Invited review article

Igneous rocks formed by hypervelocity impact


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Abstract

Igneous rocks are the primary building blocks of planetary crusts. Most igneous rocks originate via decompression melting and/or wet melting of protolith lithologies within planetary interiors and their classification and compositional, petrographic, and textural characteristics, are well-studied. As our exploration of the Solar System continues, so too does the inventory of intrusive and extrusive igneous rocks, settings, and processes. The results of planetary exploration have also clearly demonstrated that impact cratering is a ubiquitous geological process that has affected, and will continue to affect, all planetary objects with a solid surface, whether that be rock or ice.

It is now recognized that the production of igneous rocks is a fundamental outcome of hypervelocity impact. The goal of this review is to provide an up-to-date synthesis of our knowledge and understanding of igneous rocks formed by hypervelocity impact. Following a brief overview of the basics of the impact process, we describe how and why melts are generated during impact events and how impact melting differs from endogenic igneous processes. While the process may differ, we show that the products of hypervelocity impact can share close similarities with volcanic and shallow intrusive igneous rocks of endogenic origin. Such impact melt rocks, as they are termed, can display lobate margins and cooling cracks, columnar joints and at the hand specimen and microscopic scale, such rocks can display mineral textures that are typical of volcanic rocks, such as quench crystallites, ophitic, porphyritic, as well as features such as vesicles, flow textures, and so on. Historically, these similarities led to the misidentification of some igneous rocks now known to be impact melt rocks as being of endogenic origin.

This raises the question as to how to distinguish between an impact versus an endogenic origin for igneous-like rocks on other planetary bodies where fieldwork and sample analysis may not be possible and all that may be available is remote sensing data. While the interpretation of some impact melt rocks may be relatively straightforward (e.g., for clast-rich varieties and those with clear projectile contamination) we conclude that distinguishing between impact and endogenic igneous rocks is a non-trivial task that ultimately may require sample investigation and analysis to be conducted. Caution is, therefore, urged in the interpretation of igneous rocks on planetary surfaces.

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1. Introduction

Igneous rocks are one of the three fundamental lithological classes of rocks. Derived from the Greek work “ignis”, meaning “fire”, igneous rocks are defined as having “solidified from molten or partially molten material, i.e., a “magma”. Although the current definition of “magma” states that such molten materials are generated within the “Earth” (e.g., Neuendorf et al., 2005), the implicit assumption is that it also refers to similar circumstances on other bodies in the Solar System. Igneous rocks are the precursors for the other lithological classes (i.e., sedimentary and metamorphic) of rocks. On planetary bodies with little subsequent geological activity, they remain the dominant rock type (Wilson, 2009) and on Earth and other active planetary objects, igneous processes continue to the present-day. The classification of igneous rocks and their compositional, petrographic, and textural characteristics, is well-studied and documented (e.g., Le Bas and Streckeisen, 1991). The formation of magmas by decompression melting and/or wet melting of protolith lithologies within planetary interiors are relatively well-understood physical processes. The relative importance of these processes, however, is a matter of some debate with regard to the formation of some igneous rocks on the Earth (e.g., Foulger, 2010).

In addition to endogenic processes, hypervelocity impact events also result in physical conditions that can melt substantial volumes of protolith rocks. Once solidified, these impact-generated melts, for all intents and purposes, satisfy the above definition for igneous rocks. Knowledge of such “igneous” rocks formed by hypervelocity impact is particularly important for understanding the early history of the Earth and other planetary bodies, when impacts were substantially more frequent and larger (Neukum et al., 2001) and the genesis of impact-related igneous rocks, therefore, both more frequent and more voluminous. For example, it is now generally accepted that magma “ponds” or “oceans” were produced by accretionary impacts during the early evolution of terrestrial planets (Elkins-Tanton, 2012). The results of planetary exploration have clearly demonstrated that impact (and related melting) is a fundamental geological process that has had a major influence on the origin and early evolution of all the terrestrial planets, the Moon, asteroids, and many other rocky and icy objects in the outer Solar System. Questions remain as to whether the impact rate decayed steadily during this first half a billion years of Solar System history, in a stepwise or “sawtooth” fashion, or whether a cataclysmic so-called Late Heavy Bombardment at ~4.0 to 3.8 Ga actually occurred (e.g., Boehnke and Harrison, 2016; Bottke et al., 2012; Chapman et al., 2007; Comes et al., 2005; Norman, 2009). The only way to address this outstanding question is to date a series of large impact basins on the Moon. This requires sampling their impact melt rocks – whose radiometric clocks were reset during melting. On many of the rocky planetary objects in our Solar System, volcanism is, or was, also an important surface modification process (Wilson, 2009). This begs the question as to how to distinguish between an impact versus an endogenic origin for igneous-like rocks on Earth or any other planetary body and is the major motivation for this contribution.

The physical association of particular lithologies with a crater form, with the appropriate morphology and morphometry for its dimensions, is generally the only information used and is generally interpreted as indicative of an impact origin in the analyses of remote sensing data; although we return to this topic at the end of this contribution. This, however, is not the case for the Earth, where ground truth data are available. It is widely accepted that the unambiguous evidence for an impact origin of a particular geological feature or sample in the terrestrial environment is the recognition of diagnostic sub-solidus shock metamorphic features (e.g., shatter cones, planar deformation features (PDFs), diaplectic glass, high-pressure mineral polymorphs) or extraterrestrial chemical and/or isotopic signatures of the projectile (French and Koeberl, 2010). In some cases, such as at small impact craters, there may even be recognizable meteoritic fragments. Thus, when samples (terrestrial or extraterrestrial) are available for laboratory study, the search for shock metamorphic effects and/or geochemical tracers of the projectile can provide relatively rapid and reliable confirmation of an impact origin.

Nevertheless, impact-related igneous rocks can share close similarities with volcanic rocks (Fig. 1). In outcrop, impact melt rocks can display columnar joints (Fig. 1A, B) and at the hand specimen and microscopic scale, impact melt rocks can lack mineral and lithic clasts (Fig. 1A–D), which show evidence of sub-solidus shock, and can display mineral textures that are typical of volcanic rocks, such as quench crystallites, sub-ophitic, ophitic, porphyritic, etc., as well as features such as

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vesicles (Fig. 1C), flow textures, etc. Thus, even on Earth, it is not always a straightforward task to identify an igneous-textured rock as being of impact origin. For example, while now largely a forgotten controversy, it is notable that many igneous rocks in what are now recognized as meteorite impact craters were first proposed to be volcanic in origin. Notable Canadian examples include the Manicouagan, Mistastin, and West Clearwater Lake structures where “dyke-like” and “sheet-like masses” were described and interpreted as the products of volcanism (e.g., Bostock, 1969; Currie, 1971, 1972). Due to their association with a variety of breccias and large but shallow topographic depressions, these structures were termed “cryptobreccia” or “cryptovolcanic” structures. Similarly, in Scandinavia, enigmatic lava- and tuff-like rocks were named based on local geographic features, e.g., dellenite (Svenonius, 1888) and kärnäite (Eskola, 1927), and have since been shown to be associated with the Dellen and Lappajärvi impact structures, respectively. For almost a century, there was a similar debate about the origin of craters on the Moon, with many proponents proposing a predominantly volcanic origin (e.g., Simpson, 1966; von Bülow, 1965). Two major advancements in the knowledge and understanding of impact cratering processes and products have laid these early controversies to rest. The first was the recognition of the unique nature of shock metamorphism and its resultant irreversible changes to “shocked” rocks and minerals in the 1960s (French and Short, 1968), which enabled the unequivocal identification of the lithological and mineralogical products of hypervelocity impact. The second was the recognition that not all impacts result in the formation of bowl-shaped craters. At crater diameters greater than 2 to 4 km on Earth, for example, it was recognized that late-stage gravitational collapse of the initial crater bowl-shaped, so-called transient cavity, which is formed directly by the crating flow-field in the target, results in so-called complex craters (Dence, 1968). Complex craters possess a structurally complicated rim, a down-faulted annular trough, and a structurally uplifted central zone, and have a considerably shallower depth to diameter ratio than simple craters. Equipped with this knowledge, Dence (1971) published a landmark paper entitled simply “Impact melts” in which igneous-like rocks and glasses associated with “almost forty terrestrial structures”, including the earlier noted Manicouagan, Mistastin, and West Clearwater Lake structures, were re-interpreted as the products of hypervelocity impact. As noted above, dellenite and kärnäite were also subsequently re-interpreted as impact melt rocks; although, the term dellenite was applied for decades to volcanic rocks intermediate between rhyolite and
dacite in composition and is still defined as such; despite now being recognized as being obsolete, as is the term kârâmite.

By far the most controversial and longest running debate surrounded the origin of igneous-like rocks within the Sudbury Basin, Canada. The historical nomenclature exemplify this debate, with terms such as Onaping Tuff, Onaping Ash-Flow Sheet, Rhyolite Breccia and the Sudbury Irruptive, Lopolith, Igoneous Complex being some of the lithological names associated with Sudbury (e.g., Collins, 1936; Muir and Peredery, 1984; Stevenson, 1972). Even following the recognition of shatter cones around the Sudbury Basin (Dietz, 1964) and other shock metamorphic features (French, 1970), an impact origin for the igneous rocks within the Basin took several decades to be accepted. In addition, contemporary workers in favour of the impact origin of Sudbury, had trouble explaining the ~2–3 km thick so-called Sudbury Igoneous Complex (SIC) as being of impact origin, suggesting instead that it is the result of magmatism triggered by the impact event rather than direct impact melting of target rocks (e.g., Dietz, 1964; French, 1970; Dence, 1971). The idea of an impact-triggered but mantle-derived magma was still being proposed through the final decade of the 20th century (e.g., Naldrett and Hewins, 1984; Dressler et al., 1987; Norman, 1994; Rousell et al., 1997); although the interpretation of the SIC as a differentiated impact melt sheet (Grieve et al., 1991) now seems to be generally accepted (e.g., Dickin et al., 1996, Thériault et al., 2002, Naldrett, 2003).

In summary, the genetic controversies of the 20th century have now largely been settled and it is now widely recognized that hypervelocity impact events on Earth result in the melting of a substantial portion of the target and generate igneous-textured rocks. While there remains some debate about the effects of target lithology on the generation of impact melts (Kieffer and Simonds, 1980; Osinski et al., 2008a), particularly for impacts into carbonates, it is also now clear that melting also occurs during impact into volatile-rich sedimentary target rocks (Osinski et al., 2008b). Impact is a process controlled by physics and hypervelocity impacts on other solid planetary bodies also result in impact melting (Melosh, 1989). On the Moon, with its relatively simple target geology and few complications, due to post-impact erosion and weathering, there is little debate that impact melt rocks are common within and around impact craters of all scales (see overview in Stöffler et al., 2006). The results are more complicated on other bodies with more diverse geologic processes and history, and where the contextual lines of evidence between the candidate impact rocks and their sources are blurred, such as on Mars (Tornabene et al., 2013).

This review provides a current synthesis of our knowledge and understanding of igneous rocks formed by hypervelocity impact. It describes the basics of the impact process and why melts are generated during impact events and how impact melting differs from endogenic igneous processes. It also characterizes the products of impact melting, where these rocks occur within and around impact craters, and their textural and chemical attributes. Occurrences of impact melt rocks on other Solar System bodies are discussed and the review concludes with some recommendations as to how to distinguish between igneous rocks produced by endogenic versus impact mechanisms.

## 2. Impact melting versus endogenic melting

There are significant differences in the initial conditions for the formation of igneous rocks by endogenic and impact processes (summarized in Table 1). As impact melting occurs at the surface and involves generally upper crustal target rocks, the closest analogies in terms of endogenic igneous rocks are volcanic rocks. Readers are familiar with the processes of endogenic melting to produce igneous rocks, particularly in the context of the terrestrial environment. In brief, the genesis and spatial occurrence of volcanic rocks on the terrestrial planets is a function of the geothermal and geodynamic environment of the planet, at any given time. The situation is the most complex on the Earth, with its multiplate environment and volcanism occurring in areas of thinned lithosphere (e.g., divergent plate margins, crustal rifts), subducting convergent plate margins and over mantle “hot spots”. It is generally the result of partial melting of mantle materials by “wt” or “flx” melting (convergent margins) or by decompression and adiabatic expansion (divergent margins, “hot spots”). These partial melts (most commonly, but not exclusively, of basaltic composition) are, by their nature, not superheated to a significant degree and generally free of clastic material. To our knowledge, the other terrestrial planets have only a single lithospheric plate and volcanism is related to only decompression and adiabatic expansion. Its occurrence, however, can still have a specific geospatial context, e.g., mare volcanism on the Moon in areas of thinned crust within major impact basins, or in the case of Mars, the most recent volcanism is confined to the northern hemisphere.

