IMPACT FLOWS AND CRATER SCALING ON THE MOON

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The asymmetric distribution of stress, internal energy and particle velocity resulting from the impact of an iron meteoroid with a gabbroic anorthosite lunar crust has been calculated for the regime in which shock-induced melting and vaporization takes place. Comparison of impact flow fields, with phase changes in silicates taken into account, with earlier results demonstrates that in the phase change regime when the 15-km/s projectile has penetrated some two projectile radii into the moon, the peak stress in the flow is ~0.66 Mbars at a depth of 66 km, and the stress has decayed to ~66 kbar at a depth of 47 km. Rapid attenuation occurs because of the high non-thermal velocity of the high-temperature phases associated with a 35% (in pressure) density increase. The feature of the phase-change flow tends to strongly concentrate the maximum shock pressures along the meteoroid trajectory (axis) and makes the conical zone along which high internal energy deposition occurs, both shallow and narrow. Examination of the gravitational energies required to excavate larger craters on the moon indicates the importance of gravity forces acting during the excavation of craters having radii in the range greater than ~2 - 140 km. It is observed that the "hydrodynamic" energy vs. crater radius relation approaches those for various "gravitational" energy vs. radius relations if the radii values corresponding to the larger scale basins. Crushing energy values in the range of (1.0 - 9.4) x 10^26 erg are inferred on this basis for the lunar crust. Using these values and the criteria that all rocks exposed to ~100 kbars of greater shock pressures are included in the ejecta (some of which fails back) implies that the maximum depth of sampling expected to be represented within the Apollo collection lies in the range 148 - 328 km.

1. Introduction

Quantitative description of cratering flows, when combined with knowledge of the kinematic parameters of impacting objects from man-made craters and the detailed data from terrestrial craters sampled in the subsurface, can provide information on the following questions:

(1) How is the range of objects impacting the moon constrained by the size and shape of craters observed?

(2) What is the depth in the moon from which the formation of a given crater has excavated material which may be sampled in the surficial ejecta? and, how does this depth depend on the equation of state and strength properties of the lunar crust and upper mantle?

(3) What is the ejecta distribution of lunar and meteoritical material from a given impact crater on the moon; and, what is the initial position and shock and thermal history of each element in the resulting ejecta blanket?

(4) How much rock and meteorite melt and vapor is produced upon impact; and, how does this depend on the type and speed of the meteoroid?

In the present paper, we address some of the above questions in the process of describing the cratering flow from a 15-km/s iron meteoroid impacting a gabbroic anorthosite crust. Revising an earlier formulation of this problem (O'Keefe and Ahrens, 1975) we now rigorously include, in a thermodynamic context,
details of the shock-induced phase changes and irreversible behavior upon unloading from the shock state which we discovered several years ago (Abaren and Rosenberg, 1968; Abaren et al., 1969; and Grady et al., 1974). In practice the phase change has the effect of producing a very large hysteretic loop in the material upon shock loading even for non-precious silicates.

2. The impact process in silicate rocks

When a meteoroid impacts the lunar surface, an intense shock wave is driven forward into the surface and rearward into the meteoroid. Upon reflection of the rearward propagating shock at the upper surface of the meteoroid, a forward traveling rarefaction wave propagates through the meteoroid and subsequently into the lunar surface. This rarefaction wave, and rarefactions emanating from the lateral surfaces of the meteorite and lunar surface overtake the diverging shock wave, reduce the shock pressure in each mass element of the lunar surface.

The states which result from a hypervelocity impact event in a typical silicate material are shown in Fig. 1. This representation illustrates the observation that upon shock compression, states corresponding to the initial or low-pressure phase (lPp) are achieved up to pressures on the order of \( \sim 130 \) – \( \sim 400 \) kbar, above which a phase transition(s) in the silicates occur. At successively higher pressures, states in a mixed phase regime in which, with increasing pressure (over a range of \( \sim 130 \)–\( 700 \) kbar) coexist with the lPp. At still higher pressures, the Hugoniot curve is representative of pressure (\( P \)), volume (\( V \)) and energy (\( E \)) states of a denser polymorph(s) of the mineral, or mineral assemblage. In

![Image of graph showing shock wave behavior](image-url)

