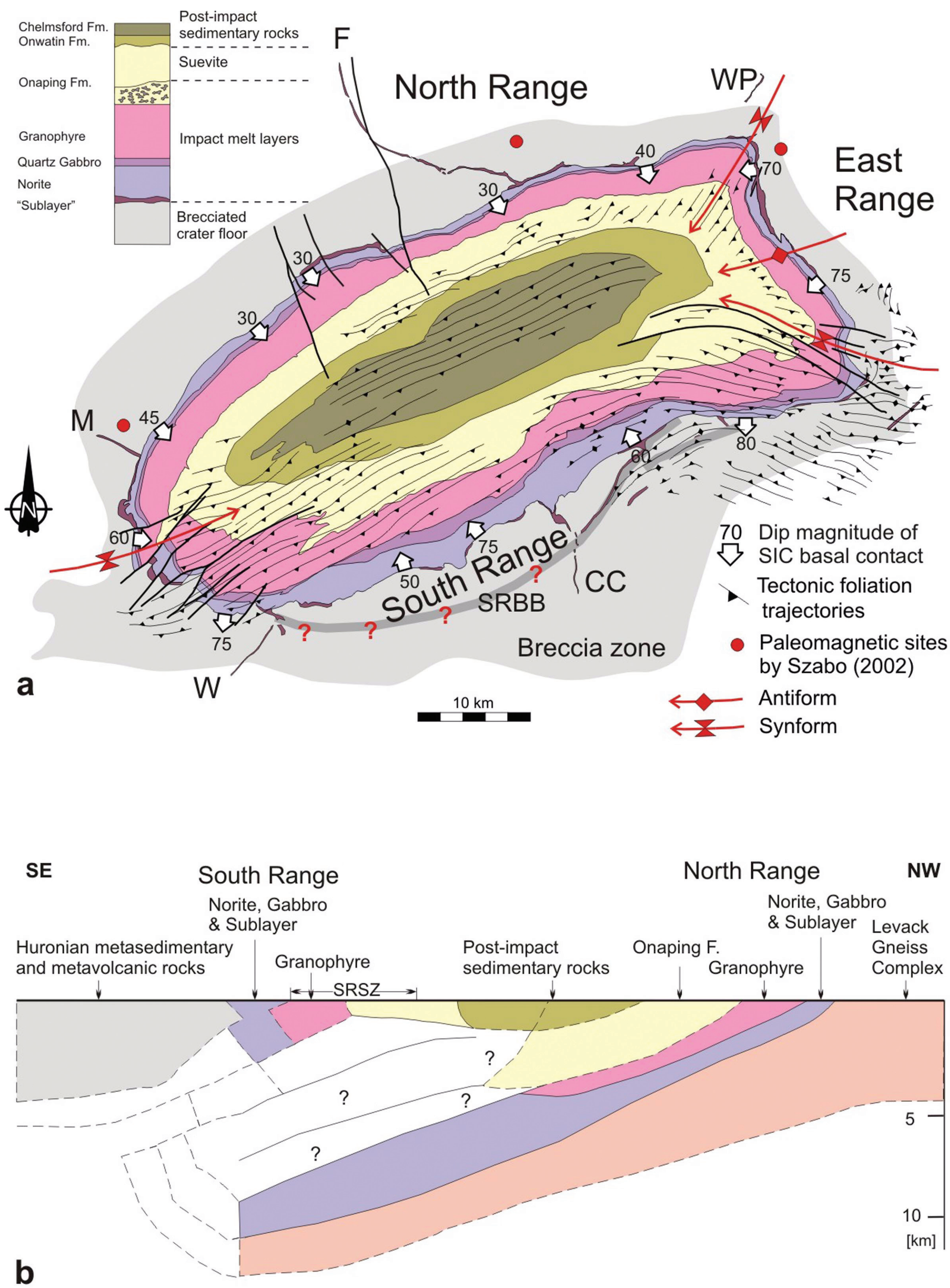

Fig. 2. a) A simplified geological map of the SIC and its Archean and Paleoproterozoic target rocks showing trajectories of post-impact foliations and the dips of the base of the SIC (modified from Cowan et al. [1999]). Offset Dikes are W: Worthington, CC: Copper Cliff, F: Foy, WP: Whistle-Parkin, M: Ministic. The continuity of the South Range Breccia Belt (SRBB) west of the Copper Cliff dike is uncertain. b) NW-SE profile across the deformed central portion of the Sudbury impact structure based on seismic profiling by Milkereit et al. (1994).

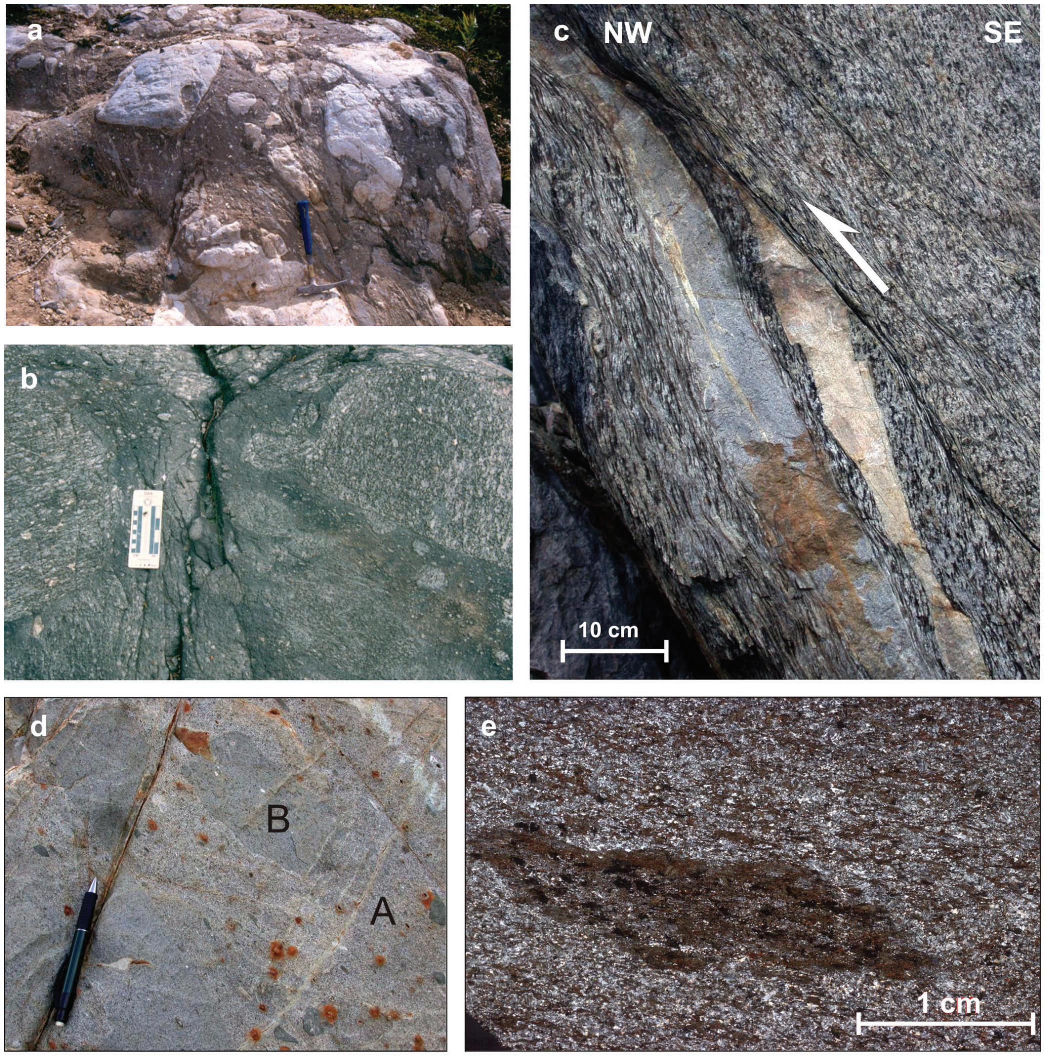

Fig. 3. Photos showing important rock types of the Sudbury impact structure. a) Clast-rich pseudotachylite in Proterozoic rocks at the intersection of Highways 144 and 17. Elliptical host rock fragments are enveloped by recrystallized pseudotachylite. Note the hammer for scale. b) Deformed clast-rich pseudotachylite in the Creighton Pluton. Note preferred alignment of elliptical clasts enveloped in recrystallized pseudotachylitic matrix, indicating post-impact strain. c) Example of a strained gabbro from the South Range Shear Zone, showing asymmetric fabric geometry indicating SE over NW sense-of-shear. d) Mingling of two melt phases in the center of the Worthington dike. Note, that the light phase (A) contains fragments of target rocks and sulphide blebs seen as rust spots, whereas the dark phase (B) is devoid of both. e) Photomicrograph (plane-polarized light) showing a stretched metavolcanic fragment in strained quartz diorite of the Worthington Offset Dike. Note the strong shape-preferred orientation of metamorphic biotite, quartz, and plagioclase. The section is cut perpendicular to the foliation and parallel to the lineation delineated by biotite and quartz.