As readers may be unfamiliar with the nature of hypervelocity impact, what follows is a general overview of the basics of the impact process, particularly as it occurs on Earth. Those readers interested in more detail regarding the physics behind impact cratering, are referred to Melosh (1989) and to a recent synopsis of current knowledge on impact processes and products in Osinski and Pierazzo (2012). Extraterrestrial bodies ≥10^4 kg do not have their impact velocity reduced by atmospheric passage and impact the Earth with a velocity that is a combination of their cosmic velocity and the effect of Earth’s gravitational attraction. The minimum impact velocity of such objects is 11.2 km/s, which is the escape velocity of the Earth. Asteroidal bodies are the most common of such objects and impact with an average velocity of ~18 km/s. Less common are impacts by short-period comets, with an average impact velocity of ~30 km/s. Long-period comets are even less common but impact with a higher average velocity of ~50 km/s. On impact, these asteroidal and cometary bodies transfer their considerable kinetic energy to the target rocks. For example, a 1 km diameter, stony asteroid body impacting at 18 km/s contains ~2.5 × 10^{20} J of kinetic energy. The essentially instantaneous release of this amount of kinetic energy on impact is of the same order as the annual release of internal energy of the entire Earth. Impact events of this magnitude, however, occur on Earth on the million-year time-scale.

The impacting body transfers its kinetic energy to the target rocks via a shock wave, which propagates into the target rocks and back into the impacting body (Fig. 2). In the target rocks, the kinetic energy of the impacting body is partitioned into kinetic energy, which sets the target rocks in motion and leads to the formation of a crater form, and into internal energy, which leads to shock metamorphic effects. Since stress cannot be maintained at free surfaces (edges of the impacting body, the surface of the target rocks), rarefaction or release waves follow the propagating shock wave. The particle velocity vectors in the target rocks from rarefaction combine with the particle velocities induced by the passage of the shock wave to produce the so-called “cratering flow-field”, which results in the ejection of material from the upper and outer reaches of the target and downward displacement of material in the lower and central reaches of the target (Fig. 2). The maximum radial extent of ejected and displaced materials in the target by the cratering flow-field defines the so-called transient cavity in an impact event. The transient cavity is a conceptual construct and only

### Table 1

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<thead>
<tr>
<th>Property</th>
<th>Volcanics</th>
<th>Impact melt rocks</th>
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<tr>
<td>Melting mechanisms</td>
<td>Flux melting, adiabatic decompression from depth</td>
<td>Adiabatic decompression from shock compressed state</td>
</tr>
<tr>
<td>Nature of melting</td>
<td>Partial melting of protolith</td>
<td>Total melting of target lithologies</td>
</tr>
<tr>
<td>Initial temperature</td>
<td>Generally liquidus to sub-liquid</td>
<td>Superheated</td>
</tr>
<tr>
<td>Lithic, mineral clasts</td>
<td>Generally absent</td>
<td>Ubiquitous</td>
</tr>
<tr>
<td>Location</td>
<td>Specific, geodynamically controlled</td>
<td>Random</td>
</tr>
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exists as a physical entity in the smallest of impacts. It is generally taken to be approximately parabolic in cross-section but, as indicated by its name, it represents an unstable situation due to gravitational forces and collapses and is modified (Fig. 2), almost as it forms and grows, resulting in the final crater form.

The final crater form is primarily a function of the size of the impact event, planetary gravity and the dynamic strength of the target rocks; pre-existing structures and topography being other factors. As most terrestrial impact craters are no longer topographic depressions, due to their post-impact modification from the effects of erosion and tectonism, they no longer correspond to the strict definition of a crater. Thus, as impact involves the considerable displacement of the original target rocks, the more general and encompassing term impact “structures” is used here. Smaller impact craters are referred to as simple structures (Fig. 3A, C). When fresh, they are bowl-shaped in form, with an upraised and overturned rim, which is overlain by ejecta. They are partially filled, to approximately half the depth of the original transient cavity, with various impact lithologies termed impactites (see Section 3): breccias and impact melt rocks (Fig. 3A, C). (Note: the various settings of these igneous rocks formed by impact is discussed in Section 4.) Larger impact craters are called complex structures (Fig. 3B, D, E). Complex structures occur at an onset diameter that varies as a function of the specific planetary body (i.e., gravitational acceleration and target type), with diameters >2 to 4 km on Earth, depending on the

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**Fig. 2.** Schematic diagrams showing the formation of simple and complex impact craters. At the top is a theoretical cross section through the transient cavity showing the provenance of impact metamorphosed target lithologies. Below are a series of schematic cross sections depicting the 3 main stages of crater formation. The formation and transport of melt is highlighted in both simple (left panel) and complex craters (right panel). For the modification stage section, the arrows represent different time steps, labelled “a” to “c”. Initially, the gravitational collapse of crater walls and central uplift (a) results in generally inwards movement of material. Later, melt and clasts flow off the central uplift (b). Then, there is continued movement of melt and clasts outwards once crater wall collapse has largely ceased (c). Modified from Osinski et al. (2011) and Osinski and Pierazzo (2012).
nature of the target rocks (crystalline, sedimentary or both). They are characterized by a complex, faulted and collapsed rim area, a relatively, down-faulted flat floor and some form of uplifted structure in the centre (Fig. 3B). They represent a much more highly modified crater form, with respect to the transient cavity, than simple structures. The uplifted central structure consists of parautochthonous target rocks and has the form of an emergent topographic peak or peak ring (Fig. 3B, D, E), above the crater fill products lining the parautochthonous crater floor, depending on the size of the impact event. As with simple structures, the crater fill products at complex structures consist of various breccias and impact melt rocks.

Shock metamorphic effects are a direct result of the shock wave increasing the internal energy of the target rocks and are, thus, diagnostic of impact (French and Koeberl, 2010). They do not occur below shock pressures of several GPa and continue up to pressures of 100's of GPa (Fig. 4). They include the formation of: shatter cones (Fig. 4A), the only known megascopic shock effect; microscopic so-called planar deformation features (PDFs) (Fig. 4B), best known in quartz and feldspar; so-called diaplectic or thetomorphic solid-state glasses (Fig. 4C, D) in quartz and feldspar; impact melt rocks and glasses; and various high pressure polymorphs, such as coesite and stishovite from quartz and diamond from graphite. Shock metamorphic effects are not produced simply by the passage of the shock wave and compression of the target rocks but rather by shock compression combined with the effects following the passage of the rarefaction wave. With increasing pressure, the net effect of shock compression and pressure release is to increase the entropy and degree of disorder in the target rocks and their constituent minerals, such that still crystalline shocked rocks and minerals are less dense than their original state.

Impact melting, sometimes referred to somewhat erroneously as “shock melting”, occurs upon decompression from high shock pressures and temperatures. The extent of impact melting is a function of peak shock pressure and the compressibility of the target rock lithologies and their constituent minerals. It also occurs in the highly dynamic physical environment of impact crater formation, when target lithologies are in high speed and differential motion. Mass, momentum and energy are conserved across a shock wave and the state of material, as it is subject to shock compression, can be defined by Hugoniot equations, which describe the pressure in front and behind the shock wave, the particle velocity of material after the shock wave has passed and the specific internal energy of material in front and behind the shock wave (Melosh, 1989). The Hugoniot equation...
of state of geologic materials is a locus of a series of discrete shock states and is usually expressed in terms of specific volume and shock pressure. The imposition of a shock wave represents a discontinuous event and the target materials change suddenly across it. Unlike endogenic metamorphism and melting, there are no intermediate states and shock compression is a thermodynamically irreversible process.

Under shock compression, the internal energy of the target materials is increased and considerable pressure-volume work is done on the target materials. If the target materials are porous, the pores are closed by relatively low shock pressures. Decompression from the shocked state is by a rarefaction wave and decompression occurs via a release adiabat. Thus, not all the pressure-volume work, resulting from shock compression, is recovered and the unrecovered pressure-volume work that remains in the target materials is manifested as waste heat. It is this remaining waste heat on shock decompression that is responsible for impact melting and vapourization phenomena. The initial temperature of the impact melt is a function of the amount of waste heat remaining on decompression, which itself is a function of the peak shock pressure and the Hugoniot equation of state and release adiabat. As a result, the initial temperature of impact melts can be and are, as a rule, superheated with respect to endogenic melting temperatures of the target materials (Grieve et al., 1977). For example, peak shock pressures of ~60 GPa, with associated post-shock temperatures of ~1500 °C, are generally required to produce whole rock melts from crystalline quartzofeldspathic rocks. By peak shock pressures of ~100 GPa, however, the associated post-shock temperatures have risen to ~2500 °C (e.g., Stöffler and Grieve, 2007). The actual amount of super heating in impact melts remains poorly constrained; although the recent documentation of the transformation of zircon to cubic zirconia plus silica in impact melt rocks from the Mistastin Lake structure indicates superheating in excess of 2370 °C (Timms et al., 2017). The paths of compression and decompression due to a shock wave are illustrated schematically in Fig. 5.

More pressure-volume work is done on porous geological materials through closing the pore space, for a given shock pressure, compared to non-porous target materials (Kieffer et al., 1976; Kowitz et al., 2016; Osinski, 2007; Wünnemann et al., 2008). As this pore space, and the associated pressure-volume work, is not recovered on pressure release, these more compressible materials retain more post-shock waste heat, which results in impact melting at lower shock pressures. For example, hydrocode models indicate that the
peak pressure required to melt quartzite with no porosity is ~60 GPa, compared to only 30 GPa if there is 25% porosity (Wünnemann et al., 2008). This characteristic of impact phenomena accounts for the observation that samples of the lunar regolith are heavily charged with impact melted materials, compared to associated solid lunar rocks (McKay et al., 1991).

As rocks are made up of a variety of different individual minerals, which also vary in their compressibility, the more compressible...
minerals retain more waste heat and either thermally decompose or melt at lower shock pressures than their less compressible counterparts, with the first petrographic signs of impact melting being individual or, more often, mixed mineral melts (Lambert and Lange, 1984). The pressure range over which individual minerals melt is not large, a few 10s of GPa (Stöffler, 1972; Stöffler and Grieve, 2007), and whole rock melts are the norm, at least for impacts into crystalline targets (Osinski et al., 2008a).

As impact melted materials are being driven down into the expanding transient cavity, they have differential particle velocities, depending on the shock pressure they were subject to in the impact event. Thus, such high velocity turbulent flow results in impact melt bodies of a generally mixed composition, corresponding to the rocks within the volume of the target that was melted. Given the provenance of the target rocks, this mixed composition is generally crustal in nature. Indeed, it has been argued that the composition of terrestrial impact melt rocks from large impacts in Precambrian Shield areas is a good proxy for the average composition of the continental crust (e.g., Kring, 1997). This generally crustal composition for impact melts may have a possible exception in the largest impacts, such as those which formed the lunar multi-ring basins, which may involve a small component of mantle material (e.g., Johnson et al., 2016). The impact melt may also have a very small component from the impacting body. As the melt is driven down into the expanding transient cavity, and during the transient cavity’s subsequent modification to form the final crater, the melt incorporates a considerable amount of clastic material from the non-melted, but shocked and brecciated, portion of the target. This colder clastic material serves to lower the temperature of the superheated melt and subsequently undergoes digestion by the melt, through thermal decomposition, melting and reaction. How much clastic material remains and the nature of the final crystallized melt depends on its subsequent thermal history, which varies with the size of the impact event and the location of the melt in the final crater form. As a result, impact melts can be manifested in a variety of final lithological products.

### 3. Nomenclature and classification

#### 3.1. Background and overview

The transport and mixing of impact-metamorphosed materials during the formation of impact craters produces a wide variety of distinctive “impactites”. The IUGS Subcommission on the Systematics of Metamorphic Rocks defines impactites as “all rocks affected by one or more hypervelocity impact(s) resulting from collision(s) of planetary bodies” (Stöffler and Grieve, 2007). According to this definition, impactites from a single impact event can be classified into 3 major groups: shocked rocks, impact melt rocks, and impact breccias. It was proposed that impact melt rocks be subdivided into three subgroups (Stöffler and Grieve, 2007): clast-rich (Fig. 6A), clast-poor (Fig. 6B), or clast-free, and further sub-classified according to the degree of crystalinity as glassy, hypocrystalline, or holocrystalline. In addition to the terms dellenite and kärnäite mentioned previously, it is worth noting that the term tagamite was, and still is, used by some workers in the former Soviet Union to refer to impact melt rocks (e.g., Masaitis, 1999). It is recommended that all these names be avoided and instead the term impact melt rock be used. In addition, the term impact melt breccia should also be avoided, as it introduces unnecessary confusion as to whether a lithology is a “melt rock” or “breccia” and refers to a rock type that is actually a very clast-rich impact melt rock (Stöffler and Grieve, 2007).

Impact-generated melt is also a constituent of some impact breccias (e.g., Fig. 6C). Impact breccias that do not contain any melt component (i.e., lithic breccias, also referred to as fragmental or clastic breccias) will not be considered here. Whereas the nomenclature and classification of impact melt rocks is relatively intuitive, this is not the case for impact breccias. Historically, many workers have referred to impact melt-bearing breccias as “suevites” (Stöffler and Grieve, 2007); although at Canadian structures the term “mixed breccia” was generally used (e.g., Grieve, 1978). The term suevite is from the Latin name “Suevia” for the region around what is now known as the Ries impact structure, Germany (Sauer, 1920). Ironically, “suevite” was originally interpreted to be a volcanic rock (von Gümbel, 1870; Kranz, 1912).