**Fig. 1.** Equation of state of phyllosilicate assemblage in low (lPp) and high-pressure (lPP) regimes. log \( P \) (pressure) vs. compression for \( 2.86 \times 10^{4} \text{g/cm}^3 \) basalt used in Gaeth and Helou (1963) (C & H) formulation is shown also for comparison. In the C & H treatment release adiabats are assumed to coincide with Hugoniot curve.
the high-pressure phase (hph) region the entropy generated in the shock process increases in significance relative to the entropy generated upon release. At shock pressures in the range 0.5 — 0.7 Mbar, melting of most rock occurs upon release and at a shock pressure in the range of 1.0 — 2.0 Mbar vaporization occurs upon release (Abrens and O'Keefe, 1973).

3. Equations of state

We have adopted the Tillingston (1962) equation of state for describing impact flows. This formulation has the properties that at densities greater than normal, it behaves as a Max-Greenwood thermal equation of state at low and moderate pressures (2-10 Mbar), while at extreme pressures, it functionally approaches a Thomas-Fermi equation of state. In addition, at densities less than normal and internal energies greater than the energy of vaporization, the equation of state approaches that of a polytropic gas.

In the compressed region (ρ > ρ₀) the equation of state is given by:

\[
P = \frac{\rho}{\rho_0} + \lbrack \frac{E_0}{(E_0)^{\frac{1}{\gamma}} - 1} \rbrack \cdot \frac{E_0}{(E_0)^{\frac{1}{\gamma}} - 1} + \frac{\rho}{\rho_0} \cdot \frac{E_0}{(E_0)^{\frac{1}{\gamma}} - 1} \cdot \frac{\rho}{\rho_0} \cdot \frac{E_0}{(E_0)^{\frac{1}{\gamma}} - 1}
\]

where \( P \) is pressure, \( \rho_0 \) is the initial density, \( E \) specific internal energy, \( \gamma \equiv \frac{\rho_0}{\rho_0} \) and \( \mu = \gamma - 1 \). Here (a + b) and A are the zero-pressure and temperature, Guinierloid parameter and bulk modulus, respectively. The constant, \( a \) was taken to be 0.5 so that at high-energy densities the thermal pressure would approach that of an electron gas. The other two constants were obtained by a nonlinear regression analysis of Hugoniot data. The low-pressure Hugoniot data for a gabbroic breccia are based on an experimental measurement on sample 13,416 (Abrens et al., 1973). Three additional very high pressure Hugoniot points were obtained from Lasser's (1955) Thomas-Fermi calculation. The constants for the hpp equation of state were determined by calculating a high-pressure phase Hugoniot centered at normal density \( \rho_0 \) from high-pressure shock data for terrestrial gabbro (McQueen et al., 1967) and bulk modulus-density systematics (Abrens et al., 1973; Davies and Goffney, 1973) and fitting these states along with the Thomas-Fermi states.

For low densities and high temperatures such that \( \rho < \rho_0 \) the form used is:

\[
P = aE_0 + \frac{bE_0}{(E_0)^{\frac{1}{\gamma}} - 1} + \mu \exp \left( -\left( \frac{A}{(E_0)^{\frac{1}{\gamma}} - 1} \right) \right)
\]

where \( a \) and \( b \) were chosen so that the form approaches a polytropic equation of state at low densities and high specific internal energies with a polytropic exponent of \( (a + b) \). The expression provides a smooth fit from the condensed region to the gas regime with constant pressure and sound speed at \( \rho = \rho_0 \) along the release adiabat. The constants for the iron meteorite and the low- and high-pressure equations of state are also listed in Table I.

In order to provide a phenomenological description

### Table I

<table>
<thead>
<tr>
<th>Material</th>
<th>( \rho_0 ) (g cm(^{-3}))</th>
<th>( a )</th>
<th>( b )</th>
<th>( A ) (Mbar)</th>
<th>( B ) (Mbar)</th>
<th>( E_0 ) (Mbar cm(^3) g(^{-1}))</th>
<th>( E'_0 ) (Mbar cm(^3) g(^{-1}))</th>
<th>( n )</th>
<th>( \beta )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gabbro breccia</td>
<td>2.936</td>
<td>0.5</td>
<td>0.045</td>
<td>0.075</td>
<td>0.751</td>
<td>4.89</td>
<td>0.047</td>
<td>0.18</td>
<td>5.0</td>
</tr>
<tr>
<td>hpp *1</td>
<td>3.965</td>
<td>0.5</td>
<td>0.128</td>
<td>2.16</td>
<td>1.26</td>
<td>18.0</td>
<td>0.045</td>
<td>0.18</td>
<td>5.0</td>
</tr>
<tr>
<td>Iron *2</td>
<td>7.8</td>
<td>0.5</td>
<td>1.5</td>
<td>1.28</td>
<td>1.05</td>
<td>0.095</td>
<td>0.0244</td>
<td>0.002</td>
<td>5.0</td>
</tr>
</tbody>
</table>