This review ends with a personal view of the most prominent unsolved structural uncertainties in the Sudbury area. For a summary of general geological characteristics of the Sudbury area, the reader is referred to Dressler (1984b), Dressler et al. (1991), and Rousell et al. (2002).

MAJOR STRUCTURAL AND LITHOLOGICAL CHARACTERISTICS OF THE SUDBURY IMPACT STRUCTURE

The synformal Main Mass of the SIC, along with its overlying Onaping Formation and post-impact sedimentary rocks (see below), is known as the Sudbury Basin (Brocoum and Dalziel 1974). At surface, the SIC varies in thickness from between about 2 km in the North and East Ranges to about 2.5 km in the South Range (Fig. 2b). Its lower contact dips mostly toward the center of the elliptical outline of the SIC in map view, but dip magnitudes vary greatly between about 35° in the North Range, up to 75° in the East Range, and between 50° and 75° in the South Range (Rousell 1984b). Near the SE lobe and the southwestern terminus of the South Range SIC, the lower contact dips steeply away from the center of the exposed SIC (Fig. 2a). Paleoproterozoic metasedimentary rocks below the South Range of the SIC are intruded by the 2.3 Ga Creighton Pluton (Frarey et al. 1982) and the 2.48 Ga Murray Pluton (Krogh et al. 1996) and are overturned to the south (Cooke 1948; Card 1965; Card and Palonen 1976). The North Range and much of the East Range of the SIC are underlain by the 2.71 Ga granulite-facies metamorphic rocks of the Levack Gneiss Complex (Krogh et al. 1984) that is intruded by NW-trending mafic dikes of the 2.45 Ga Matachewan dike swarm (Heaman 1997).

The most prominent macroscopic result of the meteorite impact in the Sudbury area is the pervasive presence of clast-rich pseudotachylite in Archean and Proterozoic target rocks. These impactites form millimeter-wide veins to hundreds of meter wide apparently tabular zones of clast-rich breccia bodies (Fig. 3a) that are spatially associated with rock discontinuities such as faults, bedding planes, and lithological contacts (Dressler 1984a; Rousell et al. 2003). Lithic fragments in these bodies may be up to many meters in diameter, are generally well-rounded, elliptical in plan view, and chiefly derived from adjacent lithological units (Dressler 1984a). The matrix of pseudotachylite veins and clast-rich pseudotachylite bodies is fine-grained to aphanitic and often shows evidence for ductile flow during brecciation (Rousell et al. 2003; Legault et al. 2003). Collectively, this points to in situ formation of these impactites due to the passage of a shock wave (cf. Kenkmann et al. 2000) resulting in subsequent frictional shearing and cataclastic flow playing an important role (Spray and Thompson 1995; Spray 1997). Pseudotachylitic breccias are most pervasive within a distance of about 10 km from the North and East Ranges and about 15 km from the South Range of the SIC (Fig. 2a). However, there also appears to be an approximately 10- to 15-km-wide concentric zone of brecciation located about 20 to 25 km north of the Main Mass of the SIC as well as isolated pseudotachylitic breccia bodies found at a distance of up to 80 km from the SIC (Dressler 1984a).

The norite of the Main Mass and the Sublayer, which occurs as discontinuous sheets below the norite, are characterized by embayments in underlying target rocks (Fig. 2a). Quartz-diorite dikes of the SIC, due to their segmentation known as Offset Dikes, appear to emanate from these embayments suggesting that these are linear depressions at the base of the Main Mass. The dikes intrude for up to about 30 km subradially into Proterozoic and Archean target rocks (Grant and Bite 1984; Tuchscherer and Spray 2002). Generally, the dikes vary in thickness between ten and hundreds of meters but taper off away from the Main Mass of the SIC. Northwest-striking dikes, e.g., the Copper Cliff and the Foy (Fig. 2a), are curved at the surface, and are more affected by apparent offsets than NE-striking dikes, such as the Worthington, Whistle and Parkin, which are rather straight at surface. Except for the Worthington dike, which dips between 60 and 80° toward the SE, the quartz-diorite dikes are reported to be subvertical (Grant and Bite 1984).

Lithological correlation of inclusions in quartz-diorite dikes with local host rock units, as well as the presence of norite and Sublayer fragments in the dikes (Grant and Bite 1984; Dressler et al. 1991), suggest transport of quartz-diorite magma away from the Main Mass of the SIC, with the dikes geochemically akin to the bulk Main Mass (Lightfoot et al. 1997; Ariskin et al. 1999; Lightfoot and Farrow 2002). Offset Dikes are also concentrically arranged around the lower contact of the Main Mass of the SIC (Fig. 2a). They truncate (and are often found within) zones of pseudotachylitic breccia, notably in Paleoproterozoic target rocks where dike margins display little or no chilling and the host rocks are devoid of thermal metamorphic effects (Grant and Bite 1984). The South Range Breccia Belt (SRBB in Fig. 2a: Spray 1997) is a spectacular example of a several-hundred-meter-wide concentric zone of pseudotachylitic breccia hosting quartz-diorite, as well as having hosted the largest Cu-Ni sulfide ore deposits in the world, the Frood-Stobie deposit.

The Main Mass of the SIC is overlain by the Onaping and the Onwatin Formations, as well as the now concentrically folded strata of the Chelmsford Formation of the Whitewater Group (Fig. 2). All contacts between these formations, and the Onaping Formation with underlying SIC, are gradational (Rousell 1984a). The Onaping Formation is a polymict impact breccia containing melt inclusions, lithic fragments, and shock-metamorphosed minerals derived from the Archean and Proterozoic target rocks (Avermann and Brockmeyer 1992; Krogh et al. 1996). Consequently, the Onaping Formation has been interpreted by some as a fall-back deposit of target rock fragments that were ejected into the atmosphere during early crater excavation (Peredery and Morrison 1984; Avermann 1999). The Onwatin Formation consists of massive to laminated pelagic argillite and siltstone deposited in a closed basin (Rousell 1984a). By contrast, greywacke of the Chelmsford Formation points to a proximal deposition by southwest-directed turbidity currents.

Rocks of the Sudbury impact structure are heterogeneously deformed, as indicated by the distribution and variety of structures as seen at surface (Fig. 2a). Much of the Main Mass of the SIC, the overlying Whitewater Group, and Huronian target rocks, including pseudotachylitic breccia and Offset Dikes south of the SIC, were affected by deformation under greenschist-facies and, locally, lower amphibolite-facies metamorphic conditions (Card 1978; Thomson et al. 1985; Fleet et al. 1987). This has led to the development of mesoscopic planar and linear mineral shape fabrics, the intensity and metamorphic grade of which decrease generally towards the north. As a consequence, the North and East Ranges, including the NE and SE lobes but also much of the norite and gabbro in the South Range and the western terminus of the SIC, are apparently not affected by mesoscopic ductile strain. Except at the eastern lobes, planar mineral shape fabrics strike parallel to the NE-trending structural grain in the Sudbury area and are either subvertical or dip moderately to steeply southward. In the lobes, shape fabrics are axial-planar to the bisectrices of the curved SIC in plan view (Cowan 1999) evoking a fold origin of both lobes

(Fig. 2a). In the Chelmsford and Onwatin Formations, planar fabrics developed as a slaty cleavage, which is axial planar to folds of the bedding planes, with doubly plunging axes (Rousell 1984b). The generation of pervasive shape fabrics in the SIC and its overlying rocks is generally attributed to the Paleoproterozoic Penokean orogeny (e.g., Sims et al. 1989).