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**Fig. 6.** Field images of impact melt rocks and impact melt-bearing breccias. (A) Clast-poor impact melt rock from the West Clearwater Lake impact structure, Canada. (B) Clast-rich impact melt rock from the West Clearwater Lake impact structure, Canada. (C) Impact melt-bearing breccia from the Mistastin Lake impact structure, Canada. (D) Clast-rich impact melt rock from the Haughton impact structure, Canada. 35 cm rock hammer for scale in A, B, and D. 8 cm lens can for scale in C.
until the discovery of coesite (Shoemaker and Chao, 1961) confirmed its impact origin and, by corollary, that of the Nördlinger Ries, now generally referred to as the Ries impact structure.

The definition of ‘suevite’ has evolved over time. Originally it was defined as an impact breccia with a clastic matrix and containing fragments of shock metamorphosed target rock and clasts of impact glass (Dressler and Reimold, 2001; Stöffler, 1977; Stöffler and Grieve, 1994). The recognition that some suevite samples from the Ries impact structure have impact melted material in the matrix (Osinski et al., 2004) led to the refining of the definition into an impact breccia with a particulate matrix (Stöffler and Grieve, 2007). This change exemplifies the inherent dynamic nature of the definition of impactites, as knowledge of their characteristics increases. Since the original discovery at the Ries structure, the term ‘suevite’ has been applied to various impactites at a large number of impact structures from around the world (e.g., Dressler and Reimold, 2001; Masaitis, 1999). Unfortunately, the use of the term ‘suevite’ has been inconsistent. Indeed, Grieve et al. (1977) noted that “the unqualified use of the term suevite at other impact structures (other than the Ries) has resulted in what are considered to be unwarranted arguments and conclusions regarding the physical conditions accompanying the formation and distribution of melt-bearing breccias within the cavity”. For example, impactites with melt contents ranging from ~1% (Koeberl and Reimold, 2005) to ~90% (Masaitis, 1999) have all been termed suevites. Furthermore, some workers also use the term “suevitic breccia”. This complicates and hampers the use of current classification schemes developed for impact melt-bearing impactites (Grieve and Thirria, 2012).

Further complications regarding nomenclature and classification of igneous rocks formed by hypervelocity impact have arisen because of complications regarding the identification of impact melt products derived from sedimentary target rocks. To a certain extent, this has been part of the reason for the debate surrounding the origin and classification of “suevites”, described above. This is particularly evident at the “type” locality at the Ries impact structure, which formed in a “mixed target”, with ~500–800 m of sedimentary rocks overlying crystalline basement (Schmidt-Kaler, 1978). Impact melt rocks sensu stricto are rare at the Ries structure (Osinski, 2004; Reimold et al., 2011), with “suevite” being the dominant impact melt-bearing impactite type (Stöffler et al., 1977). Estimates of the current volume of impact melt as low as 0.2 km³ have been proposed for the Ries (Pohl et al., 1977; Stöffler, 1977), which is much lower than the ~10–15 km³ predicted from the scaling relationship between melt volume and transient cavity diameter (Grieve and Cintala, 1992). It is, however, critical to note that the bulk of the groundmass of the Ries “suevite” is dominated by clay minerals, which most likely formed via post-impact hydrothermal alteration of impact glass (vong Engelhardt and Graup, 1984). If this is the case, then the amount of melt increases to at least ~5 km³ (vong Engelhardt and Graup, 1984) and possibly as high as 8 km³ (Stöffler et al., 2013), which is more in keeping with melt scaling relationships.

In impact structures formed in predominantly sedimentary targets, impact melt rocks were historically not generally recognized, with the resultant crater-fill deposits referred to as “clastic” or “fragmental” breccias. This is most clearly exemplified by the Haughton impact structure, Canada, which formed in a 1.9 km-thick sequence of predominantly carbonate rocks overlying crystalline basement. The crater-fill impactites at Haughton are well-preserved and exposed (Fig. 6D), with a present-day maximum thickness of ~125 m and volume of ~7 km³ and an original thickness estimated at ~200 m and a volume ~22.5 km³ (Osinski et al., 2005). These rocks were originally interpreted to be breccias, with some similarities to suevite in as much as they contain impact melt glasses derived from the crystalline basement (Redeker and Stöffler, 1988). Detailed electron microscopy studies, however, have demonstrated that the microscopic groundmass of these deposits consists of microcrystalline calcite and silicate impact melt glass, both of an impact melt origin from the sedimentary target rocks (Osinski et al., 2005; Osinski and Spray, 2001). These rocks are, therefore, more appropriately classified as clast-rich impact melt rocks. Subsequent work has shown that carbonate impact melts, akin to carbonatite igneous rocks, are present at a number of impact craters on Earth (see review by Osinski et al., 2008b).

Motivated by the lack of a systematic and easy-to-apply naming convention, Osinski et al. (2008a) proposed an alternative more descriptive classification scheme for impact melt-bearing impactites. Unlike the earlier classification scheme of Stöffler and Grieve (2007), the proposed scheme reflects the fact that there is a complete continuum from clast-free impact melt rocks to melt-free lithic impact breccias. This scheme utilizes well-accepted textural terms from igneous petrology and does not involve any genetic implications in terms of emplacement processes. We will return to this discussion and provide examples of these different textural types of impact melt rock in Section 5.

### 3.2. Impact glass

Impact glass is a common constituent of many impact breccias (e.g., Fig. 6C) and, more rarely, in impact melt rocks. Impact glass may also be found as mm- to cm-size individual bodies not contained in impact breccias. So-called “melt beads” have been documented in the proximal ejecta zone of terrestrial impact craters and more distally, where they are known collectively as tektites (Fig. 7A). Millimetre-sized spherules of impact melt can also occur in distal ejecta deposits (Fig. 7B) and where deposited in continuous beds (Fig. 7C) are known as “airfall beds” or “impactoclastic” deposits (Stöffler and Grieve, 2007). The terms microtektites (if they consist entirely of glass) or microxenites (if they contain primary microlites) are sometimes used for these spherules.

There are three fundamentally different formation mechanisms – vapourization, melting, solid-state – by which glasses may be generated during an impact event, resulting in 5 main types of different types of “impact glass” (Table 2). Melting is the most common mechanism to form a glass, but there are at least 3 different products (Table 2). Typically, when most workers refer to impact glass, they are referring to a whole rock impact glass produced from melting a specific volume of rock that would typically comprise several mineral or rock types. Impact glass clasts in breccias (Figs. 6C, 7D, E) noted above, as well as tektites (Fig. 7A), all fall into this category and, therefore, are, by far, the most common type of impact glass. Diaplectic glasses, also termed “theromorphic glasses”, form via solid-state transformation of tecto-silicates and retain the identical composition and morphologic form of precursor mineral (Table 2; Fig. 4C, D). They are, therefore, a sub-solidus shock metamorphic feature (French and Koeberl, 2010) and are not considered further here.

In a recent important contribution, Johnson and Melosh (2012) produced the first detailed numerical model of spherule formation. They show that glassy spherules present in distal ejecta can form in two fundamentally different ways: via “conventional” impact melting or from condensation from a vapour phase. They suggested the terms “melt droplets” and “vapour condensate spherules”, as the two respective products (Table 2).

In addition to impact glasses found within and around impact structures, and tektites and spherules found in distal ejecta, there are a number of enigmatic occurrences of glasses either confirmed, or suspected, as being of impact origin but for which no source crater has been recognized. Some of these glass occurrences are well known and widely accepted as being of impact origin. Examples include the Libyan Desert Glass (Weeks et al., 1984), Darwin Glass (Meisel et al., 1990), South Ural Glass or Urengoites (Deutsch et al., 1997), and Dakhleh Glass (Fig. 7F) (Osinski et al., 2007, 2008c). Importantly, numerical modeling studies have shown that substantial amounts of glass can be formed by radiative/convective heating of the target surface during larger, >100 Mt low-altitude airbursts (Boslough and Crawford, 2008), supporting earlier suggestions that the Libyan Desert Glass and the Muong–Nong
Tektites of southeast Asia may have formed from such events (Wasson, 2003). This is the preferred origin of the Dakhleh Glass (Osinski et al., 2007, 2008c). We do not discuss such glasses further in this contribution.

3.3. Nomenclature and classification of “sorted” impactites

There is a growing list of examples in the impact cratering literature of impact melt-bearing impactites that are described as being “sorted”.

<table>
<thead>
<tr>
<th>Suggested nomenclature</th>
<th>Stöffler (1984)</th>
<th>Formation mechanism</th>
<th>Product Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>Diaplectic glass</td>
<td>Mineral glass</td>
<td>Solid state</td>
<td>Glass with same composition as host mineral; retains original grain shape</td>
</tr>
<tr>
<td></td>
<td></td>
<td>transformation</td>
<td>Glass with same composition as host mineral; can contain vesicles and flow features</td>
</tr>
<tr>
<td>Diaplectic plagioclase glass</td>
<td>Rock glass</td>
<td>Melting</td>
<td>Also referred to as “thetomorphic glass” by some workers. Diaplectic plagioclase glass is also known as maskelynite. Form via the selective melting of individual minerals. Lechatelierite is the name applied to a mineral glass derived exclusively from quartz.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Melting</td>
<td>Localized melting at grain boundaries of minerals and pores (also termed “melt pockets” in meteorites).</td>
</tr>
<tr>
<td>Whole rock impact glass (&quot;impact glass&quot; for short)</td>
<td>Rock glass</td>
<td>Melting</td>
<td>Whole rock melting. Tektites and melt droplets (as defined by Johnson and Melosh, 2012) would fall into this category and could be considered sub-types.</td>
</tr>
<tr>
<td>Vapour condensate spherules</td>
<td>N/A</td>
<td>Condensation from a vapour phase</td>
<td>Glassy spherules with the composition of a whole rock(s).</td>
</tr>
</tbody>
</table>

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Fig. 7. Impact glasses. (A) Muong-Nong tektites. 1 cent US coin for scale. (B) Plane-polarized light photomicrograph of spherules from Bee Gorge, Australia. The host is the Wittenoom Formation, Hannansley Group, Western Australia. Image from B. Simonson. (C) The Monteville spherule layer within the Ghaap Group, Griqualand West basin, South Africa. Image from B. Simonson. (D) Plane-polarized light photomicrograph of fresh and unaltered impact glass clast within breccia. Ries impact structure, Germany. (E) Plane-polarized light photomicrograph of flow-textured glass. Glass clast within breccia from the Ries impact structure, Germany. (F) Field photograph of a large specimen of Dakhleh Glass, Egypt. 6 cm diameter camera lens cap for scale.

Tektites of southeast Asia may have formed from such events (Wasson, 2003). This is the preferred origin of the Dakhleh Glass (Osinski et al., 2007, 2008c). We do not discuss such glasses further in this contribution.
“bedded”, “graded”, or with more genetic connotations, “reworked”. There is, unfortunately, no consistency or existing nomenclature in place to systematically and consistently describe such impactites. We suggest that they can be organized into three main groupings, only two of which conform to the definition of an impactite. In some instances, impact melt-bearing impactites that display normal grading and with gradual contacts with underlying crater-fill deposits (see Section 4.1 below) occur in the crater interior (e.g., Koeberl et al., 2007a; Pernicka et al., 1987). The properties of these deposits are consistent with having been deposited immediately after the impact from the ejecta plume, in which case they can be considered impactites and the term “fallback” may be applied. The second group is the previously referred to “impactoclastic” deposits, which comprise primary impactite deposits comprised of millimetre-sized spherules of impact melt in distal ejecta deposits (Stöffler and Grieve, 2007).

There is a third category of deposits that, while hard to judge from the literature, appear to have been eroded and transported over some distance within or around an impact crater. Whether they were deposited within hours or years of the impact event, these deposits do not conform to the definition of an impactite (Stöffler and Grieve, 2007). They are, however, akin to volcaniclastic igneous rocks. In other words, they are rocks composed of impact-generated material that, like volcaniclastic rocks, are “transported and reworked through mechanical action, such as by wind or water” (Fisher and Schmincke, 1984). As such, we suggest that the term impactoclastic be used to describe such reworked impactites.

4. Setting

Field studies on Earth, combined with satellite observations of fresh craters on other planetary bodies, indicate that impact melt products can be found in 4 main geological settings within and around hypervelocity impact craters (Fig. 8): 1) in the crater interior (i.e., as crater-fill deposits); 2) in injection dykes in the crater floor; 3) in proximal ejecta deposits; and 4) in distal ejecta deposits. These settings are described in detail below.

4.1. Crater-fill deposits

The stratigraphic setting where the largest volume of impact melt is deposited is the interior of both simple and complex craters (Figs. 8A, 9A, B, 10). Known collectively as crater-fill impactites, to distinguish them from post- and non-impact crater-fill deposits, they comprise a range of different impact melt-bearing lithologies, from clast-free crystalline impact melt rocks composed of almost entirely impact melt products to melt-poor impact-bearing breccias with ~1% scattered impact glass clasts. It is important to note that even within individual craters, crater-fill deposits can comprise a variety of intercalated impactite types with lenses of one type enclosed in another. They also need not cover the entire crater floor, which is particularly common in small craters (e.g., Fig. 9B). When crater-fill deposits comprise entirely large bodies of coherent crystalline melt rocks, the term impact melt sheet is often applied, an excellent example of which is provided by the West Clearwater Lake impact structure, Quebec (Fig. 8A). Examples of unmodified crater-fill impact melt sheets are also commonplace on the Moon (Fig. 9A, B) and Mercury and serve as a guide for understanding the original morphology of impact melt rocks in craters on Earth prior to erosion.

