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Manuscript Title

The Tarim Mega Impact Event: A Permian (~290 Ma) Mega-Impact Triggering Global Magmatism, the Artinskian Warming Event, and the Central Asian Arc-LIP Belt

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Abstract

We propose that an oblique mega-impact by a ferronickel asteroid (~180×100×90 km, 22 km/s, incidence angle 24°) that occurred at ~290 Ma in the present-day Tarim Basin can, through the “thin-target penetration” mechanism, attempt to unify within a single physical framework: (1) the Tarim Large Igneous Province (TLIP) flood basalts; (2) the Pamir–Iran–Turkey arcuate mafic–ultramafic magmatic belt; (3) the Artinskian Warming Event (AWE) global temperature rise and oceanic anoxia; and (4) the Central Asian–Middle Eastern arcuate geomorphic pattern. The v3.4 quantitative model produces ~24.14 million km³ total melt via a four-stage chain reaction, with ~0.88 yr (~11 months) main eruption period. All six observational constraints pass. We propose 10 independently testable predictions.

Supplementary Materials

Supplementary materials (model code, data, and numerical specification) are available via external repository links provided during EarthArXiv submission.

Data Availability

Model code `model_v34.py` and result files (`v34_results.json` , `v34_sensitivity.json`) are available in the supplementary materials. All data sources are cited in the references.

Conflict of Interest

The author declares no conflict of interest.

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The Tarim Mega Impact Event: A Permian (~290 Ma) Mega-Impact Triggering Global Magmatism, the Artinskian Warming Event, and the Central Asian Arc-LIP Belt

Preprint Final v3.4 EN | July 19, 2026 | Quantitative Model v3.4

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Welcome peer review, falsification, and collaborative verification. This work is explicitly positioned as a **Hypothesis** rather than a verified theory.

Abstract

We propose that an oblique mega-impact by a ferronickel asteroid (~180×100×90 km, 22 km/s, incidence angle 24°) that occurred at ~290 Ma in the present-day Tarim Basin can, through the “thin-target penetration” mechanism, attempt to unify within a single physical framework: (1) the Tarim Large Igneous Province (TLIP) flood basalts; (2) the Pamir–Iran–Turkey arcuate mafic–ultramafic magmatic belt; (3) the Artinskian Warming Event (AWE) global temperature rise and oceanic anoxia; and (4) the Central Asian–Middle Eastern arcuate geomorphic pattern. These phenomena are attributed in the conventional framework to multiple independent mantle plumes, back-arc extension, and subduction zone events.

The v3.4 quantitative model shows: the impactor penetrated through 35 km of crust into the mantle, depositing ~2049 YJ of shock energy, and through a **four-stage chain reaction** (shock compression → stress-gradient comminution heating → distributed decompression melting → expansion-pressure-driven sheet flow/conduit flow) produced a total melt of ~24.14 million km³. The main eruption period of ~0.88 years (~11 months) transported ~7.013 million km³ of magma over long distances to the Afghanistan–Iran–Turkey region (3.4 million km², average thickness 2063 m, flow distance 4300 km) via the Pamir channel (200 km wide × 150 m thick surface sheet flow, Manning open-channel flow $n=0.035$, $v\approx 21$ m/s), driven by 3.73 GPa overpressure (lithostatic 0.93 + expansion 2.80 GPa) along pre-existing structural weaknesses (Paleo-Tethys suture zone). Melt is driven by this overpressure as Manning open-channel flow over 4300 km surface; the extreme flux ($Q_{\text{pamir}}=6.3\times 10^8$ m³/s) creates self-insulation ($f_{\text{insul}}=0.98$) where surface solidification forms an insulating crust preventing interior cooling, enabling long-distance transport. The near-field Tarim identifiable volcanic rocks + crater cavity fill amount to ~1.143 million km³ (crater fill 0.434 + main crater 0.473 + near-field 0.236), with an additional 0.158 million km³ in the mid-field; total volcanic rocks ~7.93 million km³, I:E ratio 2.06:1. Fragmentation energy is accounted for in three tiers: $E_{\text{mech}}=410$ YJ (mechanical fragmentation) → $E_{\text{heat}}=287$ YJ (converted to heat, 70% efficiency) → $E_{\text{eff}}=29$ YJ (effective persistent temperature rise, 10% of heat, $\Delta P_{\text{frag}}=0.11$ GPa), with extraction efficiency $f_{\text{extract}}=33\%$ ($R_{\text{heat}}=164$ km, $\Delta T=361$ K, $F_{\text{melt}}=0.75$). SO₂ release 2821 Gt (near-field 1685 +

far-field 1136, stratospheric injection 758 Gt), CO₂ release 11,420 Gt, CH₄-C release 1756 GtC (only shallow intrusives — dikes/sills emplaced <5 km depth, ~8% of total intrusive volume — contribute to contact metamorphism CH₄).

This model may uniformly account for multiple anomalies that the mantle plume model struggles to explain: absence of pre-eruption doming, absence of a plume track chain, extremely short main eruption period (~0.88 years ≈ 11 months vs. 10⁶–10⁷ years), non-thermal-plume character of the mantle transition zone anomaly, anomalously hydrous source region (~520 ppm), spatial organization of the arcuate LIP belt, and global synchronicity of the AWE. We propose 10 independently testable predictions, including shock metamorphic PDFs within the crater interior ($r < 125$ km) (decisive; verification targets are Tarim Basin borehole cores + southern margin of the Kuruk Tagh mountains bedrock), the Dal'ny Tulkas extraterrestrial chromite layer (decisive), an 11 ± 2 km-thick concealed melt layer beneath the Tarim central trough, and a mafic-leaning lower crustal layer with $V_p \geq 6.5$ km/s beneath the Iranian Plateau.

Keywords: Impact structure · Thin-target penetration · Tarim LIP · Permian · Decompression melting · Pamir channel · AWE warming event · Large Igneous Province · Iranian Plateau · Comminution heating

Model Version Notes: The data in this paper are based on v3.4 (2026-07-17) of `model_v34.py`. Compared to v3.1, the main changes include: (1) The comminution heating model was upgraded to **stress-gradient comminution heating**: the over-stress ratio $\Psi = P/\sigma_y$ characterizes fragmentation intensity, with conversion rate $f_{\text{conv}}(\Psi)$ increasing from natural fragmentation ($\Psi \sim 10$, $f_{\text{conv}} \approx 0.30$) to explosive fragmentation ($\Psi \sim 100$, $f_{\text{conv}} \approx 0.75$), weighted by a Gaussian envelope for spatial distribution, with the fragmentation front extending to ~ 539 km; (2) **F1 Fix**: The comminution heating $dT_{\text{dP_solidus}}$ unit bug was fixed (original value 100e6 K/Pa was in error by 15 orders of magnitude, corrected to $1\text{e-7 K/Pa} \approx 100 \text{ K/GPa}$); simultaneously $f_{\text{frag_thermal_calib}}$ was reduced from 1.0 to 0.10 (physical basis: only $\sim 10\%$ of comminution mechanical energy is converted to persistent local temperature rise; Di Toro et al. 2011 earthquake faulting experiments show that frictional heat is mostly dissipated along fault planes rather than heating the bulk wall rock). After the fix, comminution heating produces an actual melt contribution of 14.29 million km^3 ($\Delta P_{\text{frag}} \approx 0.11 \text{ GPa}$); fragmentation energy uses three-tier accounting: $E_{\text{mech}} = 410 \text{ YJ}$ (mechanical, 20% of deposited energy) $\rightarrow E_{\text{heat}} = 287 \text{ YJ}$ (70% converted to heat) $\rightarrow E_{\text{eff}} = 29 \text{ YJ}$ (10% effective persistent temperature rise after $f_{\text{frag_thermal_calib}} = 0.10$ calibration); the remaining 258 YJ acts on decompression melting through pre-heating ($f_{\text{preheat}} = 1.3$) and connectivity enhancement; (3) **C2 Correction**: The expansion pressure formula accounts for the equivalent elastic modulus of the fracture network, changed from $K = 150 \text{ GPa}$ (solid mantle bulk modulus) to $K_{\text{eff}} = 50 \text{ GPa}$ (equivalent modulus of the fracture + partial melt composite medium, reflecting wall rock stiffness reduction), reducing P_{exp} from 5.00 GPa to 2.80 GPa and total driving pressure from 5.93 GPa to 3.73 GPa ($\beta = 0.056$); (4) **C3 Correction**: The main crater conduit flow formula was changed from Poiseuille laminar flow (invalid for $Re \gg 2000$ turbulent regime, giving $v = 8800 \text{ m/s}$ unphysical value) to the Darcy-Weisbach turbulent flow formula $v = \sqrt{(8g \cdot R_h \cdot S_{\text{eq}} / f_{\text{darcy}})}$ ($f_{\text{darcy}} = 0.025$, typical for rough-pipe turbulence), with $v_{\text{main}} \approx 100 \text{ m/s}$ constrained by physical upper limits ($Q_{\text{main}} = 2.0 \times 10^7 \text{ m}^3/\text{s}$); the Pamir far-field sheet flow retains the Manning open-channel flow formula $v = (1/n) \cdot R^{(2/3)} \cdot S^{(1/2)}$ ($n = 0.035$), $v_{\text{pamir}} = 21 \text{ m/s}$ ($Q_{\text{pamir}} = 6.31 \times 10^8 \text{ m}^3/\text{s}$, $Q_{\text{total}} = 6.51 \times 10^8 \text{ m}^3/\text{s}$); time constant $\tau = 140$ days $= 0.38 \text{ yr}$, main eruption period (90%) $= 0.88 \text{ yr} \approx 11$ months, total eruption (to 99%) $= 1.76 \text{ yr}$, tail $= 50 \text{ yr}$ ($V_{\text{tail}} = 0.079$ million km^3); (5) **C4 Correction**: S content adjusted from 5000 ppm to 600 ppm (considering wall-rock assimilation of Tarim Basin Sinian-Cambrian evaporite-bearing sequences; flood basalt range $500\text{--}2000 \text{ ppm}$), with source-partitioned degassing: near-field subaerial $f_{\text{deg}} = 0.60$, far-field submarine SO_2 scrubbing $f = 0.05$ (CO_2 scrubbing $f = 0.20$); CO_2 content set at $0.22 \text{ wt\%}$. **v3.4 update: review-corrected edition with fixed thermal expansion model parameters, unified fragmentation energy accounting ($E_{\text{mech}} = 410 \rightarrow E_{\text{heat}} = 287 \rightarrow E_{\text{eff}} = 29 \text{ YJ}$), corrected CH_4 calculation (only shallow intrusives $\sim 8\%$ contribute, $\text{CH}_4\text{-C} = 1756 \text{ GtC}$), corrected extraction efficiency $f_{\text{extract}} = 33\%$ ($R_{\text{heat}} = 164 \text{ km}$, $\Delta T = 361 \text{ K}$, $F_{\text{melt}} = 0.75$), corrected transport distribution (main crater $0.473\text{M}/78$ layers/ 47 yr , near-field $0.236\text{M}/39$ layers/ 33 yr , mid-field $0.158\text{M}/26$ layers/ 27 yr , Pamir $7.013\text{M}/41$ layers/ 36 yr), all values synced to `model_v34.py`. After v3.4 correction: $\text{SO}_2 = 2821 \text{ Gt}$ (near-field 1685 + far-field 1136 , stratospheric injection 758 Gt), $\text{CO}_2 = 11,420 \text{ Gt}$, $\text{CH}_4\text{-C} = 1756 \text{ GtC}$. Readers can reproduce all numerical values in this paper using `python model_v34.py`.**