It is uncertain, however, whether Penokean deformation north of Lake Huron commenced prior to impact and how long it actually lasted. North of Lake Huron, Penokean deformation is characterized by NW-SE shortening, accomplished by doubly-plunging folds and south-dipping reverse faults (Zolnai et al. 1984), mostly under greenschist-facies metamorphic conditions (Card 1978). The most prominent structure in the Sudbury area, that is likely associated with Penokean deformation, is the South Range Shear Zone (SRSZ in Fig. 2b: Wilson 1956; Rousell 1975; Shanks and Schwerdtner 1991a), a south-dipping ductile thrust that displaced the South Range and its underlying Huronian rocks toward the NW. This zone transects the Main Mass of the SIC south of the SE-lobe (Cowan et al. 1999) and at the southwestern terminus, where strain is concentrated at the interface between Archean basement and Huronian cover rocks (Riller et al. 1998). Mineral shape fabrics in the central South Range Shear Zone dip generally to the SE and are characterized by a strong asymmetric fabric geometry (Fig. 3c), with down-dip mineral lineation indicating a top-to-NW reverse sense-of-shear (Shanks and Schwerdtner 1991a). By contrast, the eastern segment of the zone is made up of subvertical, anastomosing planar shape fabrics. This segment is kinematically interpreted as the lateral terminus of the shear zone, which accommodated reverse sense-of-shear in the central segment by right-lateral transpression (Cowan and Schwerdtner 1994).

North-striking brittle faults of the Onaping Fault System cut the North Range (Fig. 2a). Strike separations of displaced contacts of the SIC on the most western faults are less than a kilometer and the vertical component of displacement is about 150 m (Rousell 1984b). This suggest that these faults were of minor importance in terms of the shape change of the SIC. Prominent SE-dipping faults, which traverse the SIC parallel the trace of the South Range Shear Zone at surface, are more important in this respect (Dressler 1984a, Rousell 1984b). Strike separations on these faults at the southwestern terminus of the SIC are on the order of 4 km. Based on the inclination of the lower contact of the SIC in this area, pronounced differences in thickness of the SIC and the Onaping Formation are due to a reverse sense of displacement on these faults. By contrast, little horizontal displacement exists on these faults where they cut across the SIC at the SE lobe.

ATTEMPTS TO CONSTRAIN THE PRIMARY SHAPE OF THE SIC

The elliptical outline of the Main Mass of the SIC in plan view has prompted a number of workers over the past 20 to 30 years to test whether the SIC was approximately circular in plan view prior to orogenic deformation (e.g., Roest and Pilkington 1994). Also, the interpretation of the SIC as a ponded impact melt sheet (e.g., Grieve et al. 1991) prompted structural analysts towards investigating whether the layers of the Main Mass of the SIC were originally horizontal. To date, still no conclusive answer exists with respect to either of these structural problems.

Using apparently the shapes of strained carbonaceous concretions contained in greywacke of the Chelmsford Formation, Brocoum and Dalziel (1974) suggested that the SIC was most likely circular prior to orogenic deformation. Later, Clendenen et al. (1988) used the shapes and orientations of the carbonaceous concretions to calibrate measurements of the anisotropy of magnetic susceptibility (AMS) in greywacke to estimate the geometry of finite strain throughout the exposed formation. Their results indicate that the long axis of the SIC, at surface, did not undergo significant length changes (Clendenen et al. 1988). Furthermore, decomposition of the finite strain tensor suggested that the Chelmsford Formation underwent phases of compaction and layer-parallel shortening (38%) in NW-SE direction, prior to buckle folding of sedimentary strata. By quantitatively correlating the geometries of AMS fabrics and strain inferred from fragment shapes of the Onaping Formation, Hirt et al. (1993) confirmed this deformation path. They also concluded that the axial-planar foliation to buckle folds of the Whitewater Group was overprinted by strain imposed on these rocks by SE-over-NW simple shear of the South Range Shear Zone and that the SIC was most likely circular prior to deformation.

However, there are significant quantitative uncertainties associated with the above structural analyses. First of all, estimating the strain only in the Whitewater Group may not reveal much about the original geometry of the impact structure, as it neglects deformation in the now eroded portions of the structure, which covered a much larger area than the exposed portion of the Whitewater Group. Moreover, the overall heterogeneity of strain and deformation of the impact structure, and the competency contrast between the markers used as strain gages and their sedimentary host rocks, render the strain estimates inaccurate at the scale of the impact structure. For a critical assessment on the specific use of carbonaceous concretions in this respect, the reader is referred to Rousell (1984b, page 93). Concerning the use of magnetic fabrics as strain indicators, there is considerable debate as to whether magnitudes of principal AMS ellipsoid axes do correlated with those of strain ellipsoids (e.g., Borradaile 1988). Moreover, the dominant magnetic carrier minerals may have formed after deformation of a given rock, in which case the AMS fabric is unrelated to strain. For instance, pyrrhotite is a ubiquitous mineral in the Sudbury area and has been identified as the most prominent carrier of AMS in the Onaping Formation (Hirt et al. 1993). Pyrrhotite forms and is mobile under hydrothermal conditions (Schwarz

1973), which calls into question the use of AMS fabrics as strain gages to quantitatively determine the primary shape of the SIC.

Larger-scale reconstructions of the primary shape of the SIC are at present restricted to two-dimensional analyses. Shanks and Schwerdtner (1991b) applied a crude 2D finite element analysis parallel to the NW-SE diameter of the elliptical SIC, which constitutes the symmetry plane to the ductile deformation field (Shanks and Schwerdtner 1991a), to constrain the range of possible pre-deformational diameters and shapes of the SIC along this profile. Boundary conditions of these analyses include solid-body tilt of the basal contact of the norite, total ductile strain, as well as shear strain imparted by the reverse shearing on the South Range Shear Zone. The most plausible 2D reconstructions of the NW-SE diameter suggest that the presently exposed SIC could have been 74 km to 90 km wide, with differential vertical displacement between the North Range and the South Range being less than 15 km. Most interestingly, however, the presently exposed portion of the South Range constitutes the central portion and thus the deepest part of an originally funnel-shaped SIC. The analysis does not permit conclusions as to whether the SIC was circular prior to orogenic deformation.

The SIC is characterized by pronounced potential field anomalies, specifically magnetic and gravity (McGrath and Broom 1994). Roest and Pilkington (1994) used the elliptical plan view pattern of these anomalies as marker surfaces for heterogeneously destraining the SIC in plan view. This 2D reconstruction resulted in an almost perfect pre-strain circular shape of the SIC. Unfortunately, their work leaves open how the 2D strain field on which the reconstruction is based was acquired and to what extent discontinuous deformation was considered. Unlike the finite-element model by Shanks and Schwerdtner (1991a), the reconstruction does not take into account vertical stretching of rock contained in (1) steeply inclined mineral lineations of ductilely deformed rocks, (2) folds in the Whitewater Group and the SIC and (3) zones of reverse shearing. Thus, the reconstruction is based on the unrealistic assumption that the pre-deformational erosion surface coincides with today's peneplain. Most importantly, the variation in strain and its intensity predicted from the reconstruction does not correspond to observed patterns of these quantities. For example, the East and North Ranges as well as the NE lobe are effectively unstrained (see below), but the reconstruction requires large strains to be present in these areas. Based on the deficits of the plan view reconstruction, one may actually argue that it furnishes evidence against a circular origin of the SIC.

In order to test whether the curvilinear geometry of the SIC resulted from post-emplacement folding, Cowan (1999) investigated the internal state of strain of the SIC in the North and East Ranges, as well as in the NE-lobe of the SIC. In case of the SIC deformed by orogenic folding, strains formed in the solid-state are expected to be highest where the curvature of the SIC is maximal, i.e. in the NE-lobe. Correlation of AMS and petrofabrics indicate the preservation of a magmatic foliation in the norite and gabbro that is concordant to the base of the SIC and a down-dip magmatic lineation. By contrast, the granophyre is characterized by a lineated fabric, which is interpreted as a crystallization fabric that is orthogonal to the contact with the Onaping Formation (Cowan et al. 1999). Despite the evidence for some crystal-plastic deformation in the NE lobe, the preservation of wall-orthogonal fabrics in the granophyre in this lobe suggests low levels of pervasive ductile overprint on the centimeter-scale. Thus, Cowan (1999) and Cowan et al. (1999) concluded that the Main Mass of the SIC and its lower contact must have been inclined upon solidification of the SIC and prior to solid-state orogenic deformation. Numeric modelling based on AMS fabrics indicates that such deformation did not impart significant shape change to the NE lobe (< % shortening: Cowan et al. 1999).