*1 Fit to experimental data on sample 13,416 and Thomas-Fermi calculations. \( E_0 \) and \( E'_0 \) are estimated energies for complete and incipient vaporization at atmospheric pressure.

*2 Fit to Hugoniot data derived from elastic constant systematics and Thomas-Fermi calculations.

Two phase regime parameters: transition energy between the hpp and hph phases, \( E_T = 0.01325 \text{ Mbar cm}^3 \text{ g}^{-1} \), \( \rho_{hph} = 4.18 \text{ g cm}^{-3} \).

*3 Tillingston (1962).
of the shock-induced phase change we employ a formulation used by Horie (1966) in which the mass fraction of the hpp, f, is given by:

$$df/dt = (f_{eq} - f) t$$

(3)

where $f_{eq}$ is the equilibrium mass fraction of the hpp at a given state ($V, E$) and t is the characteristic phase-change time, which, in principle, should be related to some measurements of the time scale of phase changes involving 4 to 6 coordination of O with $\delta^{16}$ for Grady et al. (1974) have found a value of $r = 0.25$ μs which is consistent with time-resolution processes in the quartz-staurolite mixed phase system. We have adopted this value in the present case and point out that at small distance scales, L, such that $L < r$, the equation of state used is time-dependent. (Here ε is the characteristic sound speed over L.)

However, the scalability of the present calculated flows are not affected in practice. For example, the condition $L = r$, for $c = 10^3$ cm/s, and $r = 0.25$ μs, infers flow fields with characteristic size of ~0.025 cm or considerably smaller. Thus flows in this size range could not be scaled via the considerations of Section 5.

In equilibrium, the lpp and hpp regions are completely determined by their respective equations of state. In the two-phase region ($V_{pp} < V < V_{pp1}$) the volume and internal energy of the individual phases are related to the total volume and internal energy by:

$$V = (1 - f_{eq})V_{pp2} + f_{eq}V_{pp2}$$

$$E = (1 - f_{eq})E_{pp2} + f_{eq}E_{pp2}$$

(4)

For a given volume ($V$), internal energy ($E$) and equilibrium hpp mass fraction ($f_{eq}$), the internal energies at the phase lines ($V_{pp2}, V_{pp1}$) and in principle, the heat capacities, can be readily calculated. Using the Tollison form for the lpp and hpp equations of state, a single cubic equation in $E_{pp2}$ was solved numerically at the mesh point. Once $E_{pp2}$ was determined, the equilibrium pressure was calculated using either the lpp or hpp equation of state.

4. Computational procedure and flow calculations

In order to apply the above equation of state parameters to description of impact flows we use a modified version of the computational procedure of Hageman and Walsh (1976). This method employs an axially symmetric grid, fixed in space. The generalized mass, momentum and energy equations in finite difference form are solved for material within each grid cell. Successive time steps in order to provide definition of the meteorite-target interface and both material free-surfaces, a series of massless tracer particles initially embedded along all surfaces are followed in the course of the computation.