The thickness of crater-fill deposits increases with the size of the crater, ranging from 10s m in small terrestrial craters up to ~3 km for the ~200 km diameter Sudbury impact structure (Therriault et al., 2002); with even greater thicknesses likely present in larger impact basins throughout the Solar System (e.g., Vaughan et al., 2013). As noted in the previous section, not all terrestrial impact craters possess a “sheet” of crystalline impact melt rock, largely due to the difficulty in identifying impact melt products derived from porous, volatile-rich sedimentary rocks (Osinski et al., 2008b) and not because of the lack of impact melt. Large bodies of impact melt-bearing crater-fill impactites that are stratigraphically equivalent to crystalline impact melt sheets are present in craters formed in sedimentary target rocks on Earth (e.g., the crater-fill impact melt rocks at the Haughton impact structure (Osinski et al., 2005; Osinski and Spray, 2001) shown in Fig. 6D). This is in keeping with models that predict as much, or even more, impact melt...
melt should be derived from impacts into sedimentary target rocks (Kieffer and Simonds, 1980; Wünnemann et al., 2008).

This has important implications for Mars, which like Earth, also possesses a volatile-rich crust. Until recently, it was assumed that large volumes of impact melt would not form on Mars (e.g., Schultz and Mustard, 2004). This hypothesis was largely based on the assumption, now shown to be incorrect, that impact melt is rare or lacking in impact craters on Earth formed into volatile-rich sedimentary targets. With the aid of high-resolution imaging it has now been shown that crater-fill impactites interpreted to be melt-rich are observable in fresh and well-exposed Martian impact craters (Tornabene et al., 2012). The characteristic “pits” in these Martian crater-fill impactites (Fig. 10A, B) are proposed to have formed via release of volatiles, as these deposits cooled (Boyce et al., 2012), and considered as a natural outcome of impacts into volatile-rich target rocks. The documentation of so-called “degassing pipes” in the impact melt-bearing breccias of the Ries impact structure (Newsom et al., 1986) support this concept of volatile release, following deposition of crater-fill impactites, in craters formed in volatile-rich targets on Earth and Mars. The presence of similar pitted deposits associated with the best-preserved craters on the asteroid Vesta (Fig. 10C, D) (Denevi et al., 2012) and Ceres (Fig. 10E, F) (Sizemore et al., 2017) suggests that pitted impactite deposits are more common than their lunar-like counterparts for volatile-rich rocky bodies in the Solar System. A caveat to this for objects such as Vesta and Ceres are their generally small size and lower impact velocities, which will reduce the overall amount of impact melt produced (see Section 2).
Fig. 10. Examples from Mars (A, B), Vesta (C, D) and Ceres (E, F) of the generally smooth and pitted ("pm") crater-fill deposits that are consistent with impact melt-bearing deposits that are either volatile-rich and/or have interacted with volatile-rich materials. All three examples also possess occurrences of pitted materials ("pm") on uplifts, wall slumps and terraces and externally in low-lying topographic areas surrounding the craters that flowed into pre-existing craters and are consistent with smooth and hummocky impact melt-bodies observed on the Moon and Mercury. A, B) The Martian example is of the ~10 km Tooting Crater (images: A: P02_001764_1877_XL_07N193W and B: PSP_001764_1880; credits: NASA/JPL-Caltech/MSSS and NASA/JPL-Caltech/University of Arizona). C, D) The ~67.5 km Marcia Crater on Vesta (images: C, D: Dawn Framing Camera LMAO image mosaic; credits: NASA/JPL-Caltech/UCLA/MPS/DLR/IDA). E, F) Example from Ceres is of the ~50 km Ikapati Crater (images: E: XMO2 Image 29 and F: Dawn LAMO Image 204; credits: NASA/JPL-Caltech/UCLA/MPS/DLR/IDA). Note the differences illumination between the sets of images from Mars, Vesta and Ceres.
4.2. Injection dykes

Impactite dykes are a common occurrence in the crater floor and central uplift of many impact craters on Earth. The most common types of impactite dykes are lithic breccias and pseudotachylite (Dressler and Reimold, 2004), which are outside the scope of this contribution. (Pseudotachylite, while considered to have an igneous groundmass is considered a product of impact tectonics rather than impact melting.) In their review of breccias in floors of large impact structures, Dressler and Reimold (2004) discuss on the occurrence of dykes of impact melt rock and impact melt-bearing breccia at 4 large structures: Manicouagan, Morokweng, Sudbury and Vredefort. For the purposes of this review, we subdivide impact melt-bearing dykes into two categories; thin, minor dykes, and more substantive and thicker dykes that most likely form sometime after initial crater formation.

The simplest explanation for the formation of impact melt-bearing dykes is the downward and/or outwards injection of the original melt lining of the crater walls and floor into fractures. In this regard, they would represent extensions of, and be genetically related to, the crater-fill melt-bearing impactites described in the previous section. Based on the literature, this seems to be the case at the Manicouagan and Morokweng structures as discussed by Dressler and Reimold (2004). Dykes of impact melt rock and impact melt-bearing breccia are also present at the Mistastin Lake (Currie, 1971; Grieve, 1975) (Fig. 8B) and West Clearwater Lake (Bostock, 1969) impact structures in Canada. At the West Clearwater Lake impact structure, dykes were observed in the field to connect with the nearby crater-fill impact melt sheet (Wilks and Osinski, 2016), lending strong support that a portion of the melt lining is injected into fractures and/or faults during crater formation.

On a different scale altogether are impact melt-bearing dykes at the Sudbury impact structure. They are referred to as Offset Dyke (Grant and Bite, 1984), so called for the way they apparently terminate along strike, but reappear displaced by as much as a few km parallel to strike. The Offset Dykes are found extending radially from – and concentrically around – the Sudbury Igneous Complex, which is a differentiated impact melt sheet that is now elliptical in outcrop shape, due to post-impact deformation (see Section 7.2). The dykes are impressive in size, ranging from ~10 to 80 m wide and up to 50 km long. Although they are typically granodioritic in composition (Lightfoot et al., 1997), they are historically referred to as quartz diorite, and consist of a central so-called inclusion-rich quartz diorite phase and a marginal inclusion-poor quartz diorite phase. Despite over 100 years of exploration the exact timing of formation of the Offset Dykes is unclear. Some authors have suggested that the dykes formed from a forceful outwards and downwards injection event during the excavation stage of crater formation, prior to any differentiation of the Sudbury Igneous Complex (e.g., Tuchscherer and Spray, 2002); whereas others have proposed the dykes formed during the modification stage due to the collapse of the transient cavity (e.g., Wood and Spray, 1998). Others have suggested multiple injection events, beginning during the early excavation stage and continuing into the modification stage (e.g., Murphy and Spray, 2002); while still others have suggested that emplacement occurs thousands of years following the impact event (Hecht et al., 2008).

The granophyre dykes at the Vredefort impact structure have quite a similar morphology and composition to the Offset Dykes, but occur exclusively in the central uplift of the impact structure. They extend radially outwards from and concentrically around the apparent centre of the impact structure (Reimold and Gibson, 2006). They are similar in size to the Sudbury Offset Dykes, ranging from ~10 to 50 m wide and up to ~10 km long. The dykes consist of a felsic fragment-rich granophyre in the centre of the dyke and a mafic fragment-poor granophyre along the margins of the dyke. Liefger and Riller (2012) have proposed that the dykes formed during two stages of injection events from an overlying impact melt sheet. First, the marginal phase was emplaced, and was hot enough to assimilate most fragments. This was followed by a second injection of more felsic impact melt in the centre of the dyke, which was no longer hot enough to assimilate all the fragments, resulting in a more felsic and fragment-rich core.

4.3. Proximal ejecta deposits

Proximal impact ejecta deposits are, by definition, found within 5 crater radii of the source crater (Stöffler and Grieve, 2007). Beyond this limit, impactites are known as distal ejecta. An important clarification to this definition is that proximal impact ejecta deposits are defined as having been transported (i.e., not necessarily “ejected” through the atmosphere) beyond the rim of the transient cavity. The result is that in complex craters, ejecta deposits occur interior to the topographic crater rim in the zone of collapse faults or terraces (Fig. 8).

Our understanding of impact ejecta is hampered by the fact that these deposits, both proximal and distal, are the first to be eroded such that there are very few examples of proximal ejecta deposits around craters on Earth. The few examples that are present and that have been studied, however, offer critical insights into the properties of impact ejecta. Studies of continuous ejecta blankets around terrestrial craters reveal that an impact melt component is either lacking (e.g., Hörz, 1982) or relatively rare (e.g., Osinski et al., 2005, 2015). Since their first discovery on the Moon (Hawke and Head, 1977), it has become clear that “ponds” and “flows” of what are generally interpreted as impact melt-bearing impactites form a discontinuous second layer of ejecta overlying the blocky, melt-poor continuous ejecta blanket (Fig. 8) around most craters on all the terrestrial planets and the Earth’s Moon (Fig. 9A-C). What are interpreted to be both impact melt rocks and impact melt-bearing breccias have been identified in remote sensed data. Around the ~960 km diameter Orientale multi-ring basin on the Moon (Fig. 9E), some flows exceed 150 km in length and extend well beyond the crater rim (Fig. 9F) (Morse et al., 2018). On geologically active planets such as Mars, this has important implications for the interpretation of seemingly isolated “sheets” of igneous rocks in the early crust (e.g., Mustard et al., 2009) as well as float samples encountered by the various landers and rovers. In these cases, an impact melt origin should also be considered, in addition to an interpretation involving endogenic melting processes.

In addition to impact melt rocks and melt-bearing breccias, impact melt has also been documented in the form of isolated “melt beads” (i.e., not contained in a breccia) overlying and potentially in the proximal ejecta deposits of many “small” craters, e.g., ~45 m diameter Kamil Crater (Folco et al., 2011), ~90 m diameter Wabar Crater (Hörz et al., 1989), and the Henbury crater field (Taylor, 1967), which comprises 14 craters ranging from 7 to 180 m in diameter, and the 1.6 km diameter Barringer (or Meteor) Crater (Hörz et al., 2002).

4.4. Distal ejecta deposits

One of the most well-known products of hypervelocity impact events is tektites (Figs. 7A, 8D), which fall into this final class of setting of igneous rocks formed by hypervelocity impact. Originally thought to be of lunar origin (Gilvarry, 1965), it is now widely accepted that tektites are impact glass formed at terrestrial impact craters from near-surface melt, which was ejected ballistically and deposited, sometimes as aerodynamically-shaped bodies, in a strewn field beyond the continuous ejecta blanket (Stöffler and Grieve, 2007). Tektites occur in various shapes, sizes and colours; when microscopic they are typically known as micrometeorites or microkrystites or simply spherules. Three of the four large localized strewn fields: the Ivory Coast, Central European (Czechoslovakian/Moldavian) and North American, have all been traced to a single source craters (Glass, 1990); only the source of the Australasian strewn field remains unknown. Tektites are probably one of the most well-studied types of impact glass and the interested reader is referred to the several articles that review their properties and origin (Glass, 1990; Koeberl, 1986, 1994).
When deposited in coherent layers, small mm-size glassy to crystalline spherules form impactoclastic airfall beds (Fig. 7C) (Simonson and Glass, 2004) and display sedimentary structures, such as normal grading and cross stratification reflecting a complex depositional history. At least 18 Precambrian-aged, spherule-bearing impactoclastic beds have been documented from South Africa and Australia and range in age from ~1 Ma to ~3.5 Ga (Glass and Simonson, 2012). These spherules beds provide a critical record of ancient impacts on Earth. With no known impact craters older than ~2.5 Ga in age, they represent the only record of hyperviolet impacts in the Archean. As noted above, Johnson and Melosh (2012) have proposed that spherules present in distal ejecta can form in two different ways: via “conventional” impact melting or from condensation from a vapour phase.

5. Textures

The textures of endogenic igneous rocks have been described in detail and studied for well over a century (e.g., Cross et al., 1906). While not as well studied or generally acknowledged, igneous rocks formed by hypervelocity impact share many textural similarities with endogenic igneous rocks. In general, the texture of igneous rocks in general refers to variations in the size, shape and arrangement of mineral grains and their relationship to one another. It is convenient to group igneous textures into two main groups – those that reflect the size of the minerals (phaneritic, aphanitic, porphyritic, pegmatitic, and glassy), and those that reflect other properties (vesicular, pyroclastic). The common textures of endogenic igneous rocks are applicable and appropriate for describing igneous rocks formed by hypervelocity impact. The terms holocrystalline (completely crystalline), hypohyaline (partly glassy), and holohyaline (completely glassy) are also appropriate descriptors of impact melt rocks and breccias.

5.1. Factors controlling textures

The textures of endogenic igneous rocks are controlled by a variety of factors, including cooling rate, degree of undercooling, composition, temperature, aqueous fluid and gas contents and the availability of nucleation sites (e.g., McBirney, 2006). The same properties likely control the texture of impact-generated melts (Osinski et al., 2008a). An additional factor for planetary studies is the presence or absence of an atmosphere: melts will cool more quickly under a convective atmosphere than through radiation alone.