In summary, there is no unequivocal field-based, structural evidence that the layers of the SIC differentiated (statically) in a circular basin. Crystallization, and possibly differentiation, of the SIC may have locally occurred on host rock walls that were inclined prior to orogenic deformation. Alternatively, tilting of initially subhorizontal layers of the solidified Main Mass of the SIC occurred, without pervasive mesoscopic deformation of the SIC. Such deformation could not have affected the SIC without incorporating its Archean and Paleoproterozoic host rocks. Therefore, pre- and post-impact tectonism affecting the host rocks, along with their implications for impact-induced deformation and possible mechanisms of shape change of the SIC, respectively, are addressed next.

PRE-IMPACT TECTONISM

There is ample evidence for pre-impact tectonometamorphic and magmatic activity in the Sudbury area (e.g., Rousell et al. 1997; Rousell et al. 2002). Most important with respect to understanding the vertical displacement of rock within the impact structure are the granulite-facies metamorphic rocks of the Levack Gneiss Complex (Fig. 1). These granulites formed at 2.71 Ga and were affected by amphibolite-facies metamorphism at around 2.65 Ga (Krogh et al. 1984). This metamorphic overprint resulted from exhumation of the granulites from their depth of crystallization between 21–28 km to shallow crustal levels of about 5–11 km (James et al. 1992), the depth at which the granulites recrystallized in the contact-metamorphic aureole of the North Range (Lakomy 1990).

In contrast to the South Range, Huronian cover rocks do not underlie the North Range of the SIC (Fig. 1). This may indicate differential exhumation of Archean basement rocks with respect to exposed Huronian cover rocks prior to

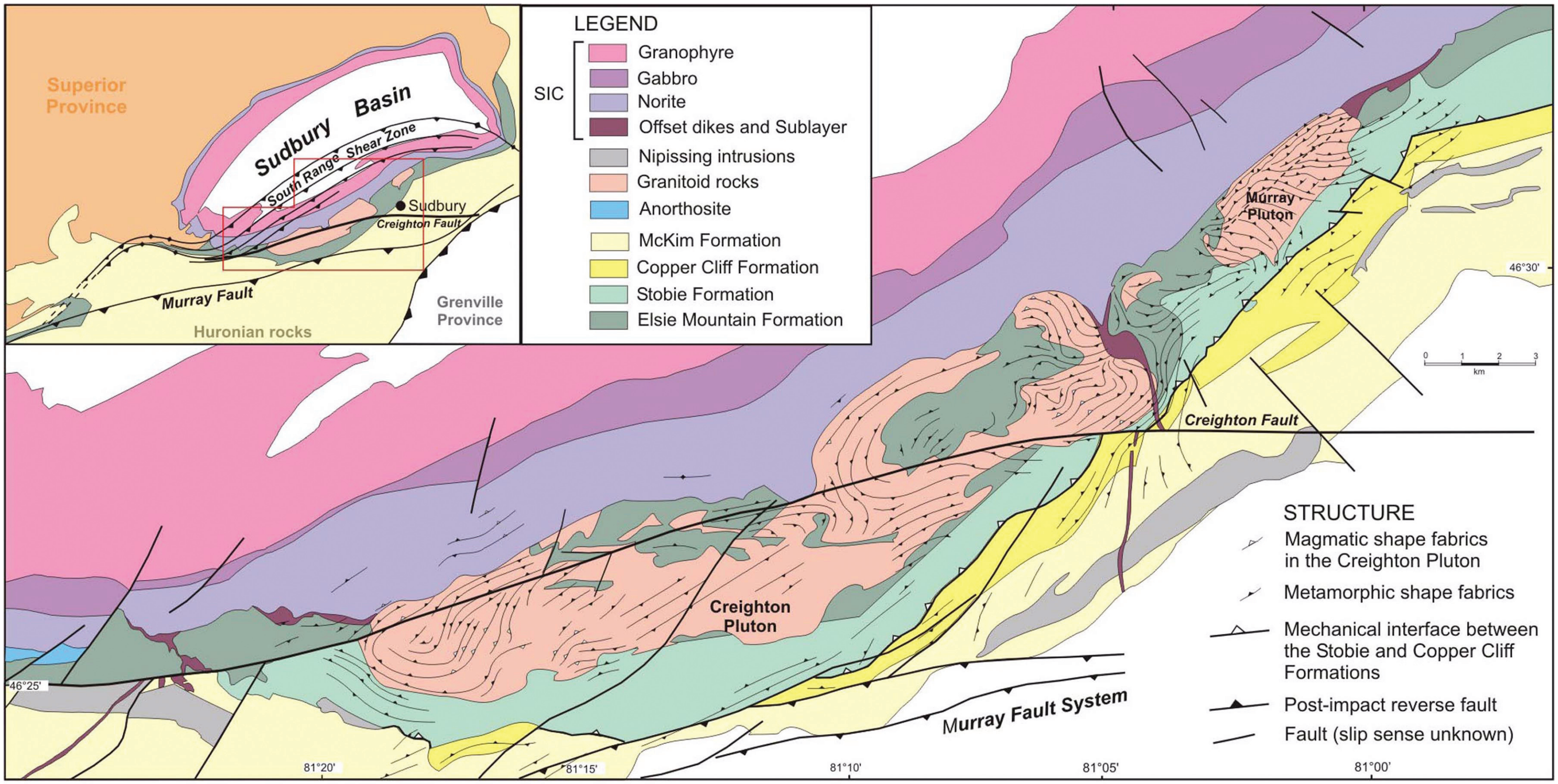

Fig. 4. Structural map of Huronian rocks underlying the South Range of the Sudbury Igneous Complex. Note the concordance of foliation trajectories at the interface between mafic and felsic metavolcanic rocks, the Stobie and Copper Cliff Formations, respectively. Mineral fabrics formed under amphibolite-facies metamorphic conditions north of this interface are juxtaposed against fabrics formed under greenschist-facies metamorphic conditions. Syntectonic emplacement of the 2.3 Ga Creighton and 2.4 Ga Murray plutons appears to be associated with amphibolite-facies metamorphism and displacement on the interface. Thus, the interface formed either as a north-dipping reverse fault or a south-dipping normal fault that rotated into its present orientation prior to the meteorite impact.

emplacement of the SIC. Cooke (1948) explained this configuration with a large NNE-trending dome structure cored by granulites of the Levack Gneiss Complex and rimmed by relics of synformal Huronian rocks about 15 km to the north and west of the SIC and overturned south-facing Huronian strata underlying the South Range (Fig. 1, see also Card et al. 1984). Formation of this structural dome, the Sudbury Basin is apparently superimposed on, is either attributed to horizontal crustal extension during Paleoproterozoic continental rifting (Rousell et al. 1997; Rousell and Long 1998) or a subsequent phase of transpressive deformation (Riller and Schwerdtner 1997; Riller et al. 1999) during the Blezardian orogeny about 2.4–2.2 Ga ago (Stockwell 1982).

The latter interpretation is based on the kinematics of deformation associated with mineral fabrics formed under amphibolite-facies metamorphic conditions in mafic Huronian metavolcanic rocks truncated by the Main Mass of the South Range SIC (Fig. 4). More specifically, these fabrics formed by fold-detachment of mafic metavolcanic strata from the interface with younger, mechanically more competent, fesic volcanic rocks during intrusion of the Creighton and Murray Plutons (Riller and Schwerdtner 1997). South of this north-dipping interface, which was already identified by Cooke (1948), Huronian rocks are generally characterized by greenschist-metamorphic mineral assemblages. This may indicate that the interface formed as a north-dipping reverse fault along which granitoid melts ascended into the southern flank of the structural dome (Riller and Schwerdtner 1997). Alternatively, the mechanical interface between mafic and felsic volcanic rocks formed as a south-dipping normal fault that rotated into its present orientation, prior to meteorite impact (Fig. 5).