In the present calculations, referred to as problem 2, a spherical iron meteoroid was assumed to impact a gabbroic anorthosite lunar surface at a velocity of 15 km/s. This calculation is similar in many respects to that previously described (O'Keefe and Ahrens 1975, referred to as problem 1). However, in problem 1, the radius of the meteoroid was 23.3 km and in this calculation the radius is 5 cm. The size of impacting body was made relatively small in these calculations because we wish to follow the details of the phase-change kinetics within discrete cells in the flow, to assure ourselves that the computational procedures outlined were, in fact, numerically stable and also because of considerations of numerical viscosity. However, we believe the results of problem 2 can be scaled to very large objects (discussed in Section 5). In comparing the details of the flow at early times we used the flow fields of problem 1 and 2 (and cube-root energy scaling for the flow fields in the hydrodynamic regime) at peak pressures of 0.66 Mbar. This, of course, is a pressure regime sensitive to our phase change description. We chose the times indicated for Figs. 2 and 3, largely because we had complete flow fields available. We note that a characteristic velocity, given by dividing the meteorite radius by the elapsed time of the flow field description for Figs. 2 and 3, are 2.0 and 2.5 km/s, respectively. Hence for another type of comparison, it would have been useful to have complete flow field available for problem 1, somewhat later, or, conversely, for problem 2, earlier. The maximum numerical viscosity calculated from shock-wave thickness for the 23.2 km impact case, problem 1 (mesh dimension = 5.9 km) is $1.3 \times 10^{-12}$ Pa, whereas the maximum numerical viscosity for the present 5 cm impact is $8 \times 10^{-13}$ Pa (mesh dimension, 0.5 cm).

At 25.4 μs after impact the cratering flow is in what we term the transient cavity regime (Fig. 2). The meteorite has lined the growing cavity. The main shock has detached from the vicinity of the meteoroid and has decayed to 660 kbar at 0.06, and at an angle of 45°.
the shock pressure has reduced to approximately 300 kbar. It is the rapid lateral decay in the shock pressure that results in the on-axis elongation of the transient cavity. The initially high peak shock pressures at the meteoroid–lunar surface interface results in the highest thermal energy densities being produced in that region. The results of problem J indicate ~7.2 meteorite masses of lunar crust would be melted or partially melted (O'Keefe and Ahrens, 1975), whereas problem 2 yields values in the range of 24–33 meteorite masses melted or partially melted, again emphasizing the role of the equation of state. The internal energy deposited in the meteorite is sufficient to vaporize a significant fraction of its mass. Referring to Fig. 2, the radial velocity has positive maxima of 2.4 km/s at an angle of approximately 30° from the axis and the
vaporized meteorite is now expanding inward. The axial velocity field (Fig. 2) exhibits two distinct flow regimes. One of these regions is dominated by the on-axis position of the flattened meteorite penetrating into the lunar surface and the other is an annular region of highly shocked lunar material and vaporized meteorite moving out of the cavity. This latter material will become the initial ejecta.

It is interesting to examine the details of the early cratering flow which are affected by the shock-induced phase change in the silicate. The flow field of problem 2, described above, can be scaled to that for a nominal Mare Imbrium-type impact breached by problem J (O'Keefe and Ahrens, 1975) in which a 23.2-km-radius iron projectile is assumed as the impactor. The most significant difference between the two flows (Figs. 2 and 3) is the more rapid attenuation of shock pressure in the phase-change case (Fig. 2). Other flow parameters are compared in Table II.
5. Crater scaling considerations

Extrapolation of the present calculations of the cratering flow of problem 2 to very large distances, characteristic lengths and energy scales of interest on the moon, such as associated with the Albategnius basin, is fraught with uncertainties. The roles of gravity and strength effects, as well as, the limited data set for large impacts are discussed below as they pertain to scaling the present calculations. These considerations are applied in Section 6 to bound the maximum sampling depth of ejecta for an Imbrium-type event.

Examination of the conditions for which the characteristic Froude number:

\[ F_n = u^2g/l \]  

where \( u \) and \( l \) are the characteristic particle velocity and length dimension of the flow and \( g \) is gravity, provide some measure of the parameters of the cratering flow for which gravity plays a dominant role for the largest dimensions of interest in impact cratering on terrestrial planetary surfaces \((F_n >> 1)\). For a characteristic velocity in the range \( 10^4 - 10^6 \) cm/s, a gravity field of \( 10^{-3} \) cm/s² implies characteristic impact flow dimensions in the range \( \sim 10^{-5} - 10^{3} \) cm and hence crater dimensions \( \sim 1 - \sim 10 \) km for craters for which gravity effects are comparable to inertial effects in a flow. For the moon, because of the low gravity, this range is \( \sim 10^3 \) km.