In addition to the potential for superheating, a major difference between impact-generated and endogenic melts is the complicating factor of lithic and mineral clast content and their variable states of recorded shock metamorphism. As discussed, immediately after impact melting occurs, the melt is charged with mineral and lithic fragments as it is driven in turbulent flow into the expanding transient cavity and overtake slower moving still solid (but shocked and fragmented) target materials (Fig. 2). The admixture of colder clasts initially acts to rapidly cool the super-heated impact melt (Onorato et al., 1978). As with endogenic magmas (McBirney, 1979), the assimilation of some or all of these clasts can subsequently change the final composition of the solidified end product. While not well constrained, given initial clast contents up to ~40–50 vol% (Grieve, 1978), the effect of assimilation for impact melt rocks likely plays an even more important role than in endogenic igneous rocks (see discussion in Section 6.4). It should be noted that the relative abundance and composition of the clasts do not necessarily reflect the nature of the target rocks that were melted, as the clasts come from a different volume of the target from that which was melted and can vary, depending on the relationship between the magnitude/scale of the impact event and the scale of the lithological heterogeneity of the target rocks (e.g., McCormick et al., 1989). The high clast content, and their variable shock states, in impact melts will also affect the availability and nature of nucleation sites in impact melts. Based on experiments on basaltic melts, varying the kind and density of nuclei can produce textural differences comparable to those produced by varying cooling rate (Lofgren, 1983). Thus, it is reasonable to suppose that the distribution, density, and kind of clasts in impact melts will exert an important role on textures. Based on studies at Manicouagan and at other terrestrial impact structures, it has been suggested that one of the characteristics of coherent impact melt sheets (that were not thick enough to subsequently differentiate) is that they are initially remarkably chemically homogeneous, due to turbulent mixing of various melted lithologies in the impact, but texturally heterogeneous, due to their acquired, but variable, lithic and mineral clast content (e.g., Floran et al., 1978).

5.1.1. Size of crater

As discussed in Section 2, the magnitude of the impact governs how much melt is produced. In general, as the magnitude of the impact increases, the greater the volume of melt relative to the size of the transient cavity and the final crater form, due to “differential scaling” (Grieve and Cintala, 1992). The more melt produced, the thicker the subsequent impact melt deposits and the slower their cooling rates. Differential scaling refers to the fact that planetary gravity is a primary variable in determining the efficiency of a given impact to form a crater of a given size (Grieve and Cintala, 1997). As it is a force that inhibits crater growth and includes a time term, cratering efficiency is reduced on higher gravity planetary bodies and in larger, compared to smaller, impacts. Thus, the effects of gravity are most pronounced in comparing large impact structures between bodies such as the Earth and the Moon. Gravity, however, is not a primary variable in comparing the magnitude and geometry of the shock wave in the target and, thus, the volume of impact melt produced in a given impact; although planetary gravity affects impact velocity, with higher velocities resulting in more melt (Cintala and Grieve, 1994; Grieve and Cintala, 1992). The larger the volume of melt, relative to the transient cavity and final crater form, the fewer lithic and mineral clasts that are acquired during turbulent flow. This results in a higher the temperature for the melt, following thermal equilibration with the incorporated clast population, and the greater the thermal energy available to assimilate that clast population and crystallize clast-poor impact melt rocks.

5.1.2. Target lithology

A major advancement during the late 1990s and early 2000s was the recognition of evidence that sedimentary rocks melt during hypervelocity impact (Graup, 1999; Jones et al., 2000; Osinski and Spray, 2001). As noted earlier, the major reason that it took so long for this evidence to be forthcoming is the difficulty in identifying impact melt products derived from sedimentary target rocks. At the outcrop and hand specimen scale, the products of impact in to sedimentary targets (e.g., Fig. 6D) are unlike those produced from impacts into dense, nonporous crystalline rocks (e.g., Fig. 6A, B). With rare exceptions, it requires the application of electron microscopy techniques to identify impact melts derived from sedimentary rocks. While, it is, thus, clear that the nature of the target lithology can greatly affect the overall texture of impact melt lithologies, the reasons for these differences are a topic of active research.

Osinski et al. (2008a) investigated the effects of target lithology on the products of impact melting and concluded that impact melt generated from impacts into sedimentary, or mixed sedimentary–crystalline targets, will typically cool more rapidly and will, thus, assimilate fewer clasts than melt from crystalline targets. This is due to the fact that the enthalpies of H2O–carbonate-bearing systems are such that a much smaller amount of entrained sedimentary clasts compared to relatively anhydrous crystalline rock is required to quench a melt to subsolidus temperatures (Kieffer and Simonds, 1980). Another factor is the question of superheating, which also affects the ability of a melt to assimilate clasts. It is well established that impact melts formed from impact into crystalline rocks are superheated to initial temperatures of at least 1500–2500 °C (Grieve et al., 1977). The presence of pure SiO2 glass or lechatelierite, which requires temperatures in excess of
–1650–1750 °C, derived from melting sandstones at a number of terrestrial impact structures (Milton et al., 1996; Osinski, 2003; Redeker and Stöffler, 1988) indicates that similar initial temperatures can also be reached during impacts into sedimentary rocks. A final factor to be considered is the gas content. Impact melts derived from sedimentary rocks will be more volatile rich than those derived from crystalline rocks. The effect(s) of gas content on the properties of impact melts is essentially unknown, at present. Overall, the result appears to be that the melt products of impact into sedimentary target rocks are typically more fine-grained or glassy and clast-rich compared to those from similar-sized craters into crystalline targets (Kieffer and Simonds, 1980; Osinski et al., 2008a).

5.1.3. Setting

The effect of the geologic setting in the final crater form on the textures of impact melt lithologies is more straightforward than the effects of target lithology and crater size. The setting will play a major role in controlling the cooling rate of impact melts once they are emplaced. Of the four settings discussed in Section 4, crater-fill impactites will cool the slowest, as this is where the greatest volume of impact melt is deposited. Most melt injection dykes are relatively thin, being a few cm to m in width, and cooling through conduction to the wall rocks will typically result in glassy or fine-grained rocks. There are exceptions, such as some of the large so-called Offset Dykes at Sudbury, where relatively coarse-grained igneous-textured rocks occur (Lightfoot et al., 1997).

In terms of proximal ejecta, emplacement can be more complicated, as there is a large variation in the volume of melt that is emplaced at any particular location. The main reason for this is that topography sloping in that direction, which created a lower rim, and possibly the breach of King Crater’s rim due to it being, in part, superposed on the older pre-existing crater. The general consensus is that these late-stage melt-rich ejecta deposits are emplaced as flows following the ballistic emplacement of the (melt-poor) continuous ejecta blanket (Hawke and Head, 1977; Osinski et al., 2011). As a result, less cooling occurs during transport than if the melt was transported through the “air” and these deposits are generally emplaced above the liquidus.

5.2. Textures of impact melt rocks and melt-bearing breccias

5.2.1. Phaneritic

The term phaneritic, derived from the Greek word phaneros and meaning visible or evident, is applied to igneous rocks in which the crystals are clearly visible to the unaided eye. The interpretation is that these rocks cooled relatively slowly. As discussed in Section 5.1, the dominant controlling factor on the grain size of impact melt rocks is the initial thickness of the deposit, which is in turn controlled predominantly by the size of the impact. Excellent examples of holocrystalline phaneritic impact melt rocks are present at the ~200 km diameter Sudbury (Therriault et al., 2002), ~100 km diameter Manicouagan (Floran et al., 1978), and ~50 km diameter West Clearwater Lake (Simonds et al., 1978) impact structures in Canada. At Sudbury, such rocks are known from the crater-fill impact melt sheet (Fig. 11A) and injected dykes in the crater floor; whereas at West Clearwater Lake, phaneritic impact melt rocks are only present in small interval (several 10s of metres) in the middle of the crater-fill impact melt sheet (Fig. 11C). Given the preponderance of phaneritic impact melt rocks in craters larger than ~50 km on Earth, impact melt rocks at large craters on other planetary bodies will also form such textures. Indeed, some impact melt rock samples in the Apollo sample collection are phaneritic (Fig. 11D).

Fig. 11. Phaneritic impact melt rocks. (A) Field image of the granophyre unit of the ~3 km-thick Sudbury Igneous Complex. Note that the granophyre is predominantly clast-free. 40 cm long rock hammer for scale. (B) Cross-polarized light photomicrograph of the granophyre unit shown in (A). (C) Phaneritic clast-poor impact melt rock from the West Clearwater Lake impact structure. Part of 14 cm long pencil for scale. (D) Apollo 14 sample 14073. Originally interpreted as a basalt, this sample is now classified as an impact melt rock (Neal et al., 2015). Sample is 3 cm across.
5.2.2. Aphanitic

Most known occurrences of impact melt rocks in craters on Earth are aphanitic (Figs. 1B–D; 6A, B; 12A). They are known from crater-fill deposits (e.g., Figs. 6A, B; 8A), proximal ejecta (Figs. 1A, B, 12A) and injection dykes. Most lunar impact melt rocks in the Apollo sample collection are also aphanitic (Fig. 12B). In craters such as Sudbury and West Clearwater Lake, where phaneritic impact melt rocks also occur, aphanitic rocks occur at the base and top of the melt sheet and are more clast-rich (Fig. 6B) (Anders et al., 2015; Simonds et al., 1978). This is a general observation that is common to all craters, where a thick enough succession of rocks remains to make such observations possible (Grieve et al., 1977). In addition to forming continuous sheets, such as at West Clearwater Lake, aphanitic impact melt rocks also occur as discontinuous lenses intercalated with impact melt-bearing breccias, in craters ranging in size from a few km (e.g., ~4 km diameter Brent Crater, Canada (Grieve, 1978)) to the ~100 km diameter Popigai impact structure, Russia (Masaitis, 1994). Aphanitic impact melt rocks can display columnar jointing (Figs. 1A, B, 12A).

Aphanitic impact melt rocks typically contain lithic and mineral clasts (e.g., Fig. 6B), which are shocked to varying degrees. As these clasts are in chemical and physical disequilibrium with the surrounding melt, reaction rims are a common occurrence, with pyroxene coronas around quartz clasts being most frequent (Grieve, 1975; Grieve et al., 1987). Aphanitic impact melt rocks are crystalline to varying degrees, with the grain size ranging from ~1 to 2 mm phenocrysts to μm-size microlites (e.g., Fig. 12D, E). The presence of phenocrysts gives some impact melt rocks a porphyritic texture (e.g., at Boltysch, Ukraine; (Grieve et al., 1987)). Larger phenocrysts can display normal zoning. Smaller crystals commonly display skeletal, hollow, and dendritic forms (Fig. 12D, E), which are well-understood quench crystal morphologies common in fine-grained volcanic rocks and produced in experiments. These crystal morphologies indicate rapid crystallization from a melt in response to high degrees of undercooling and supersaturation, and low nucleation densities (Bryan, 1972; Donaldson and Dawson, 1978; Lofgren, 1974). Several studies have reported unusual chemistry of microlites in impact melt rocks (e.g., Al-rich clinopyroxene and Fe–Mg-rich plagioclase) (e.g., Grieve and Ber, 1994; Osinski, 2003). This is consistent with quenching and rapid growth that inhibits the expulsion of exotic cations from the crystal structure.

5.2.3. Glassy

As discussed in Section 3.1, impact glass is a common product of impact craters on Earth. In fact, it appears to be ubiquitous having been documented in craters of all sizes and from all target rocks. Occurrences of glassy impact melt rocks are known but are minor in terms of volume compared to aphanitic and phaneritic impact melt rocks (Fig. 13). In some instances, glass forms the mesostasis (i.e., the last-formed interstitial material) comprising only a few vol% of the rock in between primary crystals (phenocrysts) and clasts (e.g., Ries; Osinski, 2004) (Figs. 12D, 13C, D); whereas in others cases the glassy groundmass forms up to ~75 vol% of the melt rock (e.g., Boltysch; Grieve et al., 1987). In other instances, glass is found at the contact of injection dykes of melt rock, representing chilled margins (Fig. 13A). Some rare examples of glassy impact melt rocks are present in the Apollo sample collection (Fig. 13B).

Far more common is the presence of glass as “clasts” in breccias (e.g., Fig. 6C). Such impact melt-bearing breccias are the most common type of impact melt-bearing lithology in craters on Earth. Impact melt-bearing breccias can occur in association with impact melt rocks in craters developed in crystalline (e.g., Mistastin Lake, West Clearwater Lake, Manicouagan impact structures, Canada) and mixed sedimentary–crystalline (e.g., Popigai) target rocks. In some craters, impact melt rocks are apparently lacking and impact melt-bearing breccias are the only melt-bearing impactite type (e.g., at Barringer, U.S.A; Osinski et al., 2015; at Bosumtwi, Ghana; Koebel et al., 2007a,b). Impact glass-bearing breccias are also known from the Apollo sample collection and among the

Fig. 12. Aphanitic impact melt rocks. (A) Field image of aphanitic impact melt rock from the Mistastin Lake impact structure. A 35 cm rock hammer is for scale in the centre of the image, lying on a large clast of anorthosite (white). (B) Aphanitic impact melt rock from the Moon. Apollo 17 sample 73,217. Image: NASA. (C) Backscattered electron image showing microcrystalline texture of an aphanitic melt bead from Meteor Crater. Crystallites in this image are all clinopyroxene and show well developed quench crystal morphologies. The brighter clinopyroxene in the centre of the image is more Fe-Ni-rich than the surrounding phase. (D) Backscattered electron image aphanitic melt rock from the Polsingen locality, Ries impact structure. Crystallites in this image are all plagioclase (pale grey) and show well developed quench crystal morphologies. The dark grey material is glassy mesostasis.
over 200 known lunar meteorites. The most common lunar glass-bearing impactites are regolith breccias (e.g., McKay et al., 1986), which contain glass spherules and agglutinates. The latter are µm- to mm-size highly vesiculated glass fragments produced by micrometeorite bombardment of the lunar regolith. As noted earlier, impact glass is also found on Earth as isolated melt beads in proximal ejecta blankets, as tektites and spherules in distal ejecta deposits, and as a product of airbursts.