There is a conspicuous structural continuity of subvertical Huronian strata at the contact with Archean basement rocks from east of the city of Sudbury to west of the town of Cutler (Fig. 1). As in the Sudbury area, Huronian strata in this belt are characterized by subvertical strain fabrics, which formed under amphibolite-facies metamorphic conditions (Card 1978). It is likely that this metamorphism, intrusion of about 2.4-Ga-old granitoid rocks, and the strain concentration in this belt are all related to the same tectonic process, i.e., normal-fault detachment of basal Huronian cover rocks from Archean basement rocks (Fig. 5a). Deposition of younger Huronian rocks in northeasterly trending paleovalleys north of the Main Mass of the SIC (Rousell and Long 1998; Long 2004) may well be related to such horizontal N-S crustal extension. Continued crustal extension may have caused detachment of these rocks from the underlying Archean basement (Rousell et al. 1997; Rousell and Long 1998).

Horizontal crustal extension can account for local uplift of higher-grade metamorphic rocks in the footwall of generally south-dipping detachment faults and their juxtaposition against lower-grade metamorphic rocks also at second-order normal faults (Fig. 5b). Normal faults may have also facilitated emplacement of the Creighton and Murray Plutons. As a result of N-S extension, Archean basement rocks formed elongate structural domes such as the "Chiblow-Baldwin Arch" (Card and Hutchinson 1972), that is much larger than the deformed relics of the superimposed SIC (Fig. 1). Progressive horizontal extension followed by orogenic shortening prior to meteorite impact may have led to steepening of Huronian strata and tightening of respective folds on either side of this basement-cored anticline (Riller and Schwerdtner 1997, Fig. 5c; Mungall and Hanley 2004).

Despite the evidence for the formation of the granulite-cored Paleoproterozoic structural dome in the Sudbury area, impact-induced exhumation of the Levack Gneiss Complex is also invoked. In impact models, exhumation of this complex is attributed chiefly to structural uplift during the modification of the transient cavity close to the center of the impact structure (Grieve et al. 1991, Deutsch et al. 1995). Numerical modelling of crater formation at Sudbury suggests that maximum rock uplift occurs close to the center of the impact structure and is on the order of 20 km (Ivanov and Deutsch 1999). Although this is within the range of absolute exhumation magnitudes based on paleobarometric estimates of the Levack Gneiss Complex (10 km to 23 km; James et al. 1992), it is in conflict with the results of a feldspar clouding analysis of the 2.45-Ga-old Matachewan dikes (Siddorn and Halls 2002). This analysis, as well as the unstrained nature of these dikes, suggests that exhumation of the Levack Gneiss Complex and formation of the structural dome occurred largely prior to emplacement of the Matachewan dikes. This has important consequences for estimating maximum stratigraphic rock uplift (SU) during cratering and the final diameter (D) of the Sudbury impact structure. Based on the empirical relationship between the two (SU:D is about 1:10, Melosh and Ivanov 1999), Grieve and Therriault (2000) may have overestimated the size of this impact structure (D: minimum of 250–280 km assuming a SU of 26–38 km) as the component of orogenic deformation of rock uplift is not considered in their analysis. Partial exhumation of the Levack Gneiss Complex prior to impact explains also the presence of higher-grade metamorphic rocks underlying the North Range, despite the fact that rocks underlying the South Range were uplifted on the South Range Shear Zone, with respect to the North Range by about 10 to 15 km (Shanks and Schwerdtner 1991b).

The foregoing account of pre-impact orogenic processes at Sudbury shows that geological characteristics, such as exhumed lower crustal rocks and overturned strata, which fit large-meteorite impact models well, nonetheless need to be regarded with caution. Clearly, pre-impact orogenic activity imposes major difficulties in reconstructing particle paths during crater formation. This is chiefly because at the time of impact, the crustal structure did not contain paleohorizontal

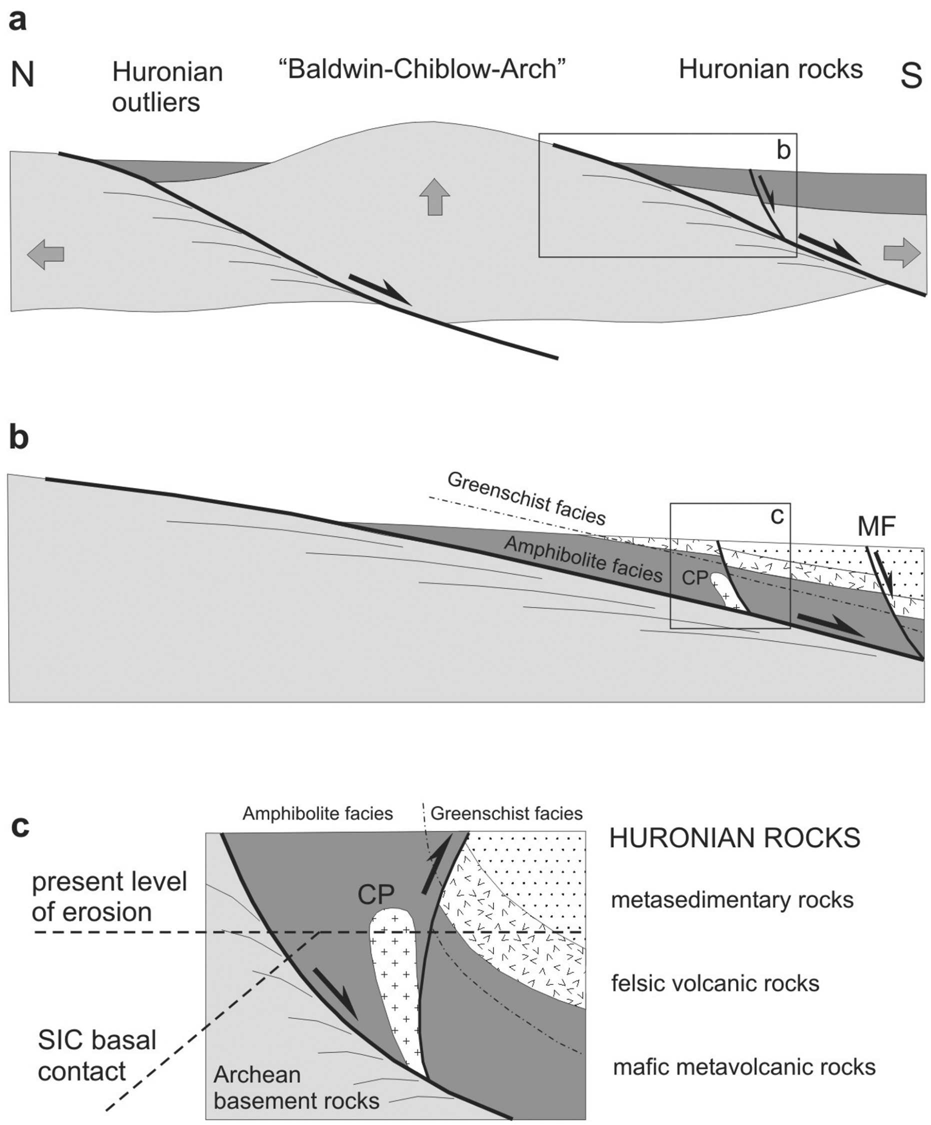

Fig. 5. Cartoon showing the possible effects of Paleoproterozoic horizontal extension in the Sudbury area. a) Formation of the elongate structural dome, the "Baldwin-Chiblow-Arch" (Card and Hutchinson 1972; cf. Fig. 1), by uplift of Archean basement rocks in the footwall of a proposed south-dipping detachment fault. Huronian rocks are deposited and subsequently detached on either side of the structural dome cored by Archean granulite-facies metamorphic basement rocks. b) Horizontal extension affected Huronian rocks, which underlie the South Range of the Sudbury igneous complex today. Subsidiary steeply south-dipping normal faults, such as the Murray Fault (MF), juxtaposes amphibolite-facies metamorphic Huronian strata, located north of this fault, against greenschist-facies metamorphic rocks, located south of the fault. This mechanical discontinuity may well have facilitated emplacement of granitoid magma. c) Geometry of structures shown in (b) after tilting and shortening prior to meteorite impact. CP in (b) and (c) denotes the Creighton Pluton (cf. Fig. 4).

surfaces or equilibrated paleoisotherms, which can be correlated readily to, for example, numerically modelled equivalents. Without a better understanding of the crustal structure at the time of impact, it is impossible to successfully correlate numerical models with observed pre-impact geological marker surfaces. One way to remedy this situation is to analyze structures formed by post-impact deformation, which can be used to reconstruct the pre-impact crustal architecture and impact-induced deformation.