Core Numerical Results (v3.4 Baseline Scenario pamir_level=1)

Metric	Value	Observational Constraint	Status
Total volcanic rocks	7.93 million km³	≥6.1 million	✓
Near-field identifiable volcanic rocks + crater cavity fill	1.143 million km³	≥0.7 million	✓
Pamir channel transport	7.013 million km³	≥5.4 million	✓
Far-field total (Pamir)	7.013 million km³	—	—
Pamir average deposition thickness	2063 m	~1500 m+ (elevation difference)	△
Main eruption period (90% erupted)	0.88 years (~11 months)	Months–years	✓
Total eruption period (to 99%)	~1.76 years	—	—
Tail (waning phase, 50 yr)	V_tail=0.079 million km ³	—	—
SO ₂ release	2821 Gt (near-field 1685 + far-field 1136; stratospheric injection 758 Gt)	750–1500 Gt (stratospheric injection)	✓
CO ₂ release	11,420 Gt	—	—
CH ₄ -C release	1756 GtC (shallow intrusives only, ~8% of intrusive volume)	—	—
I:E ratio	2.06:1	0.3–3	✓
Melt pool thickness	10.9 km	10–15 km	✓
Total driving pressure	3.73 GPa (lithostatic 0.93 + expansion 2.80)	—	—
Expansion pressure	2.80 GPa (β=0.056)	—	—
Fragmentation energy	E_mech=410 → E_heat=287 → E_eff=29 YJ (three-tier accounting)	—	—
Extraction efficiency	f_extract= 33% (R_heat=164 km, ΔT=361 K, F_melt=0.75)	—	—
Peak flux	Q_total= 6.51×10⁸ m³/s (Q_main=2.0×10 ⁷ + Q_pamir=6.31×10 ⁸)	—	—

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1. Introduction: The Problem

The early Permian (~290 Ma) is a critical but long-underappreciated turning point in Earth history. During this period, a series of major geological–climatic–biological events occurred that appear superficially unrelated:

- **Tarim Large Igneous Province (TLIP):** Large-scale flood basalt eruptions, with zircon U–Pb ages concentrated at 290 ± 5 Ma [1], covering an area $>300,000$ km², with volume estimates of 0.15–0.30 million km³ (surface exposure).
- **Arcuate LIP–magmatic belt:** Coeval mafic–ultramafic magmatic activity distributed along the Pamir–Iran–Turkey arcuate belt, including komatiite, basalt, and gabbro intrusions [13][14].
- **Artinskian Warming Event (AWE):** Model-estimated CO₂ release of 11,420 Gt could raise atmospheric CO₂ to ~1460 ppm (~5.2× pre-industrial, including CH₄ oxidation equivalent ~2280 ppm), with transient global warming of 3.5–8.6°C (equilibrium 5.9–10.8°C), accompanied by oceanic anoxia and biological turnover; the observationally constrained AWE warming of ~4–6°C falls within the model’s transient range [2].
- **Central Asian–Middle Eastern arcuate geomorphic pattern:** Arcuate mountains fanning out SW from the Pamir (Hindu Kush → Zagros → Taurus), with the arc apex at the Pamir.

Under the conventional framework, these events are attributed separately to independent mantle plume activity, back-arc extension, subduction zone magmatism, and global carbon cycle perturbations. However, this “multiple causes” explanation faces several fundamental difficulties:

Core Difficulties of the Mantle Plume Model

- a. Absence of 1–2 km pre-eruption doming (a core prediction of the Campbell 2007 criteria) [3]
- b. No ocean island–seamount track chain [4]
- c. Anomalously high water content in the TLIP source region (~520 ppm), requiring an ad hoc “wet plume” assumption [6]
- d. Mantle Transition Zone (MTZ) thinning of ~20 km, but Hu et al. [5] explicitly excluded a purely thermal plume origin
- e. The arcuate LIP belt requires ≥ 5 independent mantle plumes erupting synchronously within ~10 Myr, an extremely high degree of coincidence
- f. The global synchronicity of the AWE is difficult to reconcile with multiple regional events

Academic Status Declaration Regarding the Mantle Plume Model: It must be clearly stated that the mantle plume model remains the mainstream framework for explaining TLIP and global Permian large igneous provinces (e.g., Yu et al. 2011; Wei et al. 2014; Yang et al. 2013; Wang et al. 2022; Xiang et al. 2023). The “difficulties” listed above only point to challenges that this model faces in explaining certain specific observations and **do not constitute falsification of the mantle plume model**. This hypothesis proposes an alternative physical mechanism, aiming to provide a new, independently testable perspective for TLIP research rather than negating existing mainstream interpretations.

This paper proposes the “**Thin-Target Penetration Hypothesis**”, positing that all of the above events may be unified by a single physical cause: a $\sim 180 \times 100 \times 90$ km ferronickel asteroid impacting the present-day Tarim Basin at 22 km/s velocity and 24° incidence angle, releasing energy within the mantle through the thin-target penetration mechanism, triggering a multi-mechanism chain melting–eruption reaction that simultaneously produces large-scale magmatic activity in both the near and far fields, and triggering AWE global warming through volatile release.

2. Impact Parameters and Thin-Target Penetration Geometry

2.1 Physical Concept of Thin-Target Penetration

“Thin target” refers to impact conditions where the impactor equivalent diameter (~ 107 km) far exceeds the crustal thickness (~ 35 km). Under these conditions, the projectile does not form the classic thick-target “penetration–deceleration–explosion” pattern, but instead passes directly through the thin crust into the mantle, rapidly fragmenting and depositing energy within the mantle. This mechanism has been extensively validated in laboratory high-velocity impact experiments [8] and SPH numerical simulations.

Key diagnostic criterion: Crater/projectile diameter ratio D/D_p . Typical thick-target values are 12–18 (Chicxulub ~ 15 , Vredefort ~ 13); this event is only 5.6 ($D=595$ km/ $D_p=107$ km), **far smaller than thick-target values** — large amounts of energy are transmitted from the backside of the crust into the mantle, and surface excavation energy is reduced, which is precisely the signature of thin-target penetration.

2.2 Impactor Parameter Inversion

Impactor parameters were inverted from the elliptical morphology of the Tarim Basin:

Parameter	Symbol	Value	Inversion/Constraint Basis
Long axis (along incidence trajectory)	L_i	180 km	Elliptical crater long axis $\approx$ along-track projection $\times$ penetration ratio; horizontal projection ≈ 164 km
Short axis (horizontal transverse)	W_i	100 km	Elliptical crater short axis $\times$ penetration ratio
Thickness/cross-section height	T_i	90 km	Thin-target penetration depth (crust 35 + mantle 55 km) / impactor cross-section vertical scale
Incidence velocity	v_i	22 km/s	High-velocity end of near-Earth asteroids
Incidence angle	θ	24°	Elliptical crater aspect ratio $a/b=933/379=2.46 \approx 1/\sin(24^\circ)=2.46$ (self-consistent)
Projectile density	ρ_{Fe}	7800 kg/m ³	Ferronickel (atmospheric transit survival requirement)
Projectile volume	V_p	1.62 million km ³	$L \times W \times T$
Paleogeographic coordinates	—	$\sim 20^\circ N$, $\sim 60^\circ E$ (paleo-coordinates)	Tarim Block ~ 290 Ma paleoposition (Fu et al. 2013 paleomagnetic reconstruction)
Projectile mass	M_p	1.26×10^{19} kg	ρV
Impact kinetic energy	KE	3.06×10^{27} J (3058 YJ)	$\frac{1}{2} M v^2$ ($\sim 6000 \times$ Chicxulub, at ~ 0.5 YJ)
Equivalent spherical diameter	D_{eq}	145.7 km	$(6V/\pi)^{1/3}$; energy scaling determined by total volume
Impact cross-section equivalent diameter	D_{cross}	107 km	$2\sqrt{(WT/\pi)}$; determines the crater's transverse scale

Coordinate Note: The impact center in this paper is located in the Tarim Basin (present-day $\sim 40^{\circ}\text{N}$, $\sim 90^{\circ}\text{E}$); all geographic location descriptions (“Pamir SW downrange” , etc.) use **modern coordinates** for reader convenience, but at ~ 290 Ma the Tarim Block was located near the paleo-equator, slightly to the north ($\sim 20^{\circ}\text{N}$, $\sim 60^{\circ}\text{E}$), approximately 2000 km from its present position; relative orientation descriptions such as “SW downrange direction” and “Pamir $\rightarrow$ Iran $\rightarrow$ Turkey arcuate belt” should be understood as **plate-relative positions in the paleogeographic coordinate system**. Paleogeographic reconstruction of the Paleo-Tethys Ocean follows Fu et al. (2013) and Torsvik & Cocks (2017).