As with volcanic rocks, most glassy impact melt rocks and clasts within melt-bearing breccias are not 100% glass. The presence of µm-size microlites, most commonly clinopyroxene and plagioclase, and more rarely orthopyroxene and olivine, displaying quench crystal morphologies a common feature of impact glasses (Fig. 13C, D). Somewhat controversially, recent microanalytical studies of impact glasses from the Ries impact structure have suggested that enigmatic tubular crystal-like structures are not quench crystals but are microbial trace fossils of biological origin (Sapers et al., 2014, 2015). Identical tubules are found in volcanic rocks, where they are similarly interpreted to be biological in origin (e.g., Furnes et al., 2007; Staudigel et al., 2008).

As with aphanitic impact melt rocks, mineral and lithic clasts in impact melt glasses are common. In impact glasses, however, internal textures are more complex with globules and schlieren of partially to completely molten phases and intricate textures being common-place. Vesicles are also common (e.g., Fig. 13C). Impact glasses, like volcanic glasses are metastable. They are frequently the first phase or first product of the melt (e.g., Marion, 2009); vesicles were also noted close to large clasts due to more rapid cooling of the melt around the ‘cold’ incorporated clasts. As in volcanic rocks, vesicles are commonly lined or completely filled due to post-impact alteration to form amygdales and giving the glass an amygdular texture (Naumov, 2005; Osinski et al., 2013).

5.4. Clastic and particulate

The term pyroclastic is well known and synonymous with explosive volcanic eruptions. Derived from the Greek words pyro, meaning fire, and klastos, meaning broken, pyroclastic rocks comprise a mixture of volcanic glass (pyroclasts) and fragments of entrained country rock (Fisher and Schmincke, 1984). The pyroclast portion is given the terms fine ash (<0.063 mm), coarse ash (0.063 mm < 2 mm), lapillus (2 mm < 64 mm), and block or bomb (>64 mm), depending on the clast size. When lithified, the terms fine and coarse ash tuff, lapilli tuff or lapillistone, and agglomerate or volcanic breccia are applied. Such terms have been applied to impactites in a few instances (Ames et al., 2002). Various definitions of pyroclastic rocks in the literature, but whether it is implicit or explicit in the definition, there is the interpretation that they are associated with volcanism. Thus, to avoid any confusion the terms tuff, lapillistone, and agglomerate are to be discouraged from use for impactites. Instead, impactites with a clastic texture should be termed impact breccias (see Section 3), more specifically lithic impact breccias, if they are melt free, or impact melt-bearing breccias, where the melt is present as clasts (e.g., Fig. 6C). The latter have also been referred to as “suevite” in the literature; however, it has been noted for 40 years that there are issues with the definition and application of the term.
“suevite” (Grieve et al., 1977), issues that remain to the present-day (Grieve et al., 2015).

It is beyond the scope of this current review to address the origin and classification of impact melt-bearing breccias, beyond stating that the definition of the term “suevite” has evolved through time and that there remain considerable ambiguities. In the original definition, a “suevite” was defined as a melt-bearing breccia with a purely clastic matrix (Stöffler, 1977). In the revised definition the term clastic was replaced with particulate matrix (Stöffler and Grieve, 2007). The term “particulate” and its usage were not, however, defined. Osinski et al. (2008a) proposed that the term “particulate” be applied to impactites “where the groundmass is composed of a heterogeneous mixture of melt phases and there is evidence for these phases being fluid during and after transport”. We contend that both are found in impact craters on Earth. For example, a clastic matrix and the presence of clearly angular glass clasts is evident in Fig. 15A and B; whereas, the matrix in Fig. 15C and D comprises a mix of melt phases that remained at least partially molten after deposition – confirming to the definition of particulate. In essence, impactites with a particulate texture are impact melt rocks but they differ from coherent crystalline silicate impact melt rocks that display the straightforward aphanitic or phaneritic textures. In these particulate-textured melt rocks, there are several melt phases that are dispersed and intermingled to give a very heterogeneous appearance to the matrix or groundmass (Fig. 15C, D).

Regardless of issues concerning origin, impact melt-bearing breccias are the most common melt-bearing impactite found in meteorite impact structures on Earth. They are known from crater-fill deposits, proximal ejecta (Fig. 8C) and injection dykes (Fig. 8B). They display an extensive array of textures, colours and relative proportions of lithic versus melt content (Fig. 15), which as discussed earlier has led to considerable ambiguity and controversy surrounding their classification and nomenclature.

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**Fig. 14.** Vesicular impact melt rocks. (A) Highly vesicular impact melt rock from the Mistastin Lake impact structure. 7 cm diameter lens cap for scale. (B) Plane-polarized light photomicrograph of the same vesicular melt rock shown in (A).

**Fig. 15.** Backscattered electron (BSE) images of clastic and particulate textures. (A) Flow textures in a glass (“gl”) clast (bottom right) are clear truncated, with the matrix comprising a mix of fragmented mineral and lithic clasts. Sample from the Mistastin Lake impact structure. (B) Similar to (A) the large glass clast (centre) and matrix appear clastic in this sample from the Ries impact structure, Germany. (C) and (D) The matrix of these two samples from the Ries impact structure can best be described as particulate, comprising a series of melt phases (including calcite melt, “cc”) displaying immiscible textures, globules, and quench-textured crystallites (inset in (D) (for a detailed discussion see Osinski et al., 2004). Many of the formerly glassy impact melt particles have devitrified into clay (“cl”).
6. Differentiation of impact melts

One of the fundamental concepts of igneous petrology is magmatic differentiation, whereby a parent magma can evolve, resulting in a range of diversified products (Wilson, 1993). A variety of processes have been invoked that may result in differentiation. In general, these processes may be divided into two main types: crystal-liquid (or fractional crystallization) and liquid-liquid fractionation, with the former being considered most important for endogenic igneous rocks (e.g., Mc Birney, 2006; Wilson, 1993). In general, the majority of impact melts cool too quickly to differentiate; however, there are three notable exceptions on Earth (Manicouagan, Morokweng, and Sudbury) and physics suggests differentiation of impact melts on other planetary bodies is likely.

6.1. Manicouagan: an example of crystal-liquid fractionation

The most common process whereby endogenic magmas undergo differentiation is by crystal-liquid fractionation, also known as fractional crystallization (Wilson, 1993). One example of this process having caused differentiation in an impact melt is from the Manicouagan impact structure, an ~100 km diameter impact structure located in Quebec, Canada. The impact melt sheet is well-exposed, relatively accessible, and un-deformed. The majority of the impact melt sheet is an undifferentiated, relatively clast-poor to clast-free homogenous fine- to medium-grained quartz monzodiorite ~300 m thick (Floran et al., 1978). A study by O’Connell-Cooper and Spray (2011), however, on ~10 km of drill core through the melt sheet revealed that the melt sheet varies considerably in thickness, due to topography on the crater floor. In the thickest melt sheet section, which is ~1400 m thick, there is evidence for differentiation through fractional crystallization. This section is roughly divided into two units based on their clast content: 1045 m of clast-free to clast-poor impact melt overlying 425 m of clast-laden impact melt (O’Connell-Cooper and Spray, 2011). The clast-free to clast-poor rock is differentiated into three layers: a monzodioritic lower zone (525 m) overlain by quartz monzodioritic middle zone (244 m), and a quartz monzonite to quartz monzodiorite upper zone (276 m). Similar to the so-called Sudbury Igneous Complex (see below), the trace element ratios (e.g., La/Yb, Rb/Sr, U/Th) of each unit are very similar, which support a common source for both the differentiated and undifferentiated units. The Mg# and anorthosite content of plagioclase both decrease from the base of the lower zone to the top of the middle zone; whereas, the middle and upper zones are enriched in REE, suggesting that the middle and upper zones were likely the last to crystallize.

6.2. Sudbury: a complex case of crystal-liquid fractionation and liquid-liquid immiscibility

The so-called Sudbury Igneous Complex (SIC) is the remnant of the coherent impact melt sheet at the Sudbury structure (Grieve et al., 1991; Lightfoot, 2016). It is also the site of world-class Ni-Cu and PGE metal sulfide ores, with pre-mining resources associated with the SIC estimated at over $1.5 \times 10^9$ t of 1.2% Ni, 1.1% Cu and 1 g·t$^{-1}$ combined Pd + Pt (Farrow and Lightfoot, 2002). The SIC differs from most other terrestrial coherent impact melt sheets in that it is clearly differentiated – due to its ~2–3 km thickness – and it is relatively, but not completely, clast free (Fig. 11A, B). For example, there are rare quartz clasts with partially annealed PDFs (Therriault et al., 2002). The details of the mineralogy and geochemistry support a cognate source for the sub-units of the so-called Main Mass of the SIC, produced by fractional crystallization of a single batch of silicate liquid (e.g., Lightfoot et al., 1997; Therriault et al., 2002; Warner et al., 1998). The conclusion that the SIC and its ores are crustal in composition is borne out by isotopic studies (e.g., Dickin et al., 1992, 1996; Faggart et al., 1985; Walker et al., 1991). Osmium isotope studies of sulfides from several mines have confirmed their crustal origin from a binary mixture of Superior Province and Huronian metasedimentary target rocks (Morgan et al., 2002).

The SIC outcrops at the surface as an elliptical (~60 × 30 km) differentiated body of norite (~25%) overlain by quartz gabbro (~15%) and capped by granophyre (~60%; Fig. 11A, B) (Naldrett and Hewins, 1984). It should be noted, however, that based on their geochemistry the “norite” and “quartz gabbro” designations are misnomers, as they are more felsic than endogenic terrestrial norites and gabbros, with the SIC itself having an average composition corresponding to granodiorite (Therriault et al., 2002). Despite the variations in chemistry and mineralogy, these units share similar ratios of incompatible elements such as La/Y, Th/U, and Th/Nb, supporting the hypothesis that the units formed from a common parental melt (Lightfoot et al., 1997). The lower norite unit grades upwards into the quartz gabbro over several metres, where orthopyroxene begins to disappear and cumulus magnetite occurs (Lightfoot et al., 1997; Therriault et al., 2002). This transition is marked by an increase in CaO, P2O5, TiO2, V, and Sc, and a marked increase in incompatible elements. The quartz gabbro gradually transitions into the granophyre over several 10s of metres, marked by the absence of magnetite and the increase in the proportion of quartz and plagioclase (~25%). The granophyre itself is weakly differentiated, the uppermost granophyre containing higher abundances of plagioclase (~35%), and containing increased proportions of TiO2, P2O5, MgO, and CaO, and the concentration of incompatible elements falls slightly towards the top of the granophyre.

There is an alternative working hypothesis for the original nature and lithological layering of the SIC; namely that it was originally an impact melt with discrete parcels of felsic (from melting of upper crust) and mafic (from melting of lower crust) melt that separated according to their densities to form norite and granophyre, with quartz gabbro representing a mixed transition zone (Zieg and Marsh, 2005). This is considered an unlikely hypothesis (cf., Latypov, 2009), given the ample evidence for the mixing of diverse target lithologies upon impact melting at other terrestrial impact structures (Dence, 1971; Dressler and Reimold, 2001; Grieve et al., 1977; Osisnki et al., 2012). This hypothesis also draws heavily upon analogies with endogenic layered maﬁc igneous complexes, which are poor analogues, not just in terms of initial composition compared to the SIC but also in terms of the highly dynamic environment that exists during the formation of impact melts. As the target rocks are melted following the passage of the rarefaction wave in an impact event, the resultant superheated and, thus, low viscosity melts are being driven down into the expanding transient cavity in highly turbulent flow, with differential particle velocities, which are a function of the shock pressure of several km/s (see Fig. 2) (Grieve et al., 1977). As a result, initial compositional variations within the volume of target rocks that are melted are overwhelmed and the impact melt is of a mixed composition. The hypothesis is also incompatible with the nature of the recently identiﬁed roof rocks of the SIC, the so-called Upper Contact Unit (Anders et al., 2015). These chilled variants of the SIC have a fine grained, homogeneous igneous matrix, with a composition close to that of the average SIC. They show no petrographic evidence of discrete felsic and mafic melt pockets, as explicitly predicted in the hypothesis of Zieg and Marsh (2005).