POST-IMPACT TECTONISM

There is little doubt that the Sudbury impact structure was geometrically modified during the Penokean (1.89–1.83 Ga: Sims et al. 1989) and the Grenville orogenies (Card et al. 1984). Recent geochronological evidence suggests that the South Range Shear Zone was active during the Mazatzal orogeny 1.7–1.6 Ga ago (Bailey et al. 2004). Individual thrust surfaces of the shear zone were seismically imaged by the

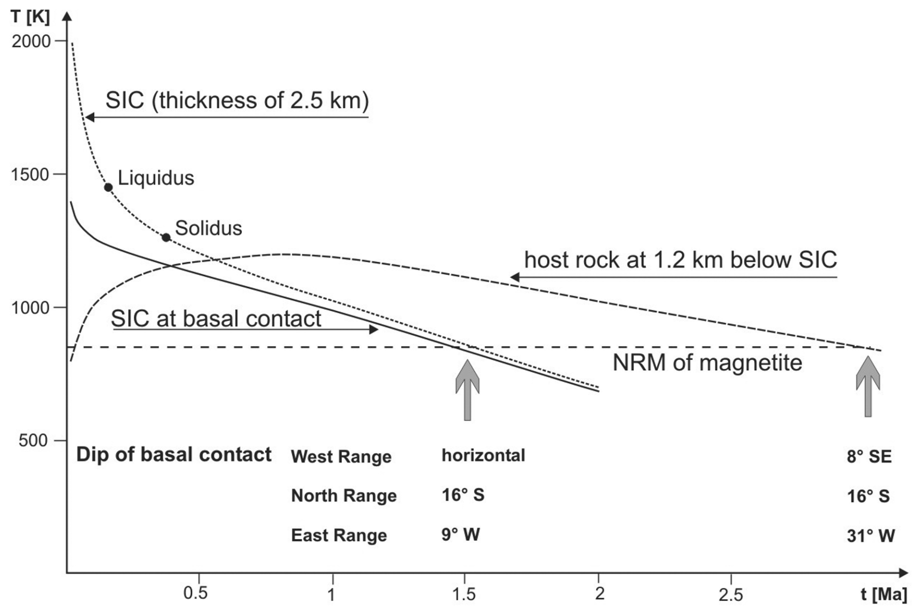

Fig. 6. Schematic temperature-time evolution of the Sudbury Igneous Complex and its contact-metamorphosed target rocks, after Ivanov and Deutsch (1999). The orientation of the lower contact of the SIC for portions of the North and East Ranges is based on paleomagnetic fold tests by Szabo (2002) and Morris (1984). NRM: natural remanent magnetization. Further explanation is given in the text.

Lithoprobe project to a depth of ~7 km below the South Range (Fig. 2b: Milkereit et al. 1992, 1994; Wu et al. 1994). Displacement of the South Range and its Huronian host rocks on this shear zone probably exceeds 8 km (Shanks and Schwerdtner 1991a). However, it is unknown whether the shear zone accommodated ductile strain as a result of large-scale buckle folding of the SIC (Cowan and Schwerdtner 1994) or merely translated the South Range toward the NW following an initial phase of folding (Riller et al. 1998).

Non-cylindrical folding of the Main Mass of the SIC along with its host rocks during Penokean deformation is advocated by paleomagnetists. Szabo (2002) determined the paleomagnetic poles in pseudotachylite of the Levack Gneiss Complex within the thermal aureole of the SIC in the North and East Ranges in order to estimate rotation magnitude of these rocks since the acquisition of the natural remanent magnetization (NRM) in magnetite, the dominant carrier of NRM in these rocks. As a result of the uncertain age of NRM acquisition, metamorphic pseudotachylite rocks underlying the South Range were not included in this analysis. Assuming that the North and East Ranges rotated around axes parallel to their strike, the Ranges were rotated backwards until their paleomagnetic poles matched in orientation. This rather crude fold test resulted in orientations of the basal SIC contact of the western North Range, the northeastern North Range, and the East Range (Fig. 2a) of 8° SE, 16° S, and 31° W, respectively. By contrast, the same fold test applied to paleomagnetic poles from the SIC (Morris 1984) gave different orientations for these Ranges, i.e., a horizontal position for the western North Range, 16° S for the northeastern North Range and 9° W for the East Range.

The difference in rotation magnitude between the SIC and its adjacent host rocks within each of the three sectors analyzed may appear incompatible with rigid rotation of the SIC along with its host rocks. Considering, however, the temperature evolution of the North and East Ranges of the SIC and its contact-metamorphosed target rocks (Ivanov and Deutsch 1999), progressive rotation of individual sections of the SIC basal contact becomes apparent. This is because the temperature of NRM acquisition in the SIC and its metamorphosed host rocks was reached at different times, i.e., about 1.5 Ma and 3 Ma after impact, respectively (Fig. 6). Consequently, rocks of the thermal aureole rotated more until acquisition of NRM for magnetite than did the SIC. Provided that the SIC rotated everywhere from the horizontal around axes that are parallel to the local strike of its lower contacts, progressive rotation of the basal contact of the SIC differs for individual sectors. Taking into account possible large errors in the calculated paleomagnetic vectors and cooling curves, this interpretation of paleomagnetic data suggests that non-cylindrical folding of the SIC commenced within less than three million years after impact, i.e., during the Penokean Orogeny.

The extent to which Archean and Paleoproterozoic target rocks were affected by post-impact rotation beyond the contact-metamorphic aureole is not accurately known. The lack of pervasive mesoscopic strain fabrics in the North Range of the SIC and its Archean host rocks may indicate that these rocks were tilted coherently during Penokean tectonism. This is corroborated by the structural continuity of the lower contact of the North Range of the SIC in seismic images (Milkereit et al. 1992; but see also Snyder et al. 2002) and suggests a maximum rigid-body rotation of 35° toward the SE. However, post-impact folding of the North Range and its Archean host rocks accomplished by larger-scale slip cannot be ruled out.

Clast-rich pseudotachylite and 1.85 Ga quartz-diorite dikes in Huronian rocks underlying the South Range are affected pervasively by subvertical planar and linear mineral fabrics (Fig. 3b, 3e), which formed under greenschist-facies metamorphic conditions (Riller and Schwerdtner 1997). This attests to NW-SE directed horizontal shortening and vertical stretching of Huronian rocks and is corroborated by the curvi-linear trace of the NW-SE striking Copper-Cliff dike at surface, in contrast to the straight NE-SW striking Worthington dike (Fig. 2a). In general, quartz-diorite dikes disposed at low angels with respect to NW-SE shortening, e.g., the Copper-Cliff and the Foy dikes, were horizontally buckled and displaced. By contrast, dikes striking at high angles to the shortening direction, e.g., the Worthington and Whistle-Parkin dikes, may have been stretched and rotated towards the direction of horizontal extension (NE-SW).

Provided that the Worthington dike was emplaced subvertically, the Huronian rocks may have been tilted by about 20° toward the NW during Penokean deformation. Evidence for minor buckling of the SIC, including shortening and/or rotation of target rocks is provided by the fanning pattern of foliation surfaces in Archean rocks immediately to the west of the SIC (Riller et al. 1998). Whether such foliation pattern is present in Archean rocks near the NE-lobe of the SIC remains to be determined. Therefore, the mechanism of tilting of the East Range during Penokean (or younger) tectonism and, thus, to what extent the Sudbury impact structure experienced non-cylindrical buckle folding are uncertain.

The principal directions of strain during late Penokean deformation are well constrained for the southern portion of the impact structure (Cowan et al. 1999). Inversion of slip directions on brittle-ductile shear faults shows that, except at the transpressional SE-lobe, shortening is directed uniformly NW-SE. This agrees well with incipient buckle folding of the impact structure (Riller et al. 1998) and with the sense of displacement on the South Range Shear Zone, as determined from asymmetric fabric geometry (Shanks and Schwerdtner 1991a). Overall NW-SE directed shortening is, however, incompatible with non-cylindrical buckle folding of the impact structure at a late stage of orogenic deformation. Since brittle deformation occurs towards the end of orogenesis, the non-cylindrical geometry of the Main Mass of the SIC must have been acquired at rather an early stage in post-impact orogenic deformation. This would agree with the interpretation of paleomagnetic data.