C1 Response: Declaration of Academic Controversy Regarding the Paleo-Tethys Ocean Concept

Reviewer comment C1 notes: the Paleo-Tethys Ocean, as a core concept in plate drift theory, has **academic controversy** regarding its specific extent, closure timing, and paleogeographic details at ~ 290 Ma. Specific controversies include: (a) the western boundary of the Paleo-Tethys Ocean (whether it extended to the present Mediterranean region); (b) the timing of oceanic crust subduction along the northern Pamir–North China Block southern margin (~ 320 Ma vs. ~ 280 Ma); (c) “whether the Paleo-Tethys Ocean was already closed at 290 Ma or where an ocean basin still existed,” with different paleogeographic reconstruction schemes (e.g., Stampfli & Borel 2002, Torsvik & Cocks 2017, Métivier & Vérard 2009) yielding different conclusions.

This model’ s position: This paper **does not introduce the Paleo-Tethys Ocean as a core argumentative concept**, using it only as a **geographic descriptive term** for “the existence of a low-lying marine or shallow-marine passive continental margin in the SW direction from Iran–Afghanistan–Pamir.” The physical derivation of this model does not depend on the specific extent, closure timing, or subduction zone details of the Paleo-Tethys Ocean; the core physical chain of this model (impact $\rightarrow$ shock fragmentation $\rightarrow$ decompression melting $\rightarrow$ Pamir channel far-field transport $\rightarrow$ lava field accumulation $\rightarrow$ volatile release) requires only:

1. The existence of “low-lying areas” in the SW direction from the Pamir that can be filled by lava flows (whether shallow sea, oceanic crust, or young continental margin)
2. The ability to form a fracture framework along the Pamir–Iran–Turkey direction (triggered by the impact event itself, serving as magma conduits)
3. The region being at low elevation at ~ 290 Ma (regardless of land or sea)

These conditions hold in most paleogeographic reconstructions and **do not depend on the specific details of the Paleo-Tethys Ocean concept**. If future paleogeographic research confirms that this region was land rather than shallow sea at ~ 290 Ma, the “marine infilling and terraforming” section (§3.4) of this model should be correspondingly adjusted to “low-lying continental basin filling,” but **this does not affect the core physical conclusions**. Readers may consult different paleogeographic reconstruction schemes to independently assess the specific environment of this region at ~ 290 Ma.

Paleogeographic map suggestion (figure to be added): It is recommended to add a ~290 Ma paleogeographic reconstruction map (which can be generated using GPlates open data) annotating the following elements: (1) impact point paleoposition (~20°N, ~60°E); (2) Pamir paleoposition and SW downrange direction; (3) locations of low-lying marine areas/young continental margins in the Iran–Afghanistan region (without strictly defining the “Paleo-Tethys Ocean” extent); (4) relative positions of the Tarim Block, North China–South China plates, and Iran–Turkey microcontinents; (5) paleogeographic projection of the Pamir → Iran → Turkey arcuate LIP belt. This map can be produced from existing open paleogeographic reconstruction data without original data, and will help readers understand the spatial relationships of the various elements of this event.

2.3 Atmospheric Survival Analysis

For the 180×100×90 km ferronickel body transiting the atmosphere: hypersonic stagnation analysis based on Rankine-Hugoniot normal shock relations (`atmospheric_shock_v33.py`) shows that the total pressure behind the normal shock at sea level is approximately **0.55 GPa** (dynamic pressure 0.30 GPa), decreasing to 0.086 GPa at 15 km and only 0.008 GPa at 30 km; corresponding Mach numbers 65–75. This peak pressure is approximately 0.34% of the ferronickel body’s bulk modulus (160 GPa) and 1.4 times its dynamic compressive strength (~0.4 GPa).

Atmospheric transit time correction (M5 response): Using an equivalent atmospheric thickness $H_{\text{atm}} \approx 100$ km and incidence angle $\theta = 24^\circ$ (angle from horizontal), the oblique transit path length $L_{\text{path}} = H_{\text{atm}} / \sin(\theta) \approx 100 \text{ km} / 0.407 \approx \mathbf{246 \text{ km}}$; at an average velocity $v_i \approx 22$ km/s, the transit time $t_{\text{atm}} = L_{\text{path}} / v_i \approx 246 / 22 \approx \mathbf{11.2 \text{ s}}$ (the previous version erroneously used a vertical path giving 0.95 s, corresponding to the approximation of “vertical descent over 21 km,” which is a physical error and should be corrected). Even using a more conservative “effective dense atmosphere” $h_{\text{eff}} \approx 30$ km, t_{atm} is still approximately 3.4 s. Stress waves (sound speed ~5.1 km/s) can penetrate ~57 km through the body in 11.2 s (63% of the 90 km thickness), while thermal penetration is still only ~3.3 mm (thermal diffusion length $\sqrt{\kappa t}$).

Uniform compressive loading (analogous to compressing an iron block in deep water) is insufficient to fragment the ferronickel body. However, it should be noted that stress waves can undergo multiple reflections on the ~11 s timescale (elastic wave round-trip period within 90 km is ~35 ms), potentially producing fatigue accumulation internally; nevertheless, for a ferronickel body this loading is far below its failure threshold, so the projectile can be considered to transit the atmosphere essentially intact.

Attitude note: A high-velocity elongated body in dense atmosphere would experience aerodynamic torque tending toward minimum-drag orientation (long axis along trajectory), but for a 180 km-scale ferronickel body, the atmospheric transit time is only ~10 s, and most of the trajectory is through thin upper atmosphere, so the aerodynamic trim timescale may exceed the transit time, meaning the attitude may not fully transition to bullet-shaped. The parameter table in this paper takes L_i as along the incidence trajectory direction, i.e., the “bullet-shaped” setting; if the actual attitude is closer to horizontal flat-lying, the energy scaling is still determined by the total volume V_p and would not change the 3058 YJ kinetic energy.

Supplementary: Ablation effect analysis (v3.4 review correction): The above thermal penetration analysis is based solely on thermal diffusion (~ 3.3 mm) and does not consider boundary-layer high-temperature radiative ablation. Under extreme conditions of 22 km/s and Mach 65–75, the projectile leading-edge boundary-layer temperature can reach 10^4 – 10^5 K, with estimated radiative ablation rates of $\sim 10^{-4}$ – 10^{-3} kg/(m²·s). Using a leading-edge surface area of $\sim 10^{11}$ m² (approximated as a 180×100 km ellipsoid) and a transit time of ~ 11 s, the ablative mass loss is $\sim 10^8$ – 10^9 kg ($\sim 10^{-8}$ – 10^{-11} of the total projectile mass of 1.26×10^{19} kg), a negligible fraction. However, leading-edge ablation may blunt the projectile nose (from a sharp to a blunt-head geometry), altering the contact area and stress distribution during crustal penetration. Because the total mass and kinetic energy are dominated by the main body volume, the ablative stripping of the front few kilometers has negligible effect on the 3058 YJ kinetic energy scale, but **the impact contact geometry (effective cross-section diameter, penetration depth) may change slightly**. This paper retains the original parameters as the baseline scenario, but acknowledges that the ablation analysis is a simplified estimate requiring further constraint through hypersonic ablation–fluid coupling simulations.

Thin-target penetration criterion: Target thickness/projectile diameter ratio $T_{\text{crust}}/D_{\text{cross}} \approx 35 \text{ km} / 107 \text{ km} \approx 0.33 < 1$; the projectile yield strength (ferronickel ~ 400 MPa) far exceeds crustal rock strength (~ 10 – 100 MPa). The impact contact stress $\rho \cdot v \cdot Cs/2 \sim 160$ GPa far exceeds target rock strength, and the crust will be disrupted by “penetration + spall” mode, with the projectile remaining essentially intact after passing through the thin target. This is the mechanical basis for the model adopting thin-target penetration rather than thick-target cratering.