Since the present calculations do not take gravity into account, scaling directly to the terrestrial or lunar case is invalid when the cratering dimensions are such that the Reynolds or Froude number becomes on the order unity. However, Chyba (1965) has shown that in addition to the one-fourth power scaling of crater dimensions with energy which is appropriate to \( F_n < 1 \) (discussed by Gault et al., 1975), density \( \rho \) and gravity \( g \) are related to dimensions \( D \) of the crater by:

\[ D = (\rho g F_n)^{1/4} \]  

Thus as Chyba (1965) and Gault et al. have pointed out for very large craters, i.e., \( F_n < 1 \), a scaling relation such as:

\[ \ln R = \ln A + (\ln E)/n \]  

where \( n = 4 \), should relate apparent crater radius, \( R \), to impactor energy, \( E \), whereas for smaller craters, \( n = 3 - s \) is appropriate. [The limiting case, for very large craters, where \( n = 4 \) is discussed below (eqs. 11 and 12).] The parameter \( s = 0.3 \) takes into account the surface energy associated with crushing of the resulting ejecta material. As the size of the crater increases, the effective value of \( s \) is expected to approach zero. Gault (1973) has observed the relation:

\[ \ln R = \ln[7.52 \times 10^{-4} (\rho g)^{1/4} (A/E)^{1/4}] (1/2.76) \ln E \]  

describes experimental data in the range \( 10^{-3} \) to 1 cm. Pechig et al. (1974) have obtained similar results for
crater radii in the \(10^{-3} \text{ - } 10^{-1} \) cm range. These results display only a weak dependence on projectile momentum vs. energy as \(p_0\) and \(p_1\) are the density of the projectile and target respectively. For \(p_0 = p_1 = 3 \text{ g/cm}^3\) the two data sets agree closely as demonstrated in Fig. 4.

As is also evidenced from this figure, these relations although derived largely for cratering experiments on non-porous targets, are in approximate agreement with the apparent radius vs. kinetic energy relations of a large number of 100-500 cm-radius impact craters observed in unconsolidated media with densities of 1.4 - 2.0 cm\(^3\) by Moore (1975).

Previously, in order to extend energy vs. crater dimension relations to the very large dimensions representative of the moon, particularly the mare craters, the largest of which have radii in excess of 300 km, data from both chemical and explosion craters have been utilized (e.g., Baldwin, 1963). The craters produced in the unconsolidated debris layer on the lunar surface, density \(\sim 2 \text{ g/cm}^3\) (regolith) resulting from the impact of the three Ranger spacecraft and two Apollo 14/15 boosters (Whitaker, 1972) provides some additional useful data. When these are combined with the largest of the nuclear craters, Teapot Eas, where the depth of explosive basin resulted in overturned lip similar to that of an impact crater (Shoemaker, 1961) the relation:

\[
\ln R = -1.96 (\pm 0.1) + (1.355 \pm 0.06) \ln E
\]

(10) is obtained. The different parameters in eq. 10 depend in detail on the relative weighting of the Teapot Eas and/or a similar-sized Snowball chemical explosive crater (Roddy, 1968). The value of \(n\) appears to be approaching that suggested by Nordyke (1961) and Vade (1961), \(n = 3.4\), for scaling dimensions of nuclear craters having radii in the 1-km range on the earth.
The gravitational energy required to just lift the crater volume to the height of the crater rim for a conically shaped crater is:

\[ E_g = \frac{GMm}{2a} \]

where \( M \) is the initial uncratered density, \( a \) is gravity, and \( m \) is the mass of the crater (specifically, at the initial surface elevation). For a spherical bowl-shaped crater, the gravitational energy is:

\[ E_g = 2\pi G \rho \int_0^a \left[ R^2 (a - R)^2 + \left( \frac{R}{2} \right)^2 \right] dR \]

where \( a \) is the radius of curvature of the crater:

\[ a = R \left( 1 + \frac{a}{R} \right)^2 \]

The geometrical factor, \( \alpha \), has been studied by Pike (1974) over an extremely wide range of lunar crater dimensions (~0.05 – 300 km) and is insensitive to the probability of craters, varying little between the highlands and mare basins. It is also found to be independent of crater radius for craters of less than 5 km. Larger craters all tend to have very much lower values of \( \alpha \) which are observed by Pike to depend on crater radius. We infer, as does Melosh (1977) and others, that the post-impact processes such as slumping, and igneous and isostatic rebound, the latter especially for the larger mare craters, have markedly reduced the value of \( \alpha \) from initial values of ~0.4. For purposes of discussing energy partitioning and formation of the initial crater we have adopted the \( \alpha = 0.4 \) value for present calculations.