IMPACT-INDUCED DEFORMATION

Shatter cones are well developed in fine-grained metasedimentary rocks of the Huronian Supergroup close to the Main Mass of the SIC. Ideally, shatter cone apices should point to ground zero of the impact. This prompted Dressler (1984b) to compile the orientation of cones apices to determine post-impact deformation. It turned out, however, that there is a considerable number of locations at which distinct shatter cone populations display variable orientations of cone apices. A more rigorous geometrical or statistical treatment of shatter cone orientations may be necessary to elucidate some of the kinematic effects of post-impact deformation at Sudbury (Gibson and Spray 1998).

The most pervasive impact-induced structures found in Archean and Proterozoic target rocks are zones of clast-rich pseudotachylite. As far as exposure permits judgement, these are dominantly planar and disposed concentrically with respect to the SIC although, as pointed out earlier, their spatial concentration in distinct rings remains to be ascertained. The distinction between, and spatial distribution of, pseudotachylite induced by shock, i.e., S-type, and frictional sliding, i.e., E-type, (Müller-Mohr 1992; Spray et al. 2004) is also unknown. Zones of clast-rich pseudotachylite vary in thickness from centimeters to hundreds of metres and are generally found at lithological contacts or faults. Thus, pseudotachylite may have formed at mechanical discontinuities upon passage of the shock wave, i.e., as S-type pseudotachylite (Spray 1998). This is corroborated by the abundance of clast-rich pseudotachylite in layered rocks, such as the Huronian, and respective paucity of pseudotachylite in lithologically more uniform granitoid rocks (Lieger 2005).

A different mechanism of pseudotachylite formation has been suggested, however, for the most prominent pseudotachylite zone, the South Range Breccia Belt south of the Murray Pluton. In places, this zone is several hundred metres thick and can be traced at surface for about 20 km from the SIC east of the Murray Pluton to the Copper Cliff dike (Fig. 2a). Spray (1997) and Spray et al. (2004) suggested that it continues south of the Creighton Pluton and merges with the SIC west of the pluton, which would make this zone about 45 km long at the present erosion surface. They further proposed that the zone has listric geometry and is inclined toward the SIC. Formation of the zone as the trace of a "superfault" has been explained by catastrophic gravitational collapse of the inner ring of the impact structure (Spray 1997) or due to central uplift formation (Spray et al. 2004), whereby clast-rich pseudotachylite formed by frictional sliding and shear-induced melting on the listric interfaces at ultra-high strain rates (Spray 1997; Scott and Spray 2000). Displacements on the faults are apparently on the order of kilometers and constrained by magnetic lineations, interpreted to represent the direction of flow during E-type pseudotachylite formation (Scott and Spray 1999).

Although E-type pseudotachylite may well be generated during cavity modification at large impact structures, the proposed mechanisms of their formation at Sudbury, and for the South Range Breccia Belt in particular, remain controversial. First, there is no evidence for any displaced

Table 1. Possible modes of *Offset Dike* formation and respective time frames of emplacement after impact.

Mode	Cratering stage (time frame)	Reference
Dilation during formation of the transient cavity, lateral and/or downward injection of melt into target rock	Excavation (minutes)	Grant and Bite (1984), Murphy and Spray (2002), Lightfoot and Farrow (2002)
Dilation in rebounding central uplift	Incipient modification (minutes)	Tuchscherer and Spray (2002) Wood and Spray (1998)
Transfer fault separating blocks of collapsing crater wall	Modification (days to years)	Scott and Benn (2002) Scott and Spray (1999, 2000)
Isostatically-driven floor fracturing	Post-cratering modification (ten thousand years)	Wichmann and Schultz (1993)
Dilation due to cooling and contraction of target rock below the melt sheet	Post-cratering modification (tens of thousands of years)	This work
Tectonism following crater formation	Post-cratering modification (tens to hundreds of thousands of years)	Therriault et al. (2002)

lithological contacts or other pre-impact fabrics where they are cut by the clast-rich pseudotachylite zone (e.g., Dressler 1984a). Second, the continuity of this breccia zone west of Copper Cliff has not yet been confirmed by geological mapping (e.g., Lieger 2005), which calls into question the existence of a continuous clast-rich pseudotachylite zone 45 km in length in Huronian rocks. Moreover, pseudotachylite in the Huronian rocks is pervasively affected by post-impact solid-state strain, in particular at the margins of pseudotachylite zones. Thus, it is unlikely that the magnetic lineation represents a primary flow fabric. Finally, frictional melting, intrinsic to E-type pseudotachylite formation, reduces the shear strength of a given fault dramatically. Growth of the width of a fault, i.e., the pseudotachylite zone, to hundreds of meters appears physically impossible. In the opinion of the author, the proposed mechanisms of E-type pseudotachylite formation, by which the South Range Breccia Belt apparently formed, are implausible.

Similar to the superfault hypothesis by Spray (1997), Mungall and Hanley (2004) explain the preservation of Huronian outliers (Fig. 2) by catastrophic normal-fault displacement on the order of 10 km juxtaposing Huronian cover against Archean basement rocks. Their interpretation is largely based on the presence of pseudotachylite decorating interfaces between Huronian cover and Archean basement rocks, as well as stratigraphic evidence for structural omission of Huronian strata next to the respective interfaces. Mungall and Hanley (2004) assume that the pseudotachylite on these interfaces is of E-type and do not discuss the possibility of S-type pseudotachylite formation. S-type pseudotachylite forms preferentially at lithological interfaces characterized by pronounced impedance contrasts (Spray 1998). Lieger (2005) showed that interfaces between granite and metasedimentary rocks in the Sudbury area are prone to formation of this pseudotachylite variant. Thus, the mere presence of pseudotachylite on any interface does not unequivocally indicate that this interface formed by large-magnitude slip as a consequence of the impact. This leaves open the possibility that large normal-fault displacements associated with Huronian outliers preceded the impact (Fig. 5a) and that S-type pseudotachylite formed on the respective pre-impact faults. The foregoing notions suggest also that the mechanism by which gravitationally driven crater modification occurred at the crustal level exposed at Sudbury remains to be ascertained.

FORMATION OF OFFSET DIKES

The formation of Offset Dikes of the Sudbury impact structure constitutes a key topic in understanding the interplay between crater formation and migration of impact melt. Table 1 summarizes the dominant modes that have been suggested to account for the formation of Offset Dikes and their respective time frames within which they occurred after impact. The foregoing notion that the composition of Offset Dikes is akin to that of the bulk Main Mass of the SIC (Lightfoot et al. 1997; Ariskin et al. 1999; Lightfoot and Farrow 2002) implies that emplacement of the dikes must have occurred prior to differentiation and, consequently, solidification of the Main Mass. Solidification of the Main Mass occurred within a few hundreds of thousands of years (Ivanov and Deutsch 1999). This excludes modes of Offset Dike formation operating on time scales on the order of millions of years, i.e., those assisted by orogenic deformation.

Emplacement of melt in Offset Dikes within minutes after the impact is also unlikely as such emplacement must have been accomplished violently by turbulent melt flow. Offset Dikes are generally characterized by two quartz-diorite phases. One is rich in host-rock fragments and sulfide blebs, whereas the other lacks both types of inclusions (e.g., Lightfoot and Farrow 2002). The inclusion-rich phase is mostly found in the center of the dike (e.g., Lightfoot and Farrow 2002; Tuchscherer and Spray 2002), where it is mingled with the inclusion-poor quartz-diorite phase (Fig. 3d). Therefore, mingling must have occurred while the inclusion-poor phase still was viscous. Evidence for mingling of both magma phases suggests non-turbulent emplacement of quartz-diorite magma. Moreover, segregation of sulfides from the silicate melt did not occur within minutes of the impact. Therefore, Offset Dikes could neither have been emplaced during formation of the transient cavity nor during gravitationally driven crater modification of the Sudbury impact structure.