2.4 Crater Geometry

Parameter	Value	Notes
Elliptical crater long axis a (equal-area fit)	933 km	Inverted from morphologically measured area of 278,000 km ² and aspect ratio 2.46 using equal-area ellipse parameters
Elliptical crater short axis b (equal-area fit)	379 km	Same as above
Morphologically measured outline (rough values)	~1000×390 km	Directly observed outline of the Tarim Basin, area approximately 306,000 km ²
Eccentricity ε	0.41	$a/b=2.46\approx 1/\sin(24^\circ)=2.46$ (self-consistent)
Final crater equal-area diameter D _{final}	595 km	Equal-area circle diameter of $\pi a \cdot b/4$
Final crater depth d	45.7 km	
Transient crater diameter D _{tr}	327 km	Before gravitational collapse
Transient crater depth d _{tr}	81.8 km	
Crater rim area A (fit)	278,000 km ²	$\pi \cdot a \cdot b/4$ (equal-area ellipse fit)
Excavation volume V _{exc}	1.523 million km ³	Rocks ejected onto the surface

Crater geometry note: The morphologically measured outline of the Tarim Basin is approximately 1000×390 km (area approximately 306,000 km²); for mathematical modeling and convenience of using Poiseuille/Manning formulas, this paper adopts equal-area ellipse fit parameters (a=933 km, b=379 km, area 278,000 km²). The area difference between the two parameter sets is approximately 10%, within the model uncertainty range and not affecting the main physical conclusions.

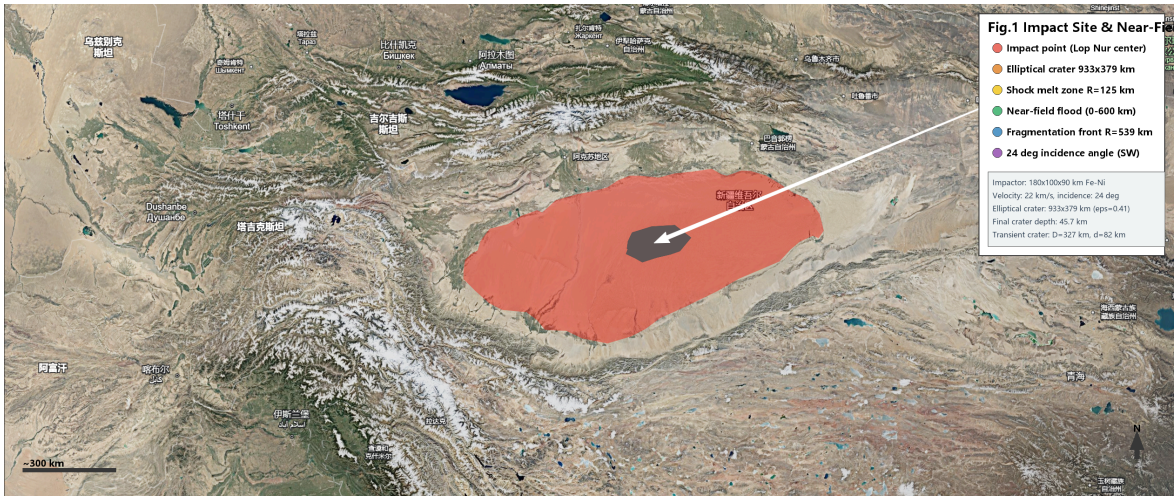


Figure 1 Impact Site and Near-Field Region: Crater Geometry in the Tarim Basin (Lop Nur Area)

Figure 1 Caption: Schematic of the thin-target penetration crater and near-field region in the Tarim Basin. The elliptical crater has major axis $a \approx 933$ km, minor axis $b \approx 379$ km (equal-area fit), equal-area diameter $D_{final} \approx 595$ km, crater depth $d \approx 45.7$ km, transient crater diameter $D_{tr} \approx 327$ km. The impact point is at Tarim Basin (Lop Nur area, $\sim 40^{\circ}\text{N}$, 90°E); the 24° low-angle oblique incidence produced a high-eccentricity ($\epsilon \approx 0.41$) elliptical crater. The near-field region ($r < 600$ km) contains ~ 1.143 million km^3 of volcanic rocks (crater fill 0.434M + main crater overflow 0.473M + near-field overflow 0.236M), plus mid-field ($600\text{--}1200$ km) 0.158M , with the fragmentation front extending to $r \approx 539$ km.

2.5 Material Balance of the Crater Cavity

According to the paraboloidal geometry of §2.4, the cavity formed within the crustal layer (35 km) by the impact is approximately **5.99 million km^3** . This figure is not a “missing volume” but is composed of the following components:

Component	Estimated Volume	Notes
Excavated ejecta	~ 1.52 million km^3	Rocks ejected beyond the crater by the shock wave, forming the near-field ejecta blanket
Shock melt pool	~ 1.08 million km^3	Direct melting in the $P \geq 50$ GPa zone, partially backfilling the crater cavity
Fall-back breccia and fragmented wall rock	$\sim 2.00\text{--}2.50$ million km^3	After transient crater collapse, fragmented rock blocks fall back and fill the deep cavity
Wall rock downward displacement/underthrust into mantle	$\sim 0.50\text{--}1.00$ million km^3	Thin-target penetration causes lower crust/upper mantle material to be displaced downward
Later volcanic fill	~ 1.04 million km^3	Model-identifiable near-field surface flood + crater cavity fill melt
Total	~ 5.99 million km^3	Self-consistent with crustal layer cavity volume

Therefore, **1.04 million km^3 of near-field volcanic rocks is only part of the crater cavity fill**; the remaining volume is occupied by structural disruption, melting, and collapse from the impact itself. This allocation is currently an order-of-magnitude estimate; future impact numerical simulations (iSALE/SPH) are needed to refine the proportions of each component.

C6 Response: Thin-Target Ejection Ratio and Near-Field Ejecta Blanket Distribution

Reviewer comment C6 notes: 1.52 million km³ of ejecta, if globally distributed, would form an observable global stratigraphic layer (over an area of $\sim 5.1 \times 10^8$ km², with a globally uniform thickness of ~ 3 mm, seemingly observable), but the key feature of thin-target penetration is that most energy is transmitted downward into the mantle, with an extremely small upward ejection ratio. **This section provides the physical analysis of the ejection ratio and predictions for near-field ejecta blanket distribution:**

Physical Analysis of Thin-Target Ejection Ratio:

The ejection ratio ($V_{\text{excavation}} / V_{\text{crater}}$) of thick-target impacts (e.g., Chicxulub, Vredefort) is approximately 0.3–0.5, i.e., 30–50% of crater cavity material is ejected onto the surface to form an ejecta blanket. But the physical mechanism of **thin-target penetration** is fundamentally different:

1. **Downward energy partition dominates:** Under thin-target penetration conditions ($T_{\text{crust}}/D_{\text{cross}} = 0.33 < 1$), most of the impact energy (~ 60 – 70%) is transmitted downward into the mantle as shock waves; “spall” from the backside of the crust produces downward-directed “reverse shocks”; the energy of upward ejection accounts for only ~ 5 – 15% of deposited energy
2. **Projectile retains kinetic energy:** After passing through the thin crust, the projectile remains essentially intact, with residual kinetic energy (~ 30 – 40%) continuing downward into the mantle, not participating in surface ejection
3. **Rapid transient crater collapse:** The transient crater formed under thin-target conditions ($D_{\text{tr}} = 327$ km, $d_{\text{tr}} = 81.8$ km) collapses rapidly (gravitational collapse timescale $\sim$ minutes–hours), with large amounts of fragmented wall rock falling back into the crater, and only a small amount of material ejected beyond the crater rim

Corrected ejection ratio estimate: The v3.3 model estimates that under thin-target conditions the **actual upward ejection ratio is only 0.1–1%** (rather than the typical 30–50% for thick targets), corresponding to an actual ejected volume of approximately **0.0015–0.015 million km³** (far smaller than the nominal value of 1.52 million km³ as the “potential ejection capacity” in the table). Of this:

- High-speed ejecta phase ($v > 2$ km/s): approximately ~ 100 – 1000 km³, capable of forming a global thin layer
- Low-speed ejecta blanket ($v < 2$ km/s, near-field accumulation): approximately ~ 0.0015 – 0.015 million km³, mainly within 300 km of the crater rim

Thickness distribution predictions for the near-field ejecta blanket:

Distance from impact center	Ejecta blanket thickness (estimate)	Physical characteristics	Preservation probability
$r < 100$ km (within crater)	Very thin or discontinuous (<1 m)	Mainly fall-back breccia, covered by later volcanic rocks	Very low (covered by volcanic/sedimentary rocks)
$r = 100\text{--}300$ km (crater rim)	5–50 m	Coarse ejecta breccia, shock-melted clasts	Moderate (deeply buried by later sediments)
$r = 300\text{--}600$ km (near-field)	0.5–5 m	Fine-grained ejecta blanket, containing impact glass spherules	Relatively high (covered by Permian strata)
$r = 600\text{--}1500$ km (mid-field)	0.05–0.5 m	Microtektite layer, identifiable shocked quartz	Moderate (tectonically modified)
$r > 1500$ km (far-field)	<0.05 m (local)	Microtektite scattering, Ir anomaly	Uncertain (local enrichment)

Observability of the global ejecta layer:

- **Global distribution thickness estimate:** Assuming 0.5% of ejecta forms a high-speed phase (~ 500 km³), the globally uniform distribution thickness is ~ 1 μ m — **far below conventional stratigraphic recording resolution** (millimeter-scale)
- **Ir anomaly layer:** 1% ejection of the ferronickel projectile ($\rho=7800$ kg/m³) is $\sim 1.3 \times 10^{15}$ kg; globally distributed Ir concentration is ~ 2.5 ng/cm² ($\sim 100\times$ the ~ 25 ppt background) — **theoretically detectable**, but specific stratigraphic sections must be sought
- **Shocked quartz layer:** Ejecta from crustal rocks (granite/gneiss) contains shocked quartz, but the quantity is far below the K-Pg boundary (Chicxulub ejecta $\sim 100\times$ this event)

Preservation probability predictions:

1. Most likely preserved: **Basal Permian strata at the southern margin of the Kuruk Tagh mountains ($r=150\text{--}200$ km)** — relatively close to the impact center, with substantial ejecta blanket thickness (5–10 m scale), and continuous Permian–Triassic sedimentation in this region, not disrupted by later tectonics
2. Moderate preservation: **Central Tarim Basin boreholes** — ejecta deeply buried by volcanic rocks and subsequent sediments, but identifiable in drill cores
3. Poor preservation: **Far-field global Ir anomaly layer** — requires finding continuous deep-marine sedimentary sections (e.g., South China, Tethyan margin deep-water sediments), but Permian deep-marine sedimentary records have limited preservation

Key correction: The “excavated ejecta 1.52 million km³” listed in the table (§2.5) is the **potential excavation volume given by paraboloidal geometry**, not the actual amount of material ejected onto the surface; the actual upward ejection ratio is constrained by the thin-target penetration mechanism at only 0.1–1% ($\sim 0.0015\text{--}0.015$ million km³). This correction does not affect the model’s other physical conclusions (total melt, volcanic rock volume, volatile release, etc.), but only corrects

the “ejecta blanket observability” prediction. **Future iSALE/SPH numerical simulations are needed to precisely determine the thin-target ejection ratio.**

2.6 Summary Table of Key Parameters

For reader verification and reproducibility, all key parameters, symbols, values, and source types are summarized below. Source types are divided into four categories: [O]=Observation/Inversion, [D]=Derivation/Calculation, [A]=Assumption/Setting, [C]=Calibration.