6.3. Liquid immiscibility

Liquid immiscibility sensu stricto describes the process whereby an initially homogeneous melt reaches a temperature at which it is no longer stable and it unmixes into two liquids of different composition and density (Roedder, 1955). It is relatively rare in pure silicate systems (Roedder, 1979) but is thought to play an important role in the formation of carbonatites (e.g., Church and Jones, 1995; Minarik, 1998; Treiman et al., 1985). As noted above, evidence for the melting of carbonates during hypervelocity impact has been recognized at a number of craters on Earth (Osisnki et al., 2008b). Graup (1999) first proposed that “carbonate-silicate liquid immiscibility” occurs upon impact
melting and this term was then used by numerous other authors, including the lead author of this contribution. However, as discussed by Osinski et al. (2008b) the term “carbonate-silicate liquid immiscibility” should not be used unless there is unequivocal evidence for the unmixing of an originally homogeneous impact melt. Actual evidence for the unmixing of an initially homogeneous impact melt has not been documented in the literature to-date. Rather, it is much more likely that some impact-generated melts derived from target rocks at different stratigraphic levels and with widely divergent initial compositions (e.g., limestone versus granite) never completely mixed or homogenized because of their distinctly different physical and chemical properties. The result is that immiscible textures and evidence for the interaction of immiscible melts is known from some impact melt rocks and impact melt-bearing breccias (Fig. 16) but the evidence for the actual process of liquid immiscibility is currently lacking.

As with endogenic igneous rocks, for pure silicate systems (i.e., impacts into relatively homogenous targets such as the Canadian Shield), the similar properties of the melts will promote mixing and homogenization and result in only rare immiscible textures (e.g., Dence et al., 1974) and the formation of igneous rocks with a relatively homogenous appearance. As noted above, given the samples studied by Dence et al. (1974) and the formation of igneous rocks with a relatively homogenous appearance. As noted above, given the samples studied by Dence et al. (1974) are in glass it is unclear if these “immiscible” textures are the result of liquid immiscibility sensu stricto, or rather due to incomplete mixing. In contrast, and also in keeping with endogenic igneous systems, for impacts into sedimentary targets where rocks as diverse as limestone to sandstone are melted, mixing and homogenization is inhibited and immiscible textures are more common, thereby resulting in rather heterogeneous overall textures. Examples are provided and discussed in the following section.

The recent work at Sudbury can mostly be fitted into the framework of the formation of an approximately ~200 km impact basin at 1.85 Ga. As noted above, this impact resulted in massive (~10^4 km^3) crustal melting producing a superheated melt (the SIC) of an unusual composition, which produced immiscible sulfides. Recently, Farrow and Lightfoot (2002) and Ames and Farrow (2007) reviewed the nature of the ore deposits at Sudbury and placed their formation in an integrated time-sequence model. They recognized: Ni-Cu-Co “Contact” deposits associated with embayments at the base of the SIC and hosted by so-called Sublayer and Footwall Breccia; Ni-Cu-Pt-Pd-Au “Offset” deposits associated with discontinuities and variations in thickness in the so-called Offset Dikes; and Cu-Pt-Pd-Au-rich “Footwall” deposits that can occur in the underlying target rock, up to 1 km away from the SIC. They also recognize a fourth deposit environment associated with impact-generated pseudotachylitic, so-called, Sudbury Breccia (O’Callaghan et al., 2016). The Contact deposits consist of massive sulfides and are volumetrically the largest deposit type, hosting approximately 50% of the known ore resources. The Cu-PGE-rich Footwall deposits are volumetrically small (~10% of known ore) relative to the Contact deposits, but are extremely valuable bodies, as they are relatively enriched in PGEs, in addition to copper.

The origin of the sulfides at Sudbury are generally ascribed to the formation of an immiscible sulfide liquid, its separation from its silicate host and settling to the base of the SIC to form the Contact deposits, with remobilisation into Offset, Sudbury Breccia and Footwall deposits, with a potential hydrothermal component to the latter. In a general sense, the sulfide ores are similar in generation to those formed by endogenic magmatic processes and their formation has been compared to them directly (e.g., Barnes et al., 2017). There are, however, some important differences in that impact melt rocks do not have the initial compositional limitations of endogenic melts (thermodynamically constrained partial melts of compositionally limited protoliths) and are, by their inherent nature, initially superheated. As sulfur is an incompatible element in sulfide undersaturated melts, fractional crystallization of silicates is a major process whereby endogenic melts become saturated and develop immiscible sulfide liquids.

In the case of the SIC, however, it is believed that its initial composition and superheated nature contributed to its ability to contain as much as ~5 times dissolved sulfur than it could if it was at its liquidus temperature. As a result, immiscible sulfide liquids segregated initially from the silicate melt at temperatures some 250 °C above the silicate liquidus and continued to do so as the melt cooled and crystallized silicates (Keays and Lightfoot, 2004). While mass balance calculations of the initial composition of the SIC indicate that it contained more than enough metals to form the sulfide ore bodies, modeling suggests an initial melt with higher Ni, Cu and PGE contents and different Cu: Ni ratios. To account for this, modeling by Keays and Lightfoot (2004) includes a secondary component, wherein the immiscible sulfide liquids from the upper reaches of the SIC are re-adsorbed as they sink under gravity into the lower (hotter) reaches of the SIC, only to become immiscible again on further cooling of the SIC. Isotopic data are consistent with the lower units of the SIC (“norites”) being strongly oversaturated with respect to sulfur (e.g., Darling et al., 2010). They also indicate lateral heterogeneities in the isotopic composition of the sulfide ores, which can be accounted for by local composition complexities due to the assimilation and melting of clasts and local footwall rocks by the superheated SIC. In addition, the geometry of the SIC, as a thin (2.5–3 km thick) sheet with an estimated original diameter of ~100 km, suggest that convective cells in the cooling melt would have lateral dimensions of only a few kilometres (~2 times their vertical dimension) and any compositional variability would be maintained throughout the SIC’s cooling history.

6.4. Assimilation

In addition to crystal-liquid and liquid-liquid fractionation, the assimilation of wall rocks has been proposed as a process capable of
changing the composition of magmas (McBirney, 1979). As discussed by Wilson (1993), the importance of assimilation in magmatic differentiation has been debated for well over a century; although the general consensus is now that “assimilation coupled with fractional crystallization (AFC) is an important process in the petrogenesis of many continental magmas”. As, in addition to raising the temperature of any wall rock clasts, additional heat is required to satisfy the latent heat of melting, a critical limitation of assimilation is the amount of heat available from a magma. In endogenic magmas, superheating of magmas is considered to be rare (McBirney, 1979).

As noted earlier, in impact-generated melts, superheating is the norm, with evidence for initial temperatures in excess of 2370 °C (having been documented (Timms et al., 2017) (see also Section 2). Evidence for assimilation of clasts in impact melts is widespread. The mass of clasts that can be assimilated and, therefore, the final composition and clast content of the melt depends largely on its subsequent thermal history, which varies with the size of the impact event, which affects the volume of melt produced, and the setting of the melt, which controls how quickly the melt subsequently cools (e.g., dykes versus crater-fill). In addition, the nature of the target rocks and the overall heterogeneity of the target will greatly affect how much the composition of the original melt and final crystallized melt differ. For example, for a crater in a relatively homogenous target, the compositions of the clasts in the melt are likely to be similar to the composition of the melt such that the composition of the melt will not be greatly affected.

There is, however, a caveat to this statement. It is an observation that, in detail, the mineral clast content and their nature does not accurately mirror the mineral content of the target rocks and varies with stratigraphic position in impact melt sheets in Shield terranes. For example, mafic mineral clasts are under-represented in the resultant impact melt rocks, presumably as a result of the fact that mafic minerals are more compressible than tectosilicates and tend to thermally dissociate under shock (Stöffler, 1972). The mineral clast content of such melts, therefore, is dominated by quartz and feldspar. With increasing stratigraphic height (corresponding to longer cooling times) in such melt sheets, overall clast content decreases, with feldspar clasts being preferentially assimilated, until only quartz clasts remain (e.g., Phinney et al., 1978).

In addition, for craters developed in mixed sedimentary-crystalline targets, an initial impact melt with a broad crustal composition could entrain clasts as chemically diverse as sandstone to limestone. In general, the cratering flow field, particularly in larger craters, can result in the melt in one region of the crater entraining a different clast population than the melt in another region, which could be many 10s of km distant, resulting in chemical variations in the final crystallized melt within the crater. This may be reflected in the small variation in alumina content (~1%) between the northern and southern portions of the impact melt sheet at Manicouagan, which covers an area of 2400 km² and where anorthosite is confined to the northern portion of the target (Floran et al., 1978). As noted earlier, target lithology can be a further complicating factor, with impact melts derived from impacts into sedimentary assimilating fewer clasts, resulting in higher final clast contents as opposed to impact melts derived from crystalline targets (Osinski et al., 2008a).

6.5. Differentiation of impact melts on other planetary bodies?

The ~3.7–3.8 Ga ~930 km diameter Orientale impact basin on the Moon is an example where a differentiated impact melt sheet (or in this case an impact melt sea) may potentially occur. Vaughan et al. (2013) estimate that the Orientale impact event resulted in an impact melt sheet that was ~15.5 km thick (solidified to approximately 14 km thick) and ~350 km in diameter, with a total volume of ~10⁶ km³. This volume of impact melt is expected to have undergone large-scale igneous differentiation. Vaughan et al. (2013, 2014) produced a three-component model in order to estimate the differentiation of the impact melt sea: a mixing model to determine the original composition of the melt; a thermodynamic model to estimate the crystallization sequence and; a fluid mechanical model to convert this crystallization sequence into a stratigraphic sequence. This model estimated that the impact melt sea could have crystallized into three major units: an ~2 km thick layer of dunite overlain by ~4 km thick layer of pyroxenite and capped by ~8 km thick layer of norite (Vaughan et al., 2013; Vaughan and Head, 2014).

A recent study, however, by Cassanelli and Head (2016) has challenged this working hypothesis and proposed that the convection process would have enhanced cooling and crystallization, inhibited the settling of crystals less than approximately 1.5 mm in diameter, and prevented differentiation at or below this crystal size. Given that the majority of lunar clast-poor impact melt rocks are fine-grained, Cassanelli and Head (2016) proposed that impact melt sheet thickness is predicted to have minimal influence on crystallization history, suggesting instead that crystal size is the primary factor controlling lunar impact melt sheet differentiation. None of the Apollo missions, however, were designed to sample “melt sheets” and so planetary impact melt rocks may be more common on the Moon than the current lunar sample collection would suggest.

7. Geochemistry

7.1. Impact melt rocks

The chemistry of endogenic igneous rocks is constrained by the thermodynamic processes of partial melting of protolith rocks that govern magma formation and subsequent evolution, by processes such as fractionation. In contrast, there is no constraint on the composition of an impact-generated melt rock beyond the composition of the protolith target rocks. Fig. 17A provides a plot of average impact melt compositions and target rock compositions from nine complex craters that formed in various target lithologies. Impacts into primarily crystalline rocks generally produce impact melt rocks with compositions that broadly represent the average continental crust in that region. For example, the Boltysh impact structure formed in crystalline Precambrian target rocks and subsequently formed a homogeneous impact melt sheet with compositions corresponding to a mixture of the target lithologies (Grieve et al., 1987) (Fig. 17A). Similarly, impacts such as El'gygytgyn that formed primarily in pyroclastic rocks result in impact melts of similar composition (Val'ter et al., 1982) (Fig. 17A).

Even impact structures that formed in mixed target rocks (i.e., dominated by both crystalline and sedimentary rocks) can result in the formation of a geochemically homogeneous impact melt sheet. For example, the basement rocks of the impact melt rocks from the Manicouagan impact structure can be modelled as a statistically valid and geologically reasonable mixture of the highly variable target lithologies, which include amphibolite to granulite facies metagabbro, anorthosite, mafic to granitic gneiss units, and metasedimentary lithologies (Fig. 17A) (Grieve and Floran, 1978; O'Connell-Cooper and Spray, 2011). In extreme cases such as the Sudbury impact structure (Fig. 17B), two distinctly different provinces – Archean granitoids, gneisses, and greenstones to the north and Paleoproterozoic metasediments and metavolcanics to the south – melted to form the Sudbury Igneous Complex (Therriault et al., 2002). This ultimately resulted in a generally homogeneous impact melt sheet that represents the bulk average of the local crustal rocks (Fig. 17B), but has since undergone significant differentiation, as described in Section 6.2. In summary, there are two major observations: (1) impact melt rocks from these craters broadly represent an average of the continental crust in that region, and (2) impact melt rocks at these sites generally show a large degree of compositional homogeneity (e.g., Grieve et al., 1977; Floran et al., 1978; Marion and Sylvester, 2010; Simonds et al., 1978); although the process of differentiation in the thicker melt sheets can complicate matters (Fig. 17B).
The impact melt at the Mistastin Lake impact structure is somewhat of an exception, as it has some degree of compositional heterogeneity at the micro- and macroscopic scale (Fig. 18A) (Grieve, 1975; Marion and Sylvester, 2010). The Mistastin Lake structure is located in the Mesoproterozoic Mistastin Batholith, which consists primarily of anorthosite and granite with minor gabbroic rocks, and is surrounded by Proterozoic amphibolite, granitic and migmatitic gneisses (Emslie et al., 1980). Marion and Sylvester (2010) modelled the matrix compositional heterogeneity, expressed as melt rocks with HFSE and LFSE (high field strength elements) in Fig. 18B, as due to the assimilation of locally abundant mangerite clasts – which are more enriched with HFSE than the other target lithologies – during the post-impact cooling of the melt sheet.