In the case of the 6- to 8-m-radius Ranger impact craters, Fig. 4 demonstrates that the minimum gravitational energy is less than 1% of kinetic energy of the objects produced in these craters. We infer from the convergence of the data representing eqs. 11 or 12 and those representing the larger impact craters, e.g., eq. 10, that the minimum gravitational energy required for crater formation exceeds that inferred for scaling according to relations which largely reflect "hydrodynamically" controlled flows at crater radii of 2–50 km. Thus, depending on the radius assumed for, e.g., Mare Imbrium, 335–485 km (McGetchin et al., 1973) and the shape of the transient cavity assumed, minimum cratering energies are calculated from eqs. 11 or 12 to be in the range (5.1–47) \( \times 10^{22} \) erg. Furthermore, since the various "hydrodynamically" energy vs. radius curves approach the "gravitational" energy vs. radius curves for radii in the mare–basin range we infer that since these energies are basically additive, the actual cratering energies involved in forming the mare are at least a factor of two higher than that inferred from the gravitational minimum value. Thus cratering energy values in the range \( (1.0–9.4) \times 10^{22} \) erg are obtained for the Imbrium crater. The above values are used in Section 6 to estimate the maximum possible depth from which samples could have been excavated by a lunar-sized cratering event.

6. Discussion

By applying the energy estimates made in Section 5 and the present calculations for the cratering flow associated with the formation of the Imbrium basin, it is possible to estimate the depth from which ejecta from such a major impact has been excavated. We presume that samples from these depths are represented in the present lunar sample collection.

We must first make the observation that in both centimetre-sized laboratory hypervelocity impact craters in rock and virtually all natural terrestrial impact craters no moderately or heavily shocked rock is found in situ in the walls or on the floor of the crater. All of this material is initially mobilized in, or ejected from, the transient crater although some of the moderately and heavily shocked material returns to the crater via the fallback ejecta (breccia) or melt. Hence by delineating the original depth in bedrock to which moderately to heavy-shock metamorphism extends, we have determined indirectly the depth of sampling. Critical assumptions also required are (a) that the mare basins were excavated by dense oblate impacting at ~15 km/s, as assumed in our calculations; (b) the radius of the initial crater in the case of Imbrium, prior to the triggering of igneous activity (melt formation) isostatic rebound and crater-wall slumping is in the range estimated by Hartmann and Wood (1971) or Short and Foreman (1972); (c) that the depth/radius ratio of the initial crater was similar to that observed by Pike (1974) for lunar craters having radii less than 5 km (i.e., \( a = 0.4 \)). With these assumptions we calculate from the flows calculated in problem 2 that if all the material which has experienced a peak shock
pressure of ~100 kbar or higher during the initial stages of casting eventually becomes ejecta, such materials extend to a depth of ~10–11 meteorite radii (Ahrens and O'Keefe, 1977). For a 15-km/s iron meteorite having energies which are estimated for the Imbrium crater to range from (1.0–9.4) x 10^{12} erg, a radius of 144.5–29.8 km for the meteoroid is inferred. Thus rocks exposed to ~100 kbar extend to depths of 148–328 km, which would represent the maximum depth of sampling expected to be represented within the Apollo collection. These depth ranges are considerably greater than those inferred by Head et al. (1975) who estimated a maximum sampling depth of only 8–27 km largely as a result of assuming that the large mare-basin craters had a considerably lower initial depth/radius ratio. Note that Head et al. (1975) estimated that the depth of maximum sampling are less than the diameter of a hypothetical meteoroid of either low- or hyper-velocity impact.

Using independent estimates of the ejecta volumes from Imbrium obtained from the work of Pike (1974) (7.8–22 x 10^6 km^3) and the present calculation of 28 meteorite masses of shock-induced melt or partial melt rock melt, implies that the percentage of melt in the casting ejecta ranges from 40 to 100%, depending upon the radius assumed for Imbrium and the transient shape (eqs 12 or 13) assumed. The latter amount of impact melt is in the range observed in studies of the lunar impact breccias.

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References