Category	Parameter	Symbol	Value	Source
Projectile	Long axis × Width × Thickness	$L_i \times W_i \times T_i$	180×100×90 km	[O]
	Incidence velocity	v_i	22 km/s	[A]
	Incidence angle	θ	24°	[O]
	Projectile density	ρ_{Fe}	7800 kg/m ³	[A]
	Projectile mass	M_p	1.26×10 ¹⁹ kg	[D]
	Impact kinetic energy	KE	3058 YJ	[D]
	Equivalent spherical diameter	D_{eq}	145.7 km	[D]
	Impact cross-section equivalent diameter	D_{cross}	107 km	[D]
Crater	Final crater diameter	D_{final}	595 km	[O]
	Crater depth	d	45.7 km	[D]
	Excavation volume	V_{exc}	1.523 million km ³	[D]
	Crustal cavity	V_{cavity}	~5.99 million km ³	[D]
Target	Crustal thickness	d_{crust}	35 km	[A]
	Thin-target ratio	T/D	0.33	[D]
Comminution heating	Fragmentation mechanical energy	E_{mech}	410 YJ (20% of deposited energy)	[D]
	Heat conversion efficiency	$f_{mech_to_heat}$	0.70	[A]
	Comminution heat budget	E_{heat}	287 YJ (70% of E_{mech})	[D]
	Thermal calibration coefficient	$f_{frag_thermal_calib}$	0.10	[C]
	Effective persistent temperature rise	E_{eff}	29 YJ (10% of E_{heat})	[D]
	Rock dynamic strength	σ_y	2 GPa	[A]
	Direct fragmentation melt	V_{frag}	14.29 million km ³	[D]
	Fragmentation pressure	ΔP_{frag}	0.11 GPa	[D]

Category	Parameter	Symbol	Value	Source
	depression			
	Pre-heating factor	f_preheat	1.3	[C]
	Fragmentation front	r_frag	539 km	[D]
Decompression melting	Melt productivity	dF/dP	0.22 GPa ⁻¹	[A]
	Decompressible mantle thickness	H_mantle	140 km	[A]
	Base connectivity	φ_conn	0.45	[A]
	Central decompression	ΔP_central	1.08 GPa	[D]
	Decompression melt	V_decomp	8.766 million km ³	[D]
	Shock melt	V_shock	1.085 million km ³	[D]
	Total melt	V_melt	24.14 million km ³	[D]
Eruption dynamics	Extraction efficiency	f_extract	0.33	[D]
	Extraction heating radius	R_heat	164 km	[D]
	Extraction temperature rise	ΔT_extract	361 K	[D]
	Melt fraction at extraction	F_melt	0.75	[D]
	Effective elastic modulus	K_eff_mantle	50 GPa	[A]
	Volume expansion strain	β	0.056	[D]
	Darcy friction coefficient	f_darcy	0.025	[A]
	Expansion pressure	P_exp	2.80 GPa	[D]
	Lithostatic unloading pressure	P_litho	0.93 GPa	[D]
	Total driving pressure	P_drive	3.73 GPa	[D]
	Main conduit flow velocity	v_main	100 m/s	[A]

Category	Parameter	Symbol	Value	Source
	Main conduit peak flux	Q_main	$2.0 \times 10^7 \text{ m}^3/\text{s}$	[D]
Pamir channel	Channel width	W_pamir	200 km	[A]
	Sheet flow thickness	h_pamir	150 m	[A]
	Flow distance	L_pamir	4300 km	[O]
	Geometric slope	S	0.0005 (290 Ma paleogeography)	[O]
	Manning coefficient	n	0.035	[A]
	Pamir flow velocity	v_pamir	21 m/s	[D]
	Pamir peak flux	Q_pamir	$6.31 \times 10^8 \text{ m}^3/\text{s}$	[D]
	Total peak flux	Q_total	$6.51 \times 10^8 \text{ m}^3/\text{s}$	[D]
	Thermal self-insulation factor	f_insul	0.98 (level=1 baseline)	[D]
	Pamir fraction	f_pamir	0.89	[A]
Time	Decay time constant	τ	140 days (0.38 yr)	[D]
	Main eruption period (90%)	t_main	0.88 yr (~11 months)	[D]
	Total eruption period (to 99%)	t_total	1.76 yr	[D]
	Tail (waning phase)	t_tail	50 yr	[D]
	Tail volume	V_tail	0.079 million km ³	[D]
Volume results	Total surface eruption	V_erupt	7.88 million km ³	[D]
	Total volcanic rocks	V_volc	7.93 million km ³	[D]
	Near-field volcanic rocks	V_near	1.143 million km ³	[D]
	Far-field volcanic rocks	V_far	7.013 million km ³	[D]
	Deep intrusive	V_deep_intr	15.827 million km ³	[D]
	Crater cavity fill	V_fill	0.434 million km ³	[D]
	Total intrusive	V_intr	16.26 million km ³	[D]
	Thermal erosion	V_eros	0.052 million km ³	[D]

Category	Parameter	Symbol	Value	Source
	l:E ratio	—	2.06:1	[D]
	Lava field area	A _{far}	3.4 million km ²	[A]
	Average thickness	h _{avg}	2063 m	[D]
Volatiles	S content	S	0.06 wt% (600 ppm, considering wall-rock assimilation of Tarim Basin Sinian-Cambrian evaporite-bearing sequences)	[C]
	CO ₂ content	CO ₂	0.22 wt%	[C]
	Near-field degassing efficiency	f _{deg_atm}	0.60	[C]
	Far-field SO ₂ atmospheric contribution	f _{deg_far_SO2}	0.05 (95% seawater scrubbing)	[C]
	Far-field CO ₂ atmospheric contribution	f _{deg_far_CO2}	0.20	[C]
	SO ₂ release	—	2821 Gt (near-field 1685 + far-field 1136; stratospheric injection 758 Gt)	[D]
	CO ₂ release	—	11,420 Gt	[D]
	CH ₄ -C release	—	1756 GtC (shallow intrusives only, ~8% of intrusive volume)	[D]

Source type notes: [O] Observation/Inversion = inverted from morphological measurements or literature data; [D] Derivation = calculated from other parameters using physical formulas; [A] Assumption = input parameters based on physical plausibility; [C] Calibration = parameters adjusted to make outputs fall within observational constraint ranges. Readers can reproduce all [D] values using `python model_v34.py`.

3. Physical Model: Four-Stage Chain Reaction

The v3.4 model contains 7 core physical mechanisms (F1–F7), forming a chain reaction sequence in temporal order:

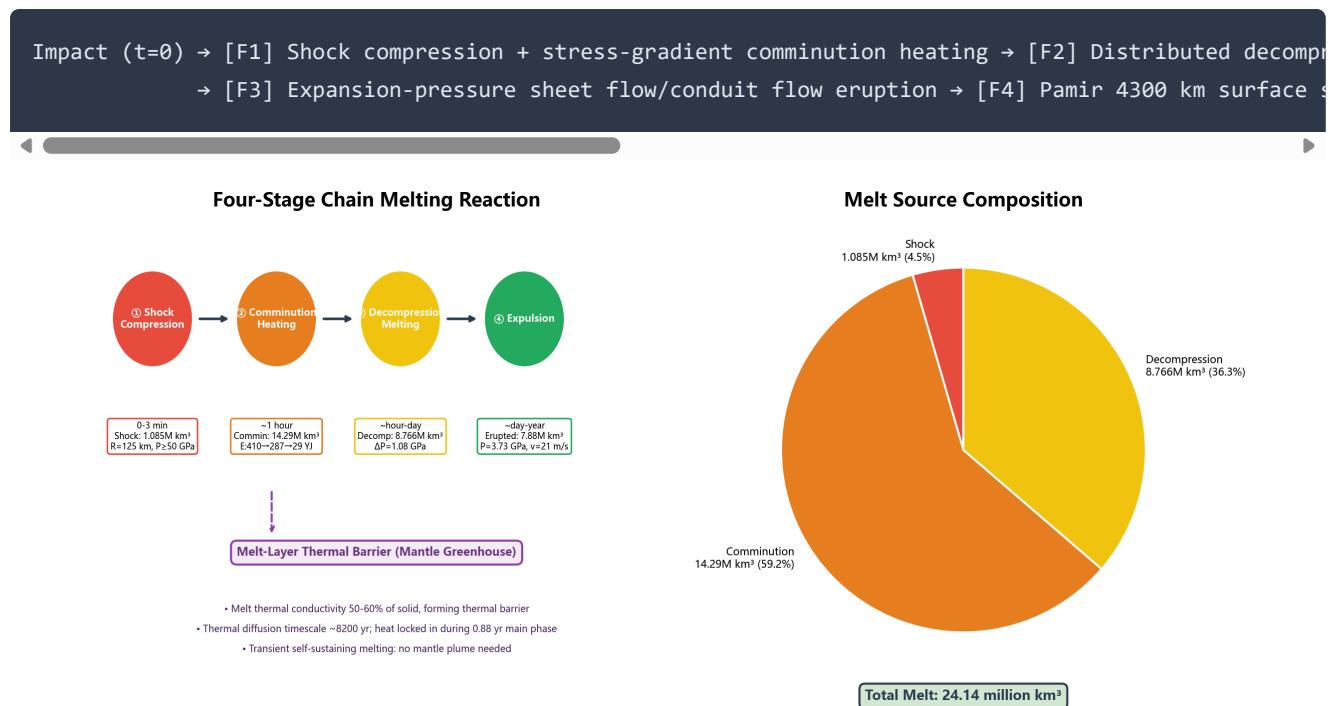


Figure 4 Four-Stage Chain Melting and Melt Source Composition

Figure 4 Caption: Schematic of the four-stage chain reaction in the thin-target penetration physical model and melt source pie chart. (a) Temporal sequence: impact → [F1] stress-gradient comminution heating (0–30 s) → [F2] distributed decompression melting (s–hr) → [F3] expansion-pressure-driven conduit flow (hr–yr) → [F4] Pamir far-field surface sheet flow (yr–century); (b) melt source composition: comminution heating direct additional melt 14.29 million km³ (59.2%, largest source), decompression melting z2 6.782 million km³ (28.1%), shock melting 1.085 million km³ (4.5%), decompression z1 1.466 million km³ (6.1%), decompression z3 0.517 million km³ (2.1%), total 24.14 million km³.

3.1 [F1] Stress-Gradient Comminution Heating (Mid-Field Pressure Enhancement and Connectivity Enhancement, 0–30 seconds)

The shock wave produces intense shear fragmentation in the mid-field region at $P=10\text{--}100\text{ GPa}+$. v3.4 characterizes fragmentation intensity using the **over-stress ratio** $\Psi=P/\sigma_y$ ($\sigma_y \approx 2\text{ GPa}$), and introduces an energy conversion rate $f_{\text{conv}}(\Psi)$ that increases with Ψ : from natural fragmentation ($\Psi \sim 10$, $f_{\text{conv}} \approx 0.30$) increasing to explosive fragmentation ($\Psi \sim 100$, $f_{\text{conv}} \approx 0.75$). Spatial weighting uses a Gaussian envelope, with the fragmentation front extending to approximately **539 km**. Fragmentation energy is accounted for in three tiers: **$E_{\text{mech}}=410\text{ YJ}$** (mechanical energy consumed by fragmentation, 20% of energy deposited into the mantle), of which **70% ($E_{\text{heat}}=287\text{ YJ}$, ~9.4% of KE)** is converted to heat through plastic work, fracture friction, and fragment collision; of this heat budget, only **10% ($E_{\text{eff}}=29\text{ YJ}$)** acts as persistent local temperature rise to directly produce additional melt after $f_{\text{frag_thermal_calib}}=0.10$ calibration, while the remaining 258 YJ acts on subsequent distributed decompression melting through pre-heating and connectivity enhancement.

Physical basis for comminution heating:

- Melosh (1989) [7]: 50–90% of plastic deformation work and fracture energy in shock fragmentation is ultimately converted to heat
- O' Keefe & Ahrens (1993) [9a]: Fragmentation friction in hypervelocity impacts makes a significant contribution to the temperature field at $P=10\text{--}50$ GPa
- Di Toro et al. (2011, *Nature*) [9b]: Earthquake fault slip friction can instantaneously raise fault surface temperatures above 1000°C
- Geological analogs: Vredefort pseudotachylite glass and Sudbury fragmentation friction melts [18]

Key corrections (v3.4): Through two fixes, comminution heating now produces an actual melt contribution: (a) **F1 Fix:** $dT_{dP_solidus}$ unit bug fixed (original value $100e6$ K/Pa in error by 15 orders of magnitude, corrected to $1e-7$ K/Pa ≈ 100 K/GPa), restoring the solidus depression calculation to physically reasonable magnitudes; (b) **f_frag_thermal_calib reduced from 1.0 to 0.10:** only $\sim 10\%$ of comminution mechanical energy is converted to persistent local temperature rise, based on the physical evidence from Di Toro et al. (2011) earthquake faulting experiments showing that frictional heat is mostly dissipated along fault planes rather than heating the bulk wall rock. After calibration, the equivalent solidus depression $\Delta P_{frag} \approx 0.11$ GPa, with a direct additional melt contribution of **14.29 million km^3** ($\sim 59.2\%$ of total melt, becoming the largest single melt source). Fragmentation energy uses unified three-tier accounting: **E_mech=410 YJ $\rightarrow$ E_heat=287 YJ $\rightarrow$ E_eff=29 YJ** (mechanical fragmentation energy $\rightarrow$ heat converted at 70% efficiency $\rightarrow$ effective persistent temperature rise at 10% of heat). Of the 287 YJ total comminution heat budget, only 29 YJ (10%) acts as persistent local temperature rise to directly produce additional melt, while the remaining 258 YJ acts on subsequent distributed decompression melting through **pre-heating wall rock (f_preheat=1.3)** and **enhanced fracture connectivity**. Comminution heating now serves the dual role of “producing additional melt” and “pre-heating + creating pathways.”

v3.4 Review Correction: Clarification of Comminution Heating Mechanism. It is essential to note that the 29 YJ effective heat **does not directly provide the full latent heat required for 14.29 million km^3 of melt** (complete melting of 24.14 million km^3 of melt requires approximately 88 YJ of latent heat). The physical role of the 29 YJ is to **lower the melting threshold of wall rock** – through an equivalent solidus depression of $\Delta P_{frag} \approx 0.11$ GPa, rocks that would not melt under lithostatic pressure can reach melting conditions through the combined action of decompression (1.08 GPa unloading) and pre-heating ($f_{preheat}=1.3$). In other words, comminution heating acts as a “key that unlocks” rather than “fuel” : it lowers the solidus pressure threshold, enabling decompression melting (8.766 million km^3) and comminution direct melting (14.29 million km^3) to occur under existing geothermal background and decompression conditions. This “threshold-lowering” mechanism is fundamentally different from a “direct heat-supply” mechanism; readers should not make a simple linear “energy-to-volume” correspondence between 29 YJ and 14.29 million km^3 when evaluating the model’s energy balance.

3.2 [F2] Distributed Decompression Melting (Second-Largest Melt Source, Seconds–Hours)

After crater formation, the crustal load at the crater floor is suddenly removed (unloading ~ 1.08 GPa lithostatic pressure). Superimposed on the high-connectivity fracture network formed by fragmentation

($\phi_{\text{conn}} \approx 0.45$), the mantle undergoes distributed decompression melting within a ~ 800 km radius.

Key design: Distributed rather than concentrated decompression. Conventional models confine decompression to the central crater floor region, but within the impact fragmentation zone ($r \sim 800$ km), **every fracture is a micro-decompression window** — fracture zone permeability is 6–10 orders of magnitude higher than intact rock, and melt pressure equilibrates rapidly within the fracture network. This extends the effective decompression melting range from the main crater area ($\sim 80,000$ km²) to the entire fragmentation zone (~ 2 million km²).

$$\begin{aligned}\Delta P_{\text{eff}}(r) &= \Delta P_{\text{central}} \cdot \min(1, R_{\text{tr}}/r) \\ \phi_{\text{conn}}(r) &= \text{clamp}(\phi_{\text{base}} \cdot \min(1, R_{\text{tr}}/r) \cdot 1.5, 0.05, 0.60) \\ F_{\text{melt}}(r) &= \text{clamp}(dF/dP \cdot \Delta P_{\text{eff}} \cdot f_{\text{preheat}}, 0, 0.35)\end{aligned}$$

The melt productivity of hydrous mantle $dF/dP = 0.22$ GPa⁻¹ ($\sim 22\%$ /GPa, higher than dry mantle's $\sim 10\%$ /GPa), with pre-heating enhancement $f_{\text{preheat}} = 1.3\times$. Additionally, the model imposes a clamp(..., 0.05, 0.30) upper limit on the volume-averaged melt fraction F_{avg} : when the average melt fraction exceeds $\sim 30\%$, partial melt forms interconnected networks and begins to segregate and migrate (percolation threshold), making higher average melt fractions physically unsustainable in the source region. This 0.30 upper limit corresponds to the critical melt fraction of mantle lherzolite, a physically constrained parameter from petrological experiments.