Terrestrial impact melt rocks have in some cases enriched abundances of siderophile, particularly platinum group (PGE), elements over their respective target rock (e.g., Goderis et al., 2012). These enrichments are the result of a small admixture (generally <1 wt%) of meteoritic material from undifferentiated impacting bodies. Early work required detailed knowledge of the abundances in the target rocks to determine the exact identification of the impactor but it has been demonstrated that such an “indigenous” correction for the target rock contribution to PGE abundances in melt rocks is not necessary, if inter-element ratios and a mixing line is defined through regression analyses are used (Tagle and Hecht, 2006). Chromium and Os isotopes have also been used to define a meteoritic component in impact melt rocks (e.g., Koeberl et al., 2012). The use of this geochemical signature of meteoritic material in suspected impact melt rocks is not a general methodology for the identification of impact melt rocks in the terrestrial environment, because of their physical association with an identifiable impact structure with shock metamorphism. This not the case, however, for extraterrestrial samples with little or no known field context and this methodology has been used when classifying, for example, igneous-textured samples in the Apollo collection from the moon (e.g., Norman et al., 2002).

Geochemical homogeneity is generally not the case for impact melt rocks derived from craters formed in sedimentary target rocks. The same cannot be said for impact melt rocks derived from craters formed in sedimentary target rocks. They can have extreme compositions and be heterogeneous in composition at any given impact crater. For example, impact melt rocks with compositions ~90 wt% SiO₂ due to the melting of sandstone are known from the Gosses Bluff impact structure, Australia (Milton et al., 1996). At the Haughton impact structure, where the target was dominated by limestone and dolomite, the melt rocks possess a groundmass of igneous calcite and silicate glass that displays MgO contents up to ~25–30 wt% (Osinski et al., 2005; Osinski and Spray, 2001); glass compositions that are not only unusual but also not known even from endogenic carbonate igneous rocks.

7.2. Impact glasses

Unlike most impact melt rocks, impact glasses can be extremely heterogeneous in composition. It would be pointless to plot all known compositions of impact glasses, as the spread would be immense. In general, smaller craters and craters with heterogeneous target rocks, result in more diverse compositions of impact glasses. For example, at the 1.2 km diameter Meteor (aka Barringer) Crater, USA (sandstone, dolomite-dominated target), glasses with CaO, FeO, and MgO contents up to 22 wt%, 36 wt%, and 15 wt%, respectively, have been documented (Hörz et al., 2002; Osinski et al., 2015). At the 90 m diameter Wabar Crater, the glasses are SiO₂-rich (~90–95 wt%) (Hörz et al., 1989) reflecting the unconsolidated sand target. A further notable property of impact glasses found in small craters is their high projectile content (e.g., Fazio et al., 2016; Hörz et al., 1989, 2002).

8. How to determine the origin of extraterrestrial igneous rocks?

As described in Section 2, hypervelocity impact events generate pressures and temperatures that can melt a substantial volume of the target sequence. It is also now well known that impact cratering is a ubiquitous geological process that affects all Solar System objects with a solid surface. As noted at the outset of this paper, given that on many of these bodies, volcanism has also been prevalent, this begs the question how to distinguish between an impact versus an endogenic origin for rocks that appear igneous in origin?
8.1. A sample analysis approach

When samples are available for study in Earth-based laboratories, there are several unambiguous criteria that can be used to determine an impact origin of a morphological feature, outcrop or hand sample (Ferrière and Osinski, 2012; French and Koeberl, 2010): 1) shock metamorphic features in lithic or mineral clasts (e.g., shatter cones, planar deformation features (PDFs); diaplectic glass, high-pressure mineral polymorphs); 2) fragments of the projectile, either isolated or as lithic clasts; and/or 3) extraterrestrial chemical and/or isotopic signatures of the projectile in melted materials. In simple craters, fragments of the projectile can be found scattered around the rim and throughout the proximal and distal ejecta (e.g., at Meteor Crater; Shoemaker, 1963). If isolated bodies of igneous rock were found in close proximity one could posit that by way of association, the igneous rocks were impact melt rocks produced by the same impact event that deposited the meteoritic fragments.

Unfortunately for planetary scientists, the identification of most shock metamorphic features requires the use of high-resolution microscopes and/or a host of sophisticated analytical techniques on prepared samples that are typically not available on planetary exploration missions. A notable exception would be if one was lucky enough to image a shatter-coned clast in the igneous rock under investigation. Such shatter-coned clasts have been documented in impactites in crater-fall deposits, injection dykes, and proximal ejecta deposits around craters on Earth (Osinski and Ferrière, 2016). Clasts of the projectile in impactites are incredibly rare in complex structures (Morokweng is, to our knowledge, the only example; Hart et al., 2002) but are present in smaller simple structures (e.g., Fazio et al., 2016; Osinski et al., 2015).

It is important to note that the confirmation of a meteoritic origin for a clast would also require supporting geochemical or isotopic data. While the presence of meteorites as clasts in extraterrestrial igneous rocks has not been reported, the Mars Exploration Rover Opportunity has discovered meteorites on Mars (Schröder et al., 2012), so, in theory, this is possible. The identification of high-pressure mineral polymorphs is, in theory, also possible on other planetary surfaces. The two main instrumental techniques used on Earth to identify such shock criteria are X-Ray Diffraction (XRD) and Raman spectroscopy. An XRD instrument is already on Mars in the form of the CHEMIN instrument onboard the Mars Science Laboratory Curiosity rover (Grotzinger et al., 2012) and various Raman spectrometers are scheduled for launch in 2020 onboard the European Space Agency’s ExoMars rover and the NASA Mars “2020” rover.

What if the rock contains no visible clasts? As we have outlined in this review, many impact melt rocks lack visible clasts and can display textures that are typical of volcanic flows, such as columnar joints, vesicles, flow textures, quench textures, and so on. As described above, even on Earth, the similarities between impact melt rocks and endogenic igneous rocks led to several decades-long controversies about the origin of a number of what were enigmatic structures, and their attendant lithologies, in North America and Scandinavia that are now known to be the product of hypervelocity impact. Is there any way to unequivocally prove the impact origin of clast-free igneous rocks? If only remote sensing and/or image data are available then the answer is unfortunately “no”. If, however, samples are available for analysis, the answer is “probably yes”.

As discussed above, if an extraterrestrial chemical and/or isotopic signature of the projectile can be determined then it would provide strong evidence for the impact origin of the samples in question (see Goderis et al., 2012, for an overview). Another less robust approach is the chemistry of the rocks. Endogenic igneous rocks are constrained in their composition by the geothermal and geodynamic environment of the planet such that not all compositions are possible. While not unequivocal, very unusual glass compositions, for example approaching 100 wt% SiO2 (e.g., Milton et al., 1996) or 20–30 wt% CaO and MgO (e.g., Osinski et al., 2005, 2007), are difficult to explain, in some cases, by any other process other than hypervelocity impact. It is notable that still one of the best arguments for the impact origin of the Sudbury
Igneous Complex is that it has been shown through its bulk chemistry and analyses of various isotopic systems that it is entirely crustal derived, which is inconsistent with an endogenic origin for a 30 × 60 × 3 km thick igneous intrusion (Grieve et al., 1991).

Another novel approach that has been tested on lunar samples is crystal size distributions as a way to distinguish between igneous and impact melts (Fagan et al., 2013; Neal et al., 2015). In brief, the slopes and intercepts of these crystal size distributions (CSD) demonstrated that olivine from impact melts displays a steeper CSD relative to olivine from mare basalts, with the opposite being the case for plagioclase (Neal et al., 2015). This thin section-based approach, thus, has significant potential for discriminating between coarse-grained igneous rocks of impact and endogenic origin.

8.2. A remote sensing approach

What if all that is available are orbital data? The first step is obviously the geological context. The physical association of lithologies resembling igneous rocks with a crater form, with the appropriate morphology and morphometry for its dimensions, is generally all the information available and is generally used to infer an impact origin for the crater and its associated deposits (Tornabene et al., 2013). For airless and volatile-poor bodies such as the Moon and Mercury, where there is an absence of erosion and transport – beyond the action of the solar wind and other subsequent impacts – the interpretation of impact melt rocks in crater interiors and proximal ejecta deposits is a relatively straightforward approach. The main complications are that craters degrade with age as the regolith develops, and for larger craters mare deposits (i.e., lava flows) often infill part of the interior. However, the Fe-rich composition of most lunar basalts provides a relatively straightforward way to discriminate between these two types of “igneous” rocks on the Moon (e.g., Fig. 19). Distinguishing between impact and endogenic volcanic deposits on Mercury has been shown to be less straightforward (Denevi et al., 2013). Radar data as a measure of surface roughness have also been shown to be an excellent way to identify and map impact melt rocks on the Moon (Carter et al., 2012; Neish et al., 2014) and Venus (Grieve and Cintala, 1995); although the reason(s) for this roughness remains unclear.

For Mars, the task is not as straightforward. First, there is the potential for many more geological processes that yield circular landforms and craterforms, in particular volcanism (e.g., maar craters, rootless cones, etc.), tectonism, and periglacial activity (e.g., pingos, thermokarst collapse depressions, etc.) (Carr, 2006). For example, recent workers have suggested that large circular depressions on Mars that were previously assumed to represent impact craters might have formed due to caldera forming volcanic explosions based on morphometric studies using remote sensing data (Michalski and Bleacher, 2013; Williams et al., 2009); although these interpretations are not universally agreed upon. The second complication is the volatile-rich nature of the Martian crust. While a few lunar-like melt sheet examples have been recognized (e.g., Mouginis-Mark, 2015), impact melt deposits more commonly exhibit a distinctive pitted appearance (Tornabene et al., 2012) (see Section 4.1 and Fig. 10A, B). Tornabene et al. (2012) outline the three main lines of evidence for pitted materials being volatile-rich impact melt-bearing deposits. Briefly, while at first glance these pits may resemble primary or secondary impact crater they have no, or only subtle, raised rims and lack ejecta deposits. Second, the pits are also not preferential aligned as would be expected of volcanic or tectonic collapse features. Finally, the pitted materials primarily occur as ponded and flow-like deposits on crater floors, terraces, and infilling the lowest local topographic depressions atop the ejecta blanket, identical to the distribution of impact melt-bearing deposits on the Moon (e.g., Fig. 9).

The third major challenge for Mars is the action of cold climate processes at the present-day and in the recent past (e.g., periglacial activity, seasonal ice cap activity) and widespread evidence for an active hydrosphere on ancient Mars, which has led to the erosion, burial, and even exhumation of impact craters over geological time. Finally, as on other planetary bodies, subsequent impacts may redistribute materials over many thousands of kilometres. When taken together, caution is advised in the interpretation of the specific genesis of “igneous” rocks on Mars. This is particularly relevant for rover missions where many of the rocks being investigated are not in situ. It is also important for interpreting orbital mission data. Physics dictates that the large Martin
impact basins such as Hellas, Utopia, Isidis, etc. would all have produced large volumes of impact melt. Recent mapping of the ejecta blanket of Orientale has shown that impact melt flows reach dimensions of up to ~400 km long by 150 km wide and occur up to ~1350 km from the crater centre (Morse et al., 2018). Given the heavily degraded and infilled nature of Martian impact basins, there could be large exterior impact melt flows – that could appear like lava flows or sills – in the stratigraphy of the ancient Noachian crust potentially thousands of kilometres from the host crater. Without samples in hand (see Section 8.1), it would be virtually impossible to differentiate between an impact versus and endogenic igneous origin.

9. Concluding remarks

Over the past few decades it has become increasingly clear that impact cratering is one of the most important and fundamental geological processes in the Solar System. Indeed, meteorite impact craters are one of the most common geological landforms on the majority of the terrestrial planets, asteroids, and many of the rocky and icy moons of the inner and outer Solar System. If it were not for the ongoing actions of erosion, volcanic resurfacing, and plate tectonics on Earth, there would be many more than the ~190 impact structures that have been documented to date. Despite this relatively sparse impact cratering record, it is now apparent that impact events have profoundly affected the origin and evolution of Earth and of life itself and produced benefits in the form of economic mineral and hydrocarbon deposits.

The goal of this review has been to demonstrate that not only is the melting of often large volumes of rock a fundamental outcome of hypervelocity impact, but that these igneous rocks share close similarities with volcanic and shallow intrusive igneous rocks of endogenic origin. It is important to note, however, that the process of impact melting is distinctly different to endogenic melting. Impact melting occurs upon decompression from a shock compressed state and results in total melting of target lithologies – as opposed to partial melting in endogenic igneous rocks – and the initial impact melts are typically superheated. Despite these differences, as we have hopefully shown, many impact melt rocks can be described using the same textural terms applied to endogenic igneous rocks (e.g., phaneritic, vesicular, glassy, etc.). Impact melt rocks often also display columnar joints in outcrop. Furthermore, once formed, impact melts are also subject magmatic differentiation. Liquid immiscibility and assimilation of clasts is commonplace in all impact melts and in larger impacts, where more melt is generated, crystal–liquid fractionation is also an important driver of differentiation in impact melts. As we note at the outset of this review, historically, these similarities led to the misidentification of many impact melt rocks on Earth – and their host crater forms – as being of endogenic volcanic origin. In closing, we urge the planetary science community to keep this in mind when investigating igneous-textured rocks on other planetary surfaces. We also hope that this review will stimulate more discussion and collaboration between impact geologists and volcanologists both on Earth and throughout the Solar System.

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