Based on the superfault hypothesis of Spray (1997) and analysis of AMS fabrics of the Copper Cliff Dike and the South Range Breccia Belt, Scott and Benn (2002) and Scott and Spray (1999, 2000) considered Offset Dikes and concentric dikes (i.e., the Frood Stobie) as impact-melt-filled discontinuities separating crater wall segments, as they collapsed into the impact melt pool. The direction of injection of impact melt between collapsing crater wall segments was inferred from subvertical magnetic lineations in the quartz-diorite dike rocks. Again, the structural continuity of pre-impact lithological boundaries across the margins of the Copper Cliff dike (Dressler 1984a) casts doubt on such an origin for Offset Dikes. Also, the centro-symmetric strain field imposed by such slumping of collapsing crater wall terraces between radial discontinuities would be unfavorable for melt propagation through the respective discontinuities. This is because deformation imposed by centro-symmetric collapse on the discontinuities is characterized by transpression (Kenkmann and von Dalwigk 2000). Furthermore, the interpretation of magnetic lineations representing magmatic lineations is unlikely, especially since pyrrhotite, a mineral which recrystallizes under greenschist-facies metamorphic conditions (Schwartz 1973), has been identified as the dominant carrier of the AMS fabric. Finally, superposition of two planar fabrics, e.g., a relic magmatic and a metamorphic foliation in, for example, the Worthington Offset Dike (Fig. 3e) readily generates a prolate magnetic fabric. This has been documented for AMS fabrics in the Murray Pluton (Riller et al. 1996) and may as well account for the subvertical lineation in quartz diorite dikes.

The foregoing discussion implies that generation of Offset Dikes more likely occurred within tens of thousands of years after impact. In the view of the author, the appropriate time frame and strain field, i.e., radial and tangential dilation, for Offset Dike formation, is provided by the decrease of the post-impact temperature in, and late-stage crustal readjustment accomplished by gravitational spreading of, target rocks below the Main Mass of the SIC. Cooling of this thermal anomaly below the crater, both induced by shock-heating as well as by uplift of hotter mid- to lower-crustal rocks upon modification of the transient cavity, may have caused the target rocks to thermally contract and, thus, to fracture. As a result, radial and concentric fractures formed preferentially in target rocks already mechanically weakened by clast-rich pseudotachylite and were filled from above with largely undifferentiated impact melt. Such fracturing explains the conspicuous hexagonal geometry, evident by angles of approximately 60° between the strike of the three major directions of Offset Dikes (Fig. 2a: NE-SW, NNW-SSE and WNW-ESE) and is typical for jointing in cooling rock masses. Thermal corrosion of rocks at the crater floor induced by the superheated impact melt can account for a number of observations made in previous studies (e.g., Grant and Bite 1984; Lightfoot and Farrow 2002; Tuchscherer and Spray 2002). For example, it lead to fragmentation of the crater floor and widening of dikes, i.e., generation of "embayments", near the base of the Main Mass. Thermally corroded target rock fragments were entrained along with segregated sulfide blebs in the impact melt, which, at a later stage, sank into the fractures where it mingled with melt devoid of fragments and sulfides. Mingling between individual impact melt phases (Fig. 3d) occurred preferably in the centers of Offset Dikes as these widened incrementally upon cooling of the host rocks.

The strain field caused by cooling of the impact-induced thermal anomaly below the crater may be similar to the one induced by isostatically driven crater floor fracturing (e.g., Wichman and Schultz 1993). This mechanism is, however, not regarded as an important consequence of a large meteorite impact (Melosh and Ivanov 1999) and is expected to operate on a much shorter time scale (about ten thousand years) than equilibration of the thermal anomaly below the crater. Nonetheless, it remains to be explored whether isostatically driven deformation would result in the observed geological characteristics displayed by the Offset Dikes.

UNSOLVED STRUCTURAL PROBLEMS

The foregoing account has discussed the effects of pre-impact and post-impact orogenic deformation and how they influence estimates of parameters such as magnitude of structural uplift, tilting of pre-impact (Huronian) strata and displacement on major discontinuities at Sudbury. Major uncertainty remains, however, as to (1) the cause for the steeply dipping SIC in the East Range, (2) whether the Sudbury impact structure is a peak-ring or a multi-ring impact basin, and (3) the mechanisms accomplishing rock flow during crater modification.

1. There is little doubt that the Main Mass of the SIC represents a folded impact-melt sheet. Since the NE- and SE-lobes of this melt sheet are synforms (the axes of which plunge toward the center of the Sudbury Basin) the East Range represents likely a westward plunging antiform, which accounts for the conspicuous curvature of the East Range at surface (Fig. 2a). The mechanism of large-magnitude tilting of the melt sheet in the East Range and distortion in the NE- and SE-lobes remain, however, to be elucidated. Cowan et al. (1999) showed that less than 5% of ductile folding strain was imparted to the rocks of the NE lobe. Localized brittle or brittle-ductile deformation may, therefore, have accomplished the shape change of the Main Mass and its underlying granitoid basement rocks, while maintaining structural continuity of the melt sheet on the map scale. If so, the scale and kinematics of the respective discontinuities need to be identified.

2. Based on the circumferential distribution of subvertical Huronian strata around and their direction of

stratigraphic younging away from the exposed SIC, it is conceivable that exposed Archean rocks next to the North and East Ranges constitute remnants of the peak ring (ring 1 in Fig. 1). Peak rings are characterized by vertical thickening of rock during collapse of the transient cavity and the central uplift. Consequently, concentric folds and reverse faults are expected to be present in this zone. In this regard, the mechanism of circumferential tilting of Huronian strata along with the underlying Archean basement rocks is also of importance. More specifically, it needs to be discerned to what extent the steep attitude of Huronian rocks is due to pre-impact orogenic doming or peak-ring formation. This can be accomplished by examining the kinematics and deformation mechanisms associated with discontinuities that accomplished tilting of Huronian strata.

Prompted by the identification of a concentric lineament pattern by Butler (1994), Spray et al. (2004) proposed the existence of three other rings (rings 2–4 in Fig. 1). Two of these rings are apparently associated with E-type pseudotachylite (Thompson and Spray 1994; Spray and Thompson 1995) and interpreted to delineate the traces of normal fault zones formed during cavity modification. Although this is an intriguing hypothesis, there is little conclusive evidence as to the genetic relationship between proposed rings (faults), lineaments, and pseudotachylite. This is largely because of uncertainty as to (1) the origin of lineaments (Butler [1994] considers also orogenic doming in this respect), (2) spatial association between pseudotachylite zones and proposed rings, (3) whether all of the pseudotachylite is of E-type, and (4) the lack of evidence for normal faulting next to proposed rings. Field structural analyses complemented by remote sensing techniques may help to clarify these issues, including the role of pre-existing crustal discontinuities during cratering and thus whether or not the impact structure was originally a multi-ring basin.

3. There is a remarkable structural continuity of pre-impact lithological units, most importantly steeply northward dipping (overturned) Huronian strata at the interface with Archean basement rocks between Sudbury and the town of Cutler (Fig. 1). This continuity in Huronian cover rocks persists well beyond the modelled maximum diameter of the transient cavity (Ivanov and Deutsch 1999). At present, this is difficult to reconcile with impact-induced differential rotation of Huronian strata within the transient cavity, unless post-impact deformation rotated the overturned strata back to their pre-impact configuration. Moreover, generation and collapse of the transient cavity did not result in large offsets of Huronian strata, not even at major clast-rich pseudotachylite zones, at the present erosion level. Thus, closure of the transient cavity seems to have been achieved by coherent rock flow, i.e., continuous deformation operating at least on the scale of metres, rather than by the generation of discontinuity surfaces resulting in several hundred meters or even kilometers of differential displacement of rock.

It should also be noted that evidence for the possible existence of acoustic fluidization (Melosh 1989; Melosh and Ivanov 1999) has not been found at Sudbury. Also, the mechanism(s) that produced the large thickness of clast-rich pseudotachylite zones remains to be identified. Single-slip events are unlikely to account for the observed thicknesses of pseudotachylite, as shear-induced melting on a shear fracture would reduce its friction and thus inhibit further melting by this process (Melosh 2005). As high slip rates over a longer time interval seem necessary to produce thick pseudotachylite zones, it is conceivable that the localized generation of clast-rich pseudotachylite is genetically related to acoustic fluidization. Microstructural analyses on rocks underlying the North and East Ranges of the SIC may help to constrain better the modes of rock flow during large-impact cratering.

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