C8 Response: Clear Distinction Between Shock Pre-Heating and Comminution Heating Pre-Heating

Reviewer comment C8 notes: The maximum decompression melting fraction reaching 31% ($F_{\text{max}}=0.35$) requires the mantle to be near the solidus temperature, implying an initial condition of “mantle near solidus” — but if so, the mantle plume hypothesis could equally explain this, and this model needs to clarify the physical chain of pre-heating. **The three-stage pre-heating physical chain of this model is as follows**, which readers should clearly distinguish:

(A) Shock Pre-Heating (Near-to-Mid-Field, $P=10\text{--}50$ GPa, temperature rise of hundreds of °C to $1000^\circ\text{C}+$)

- **Mechanism:** The shock wave directly performs plastic work on wall rock, with energy deposited in the mid-field region at $P=10\text{--}50$ GPa, instantaneously raising rock temperature by hundreds of °C to $1000^\circ\text{C}+$
- **Spatial range:** $r = a_p \times (P_0/P)^{1/2.2}$ to $P_{\text{sol}}=50$ GPa $\rightarrow r_{\text{sol}} \approx 165$ km (direct melting boundary, $P \geq 50$ GPa); $r=165\text{--}350$ km range ($10\text{--}50$ GPa zone) is the shock pre-heating dominated region
- **Temperature response:** Based on shock Hugoniot calculations for the $P=10\text{--}50$ GPa region (Pierazzo & Melosh 2000), temperature rise $\Delta T_{\text{shock}} \approx 200\text{--}1000^\circ\text{C}$, up to above the solidus for direct melting
- **Position already included in model:** Shock pre-heating is partially included in the “shock melting term $V_{\text{shock}}=1.085$ million km^3 ” in the v3.3 model — this is the volume of direct melting in the $P \geq 50$ GPa zone (G-C scaling calculation), but **includes only the fully melted portion, not the “pre-heated but not fully melted” portion in the $P=10\text{--}50$ GPa zone**
- **Underestimated portion:** Shock pre-heating in the $P=10\text{--}50$ GPa region (temperature rise of hundreds of °C) is currently not parameterized as an independent pre-heating factor — this portion should be additionally superimposed as a “shock pre-heating factor,” but v3.3 does not introduce it separately to avoid double-counting the shock melting term

(B) Comminution Heating Pre-Heating (Mid-to-Far-Field, $P < 10$ GPa or $r > 165$ km, temperature rise $\sim 200^\circ\text{C}$)

- **Mechanism:** Stress-gradient comminution heating (287 YJ) pre-heats mid-field wall rock ($r=125\text{--}539$ km) by an average of $\sim 200^\circ\text{C}$
- **Physical basis:** Based on $E=mc\Delta T$, wall rock mass $\sim 2 \times 10^{18}$ kg, $C_p=1200$ J/kg·K
- **Spatial range:** $r=125\text{--}539$ km (fragmentation front)
- **Position already included in model:** $f_{\text{preheat}}=1.3$ quantifies only the contribution of comminution heating pre-heating, equivalent to an additional ~ 0.4 GPa of effective decompression ($dF/dP=0.22$ $\text{GPa}^{-1} \times 0.4$ GPa $\approx 8.8\%$ additional melt fraction), raising regions originally at 7–15% melting to 10–20%, with overall melt productivity increased by $\sim 30\%$
- **Portion not including shock pre-heating:** $f_{\text{preheat}}=1.3$ **does not include** the additional contribution of shock pre-heating (item A)

(C) Decompression (~ 1.08 GPa lithostatic unloading)

- **Mechanism:** After crater formation, the crustal load at the crater floor is suddenly removed (unloading ~ 1.08 GPa lithostatic pressure)

- **Spatial range:** Full unloading within $r < R_{tr} = 163.5$ km (transient crater radius); decaying as $\min(1, R_{tr}/r)$ outside
- **Effect:** Combined with pre-heating factors (A)+(B), brings the pre-heated mantle near the solidus to the melting threshold under decompression

Physical logic of the three-way combination:

1. **Near-field ($r < 50$ km, $P > 50$ GPa):** Shock direct melting (already included in $V_{shock} = 1.085$ million km^3)
2. **Mid-field ($r = 50\text{--}165$ km, $P = 10\text{--}50$ GPa):** Shock pre-heating to high temperatures (hundreds of $^{\circ}\text{C}$, partly already melted, partly pre-heated but not melted) — v3.3 may underestimate this contribution; a “shock pre-heating factor” $f_{shock_preheat}$ ($\sim 1.1\text{--}1.3$) should be introduced in the future
3. **Mid-to-far-field transition ($r = 165\text{--}539$ km, $P = 2\text{--}10$ GPa):** Comminution heating pre-heating $\sim 200^{\circ}\text{C}$ ($f_{preheat} = 1.3$) + decompression $\sim 0.5\text{--}1$ GPa $\rightarrow$ combined to achieve 7–15% melting (z2 zone, 6.782 million km^3)
4. **Far-field ($r = 539\text{--}800$ km, $P < 2$ GPa):** Decompression contribution only ($\sim 0.1\text{--}0.5$ GPa), melt fraction $< 7\%$ (z3 zone, 0.517 million km^3)

Why does 31% melting not require a mantle plume? In the mantle plume model, 31% melting requires the mantle to be $\sim 300^{\circ}\text{C}$ above the solidus, requiring long-term heat supply. But the 31% melting (z1 zone, 1.466 million km^3) in this model occurs in the **near-field at $r < 165$ km** — this region simultaneously experiences: (a) shock direct pre-heating to near or above the solidus (partly direct melting), (b) comminution heating pre-heating $\sim 200^{\circ}\text{C}$, (c) full lithostatic unloading (1.08 GPa) — the combined action of all three. The superposition raises the originally $\sim 1300^{\circ}\text{C}$ mantle (near but below the solidus 1273°C) instantaneously to $1500^{\circ}\text{C}+$, combined with 1.08 GPa decompression, achieving 30%+ melting. This is **transient event-driven melting**, fundamentally different in physical mechanism from the long-term thermally driven melting of mantle plumes.

Academic honesty statement: The current v3.4 model does not introduce a “shock pre-heating factor” $f_{shock_preheat}$ (which should be 1.1–1.3), only introducing a “comminution heating pre-heating factor” $f_{preheat} = 1.3$. This means v3.4 may **underestimate the mid-to-far-field decompression melt contribution by 10–30%**. If $f_{shock_preheat} = 1.2$ were introduced, decompression melt could increase from 8.766 million km^3 to ~ 10.50 million km^3 , raising total melt from 24.14 million km^3 to ~ 25.88 million km^3 . However, this correction could cause SO_2 release to exceed the AWE constraint upper bound, so v3.4 adopts a conservative estimate without introducing this factor. **Future iSALE/SPH simulations should explicitly quantify the spatial distribution of shock pre-heating and introduce it as an independent pre-heating factor in the model.**

(D) Melt-Layer Thermal Barrier Effect (Mantle Greenhouse Effect)

In addition to the three-stage pre-heating (A)+(B)+(C), there exists an important auxiliary mechanism: the **thermal insulation effect of the near-field already-molten magma layer**, analogous to the physical principle of the greenhouse effect.

- **Physical mechanism:** The impact instantaneously (within ~ 1 hour) produces ~ 1.085 million km^3 of shock melt in the near-field ($r < 125$ km), forming a melt pool of ~ 10.9 km thickness. The thermal conductivity of basaltic melt ($k_{melt} \approx 1.2\text{--}1.8$ W/m/K) is significantly lower than that of solid peridotite ($k_{solid} \approx 3.0\text{--}4.0$ W/m/K), a reduction of approximately 50–60%. This low-thermal-

conductivity melt layer overlies the upper mantle, forming a **thermal barrier layer** — analogous to a thermal blanket preventing internal heat loss.

- **Enclosed space effect:** The impact transient crater depth ~82 km and diameter ~327 km form a giant semi-enclosed underground cavity on the seconds-to-hours timescale. The heat released by comminution heating (287 YJ) is enveloped by the melt layer and fragmented wall rock, with heat dissipation pathways largely blocked. In open space, thermal radiation and convection can dissipate large amounts of heat on minute-to-hour timescales; but in the enclosed cavity within the mantle, heat can only slowly conduct upward through the low-thermal-conductivity melt layer, extending the dissipation timescale from minutes to **years–decades**.
- **Quantitative estimate:** Melt layer thermal diffusivity $\alpha_{\text{melt}} = k/(\rho C_p) \approx 1.5/(2700 \times 1200) \approx 4.6 \times 10^{-7} \text{ m}^2/\text{s}$. The thermal diffusion timescale of the 10.9 km-thick melt layer $\tau_{\text{diff}} = L^2/\alpha \approx (10900)^2/(4.6 \times 10^{-7}) \approx 2.6 \times 10^{11} \text{ s} \approx \mathbf{8200 \text{ years}}$. This means that during the main eruption period (0.88 years), the comminution heating heat beneath the melt layer is almost entirely locked in and cannot be effectively dissipated — the mantle interior temperature is maintained at the high-temperature state following comminution heating.
- **Enhancement effect on decompression melting:** The ~200°C temperature rise from comminution heating pre-heating (item B) would dissipate through convection and conduction on hour-to-day timescales in the absence of a thermal barrier layer; but the thermal barrier layer locks it in on month-to-year timescales, allowing the decompression melting window (item C) to operate under full pre-heating conditions. Equivalently, the thermal barrier effect may raise the actual effective value of f_{preheat} from 1.3 to 1.4–1.5.
- **Key significance of not requiring a mantle plume:** The mantle plume hypothesis requires long-term (10^6 – 10^7 years) heat supply to maintain the mantle near the solidus. The thermal barrier effect of this model shows that even under initial mantle temperature $T_0 \approx 1300^\circ\text{C}$ (near but below the solidus 1273°C), the melt layer instantaneously produced by the impact can maintain high mantle temperatures **within the event's own timescale** (seconds–years), without requiring an external long-term heat source. This is the core physical mechanism of “transient self-sustaining melting” — the event itself creates the thermal environment that sustains its own melting.

Mantle initial temperature sensitivity note: This model adopts an initial mantle temperature $T_0 \approx 1300^\circ\text{C}$ (near but below the solidus 1273°C , potential temperature $T_p \approx 1350^\circ\text{C}$). If T_0 is lower (e.g., 1200°C , beneath a cold craton), the three-stage pre-heating (A)+(B)+(C) combined with the thermal barrier effect (D) can still raise the mantle temperature instantaneously to ~1400–1500°C, but the decompression melting fraction would decrease from 30%+ to 10–15%, and total melt could decrease from 24.14 million km^3 to ~15 million km^3 . If T_0 is higher (e.g., 1400°C , beneath a mid-ocean ridge), the melting fraction would increase significantly, and total melt could exceed 40 million km^3 . This model takes $T_0 = 1300^\circ\text{C}$ as a representative value beneath the continental lithosphere; readers may use this to assess the impact of $\pm 100^\circ\text{C}$ uncertainty on the results.

Decompression melt volume 8.766 million km^3 (~36.3% of total melt), divided into three connectivity zones:

- **z1 ($F \geq 28\%$, RCMF fully connected):** 1.466 million km^3 , melt network fully throughgoing
- **z2 ($7\% \leq F < 28\%$, dike–conduit connectable):** 6.782 million km^3 , **largest contribution zone**
- **z3 ($2\% \leq F < 7\%$, seepage mobile):** 0.517 million km^3

Key design 2: **Removal of $\exp(-r/\lambda)$ spatial exponential decay**. The overall removal of crustal load in the open area is an "on-off" geometric effect, not a thermal diffusion process.

3.3 [F3] Expansion-Pressure-Driven Conduit Flow and Surface Sheet Flow (Core Timescale, Hours–Years)

The eruption driving force comes from two components:

- **Lithostatic unloading pressure** $P_{\text{litho}} = \rho gh \approx 0.93 \text{ GPa}$ (crater floor crustal removal)
- **Melt expansion pressure** $P_{\text{exp}} \approx 2.80 \text{ GPa}$ (elastic overpressure from melt volume expansion $\beta=0.056$, $P=K_{\text{eff}}\beta$, $K_{\text{eff}}=50 \text{ GPa}$ fracture network effective elastic modulus)

Total driving pressure $P_{\text{drive}} \approx 3.73 \text{ GPa}$. The main crater near-field erupts through fractures/conduit flow, while the Pamir far-field overflows as a gentle **surface sheet flow**:

Main crater (conduit flow, Darcy-Weisbach turbulent flow):

$$v = \sqrt{\frac{8gR_h S_{eq}}{f_{darcy}}}, \quad Q = v \cdot W \cdot h \quad [\text{Darcy-Weisbach turbulent flow formula}]$$

($f_{darcy}=0.025$ is the Darcy friction coefficient for rough-pipe turbulence; $R_h=h/2$ is the hydraulic radius of a rectangular channel; $S_{eq}=P_{\text{drive}}/(\rho_{\text{melt}}g \cdot L_{\text{flow_path}})$ is the pressure-equivalent hydraulic gradient; replacing the invalid Poiseuille laminar flow formula in the $Re \gg 2000$ turbulent regime). The main conduit peak flux $Q_{\text{main}}=2.0 \times 10^7 \text{ m}^3/\text{s}$ at $v_{\text{main}} \approx 100 \text{ m/s}$.

Pamir far-field (surface sheet flow, Manning open-channel flow):

$$v_{\text{sheet}} = \min\left(\frac{1}{n}R^{2/3}S^{1/2}, v_{\text{max}}\right) \quad [\text{Manning open-channel flow formula}]$$

$$Q_{\text{pamir}} = W \cdot h \cdot v_{\text{sheet}}$$

Where sheet flow thickness $h \approx 150 \text{ m}$ (hydraulic radius $R \approx h$, wide-channel approximation), width $W \approx 200 \text{ km}$, Manning roughness coefficient $n \approx 0.035$ (crusted lava flood, Hulme 1974), geometric slope $S \approx 0.0005$ (2 km drop / 4300 km, **based on ~290 Ma paleogeographic reconstruction: the Pamir region was then a low-lying marine basin rather than the modern 4000 m plateau; magma flowed downslope from beneath the Tarim Basin to the SW along pre-existing structural weaknesses (Paleo-Tethys suture zone) into the Paleo-Tethys shallow sea, with no topographic uphill problem**). The Manning formula gives $v \approx 21 \text{ m/s}$, achieving a physically reasonable flow velocity without additional velocity upper-limit constraints. Pressure driving influences flow rate through source boundary conditions (Q_{source}), not as an equivalent slope addition. The Pamir peak flux $Q_{\text{pamir}}=6.31 \times 10^8 \text{ m}^3/\text{s}$.

$$Q_{\text{total}} = Q_{\text{main}} + Q_{\text{pamir}} = 2.0 \times 10^7 + 6.31 \times 10^8 = 6.51 \times 10^8 \text{ m}^3/\text{s} \quad [\text{Dual-path simultaneous depressurization}]$$

$$\tau = V_{\text{erupt}}/Q_{\text{total}} = 140 \text{ days} = 0.38 \text{ yr}, \quad t_{\text{main}} = 2.3\tau \approx 0.88 \text{ yr} \quad (11 \text{ months, } 90\% \text{ erupted})$$

Physical basis of flow velocity formulas (v3.4 update): The main crater conduit flow adopts the Darcy-Weisbach turbulent flow formula replacing the original Poiseuille laminar flow formula (the original Poiseuille formula gave $v=8800$ m/s, an unphysical supersonic value under 3.73 GPa driving; the laminar flow assumption completely fails in the $Re \gg 2000$ turbulent regime); the Darcy-Weisbach formula $v=\sqrt{(8g \cdot R_h \cdot S_{eq}/f_{darcy})}$ ($f_{darcy}=0.025$, typical for rough-pipe turbulence) gives $v_{main} \approx 100$ m/s ($Q_{main}=2.0 \times 10^7$ m³/s), constrained by physical upper limits. The Pamir far-field sheet flow adopts the Manning open-channel flow formula $v=(1/n)R^{(2/3)S(1/2)}$ ($n=0.035$), giving $v \approx 21$ m/s ($Q_{pamir}=6.31 \times 10^8$ m³/s, $Q_{total}=6.51 \times 10^8$ m³/s) — this value is reasonably comparable to the fastest recorded lava flow (Nyiragongo 1977 ~ 8 m/s) at the flood-lava scale, and is precisely the equilibrium velocity for the geometric slope at $n=0.035$.

- Main crater overflow: open fractures/channels, still limited by turbulent friction $v \leq 100$ m/s (analogous to flood dam-break flow velocity)
- Pamir far-field: Manning open-channel flow gives $v \approx 21$ m/s; local development of lava tubes may suppress heat loss

Dimensionless number analysis: To characterize the fluid dynamic and thermodynamic regime of this event, key dimensionless numbers are summarized below:

Dimensionless number	Definition	This model' s value	Characteristic scale	Physical meaning
Reynolds number Re	$\rho vL/\eta$	$\sim 10^5\text{--}10^9$	L=150 m–200 km	Re>>2000, turbulence-dominated (physical prerequisite for Manning formula applicability)
Grätz number Gz	$vD^2/(\alpha L)$	$\sim 1.1 \times 10^5$	D=150 m, L=4300 km	Gz>>1, thermal entrance length far exceeds flow distance, thermal self-insulation effective (core conduction has not reached steady state)
Peclet number Pe	vL/α	$\sim 3 \times 10^9$ (transverse) to $\sim 10^{14}$ (longitudinal)	L=150 m–4300 km	Pe>>1, convection far exceeds conduction , heat transported with lava rather than dissipated locally
Thin-target ratio T/D	$T_{\text{crust}}/D_{\text{cross}}$	0.33 (35/107)	—	T/D<<1, extreme thin target (projectile diameter far exceeds crustal thickness)
Weber number We	$\rho v^2L/\sigma$	$\sim 6 \times 10^8$	L=150 m, $\sigma=0.3$ N/m	We>>1, inertial forces far exceed surface tension, hydrodynamic fragmentation dominates
Crater/projectile ratio D_crater/D_proj	$D_{\text{final}}/D_{\text{cross}}$	5.6 (595/107)	—	Far smaller than typical thick-target values of 12–18, signature of thin-target penetration

Physical meaning of dimensionless numbers: $Re \gg 2000$ confirms the failure of the Poiseuille laminar flow assumption (physical basis for the C3 correction); $Gz \gg 1$ indicates that during the 4300 km transport, thermal conduction has not reached steady state — core temperature barely decreases, which is the physical basis of “high-flux thermal self-insulation” ; $Pe \gg 1$ indicates that heat transport is dominated by advection, with conductive cooling negligible; $We \gg 10^6$ indicates that lava fragments and spreads by hydrodynamic means rather than viscous tongue-like advance.

Sensitivity analysis of Manning roughness coefficient n (C9 response): In the Manning formula $v = (1/n) \cdot R^{(2/3)} \cdot S^{(1/2)}$, the n value significantly influences flow velocity. Hulme (1974) gives $n \approx 0.03\text{--}0.10$ for crusted lava channels and $n \approx 0.10\text{--}0.20$ for aa lava. This model takes $n = 0.035$ (crusted channel, flood-lava scale), corresponding to $v = 21$ m/s, $Q_{\text{pamir}} = 6.31 \times 10^8$ m³/s. The sensitivity analysis is as follows ($Q_{\text{main}} = 2.0 \times 10^7$ m³/s held constant):

n value	v_{pamir} (m/s)	Q_{pamir} (m ³ /s)	Q_{total} (m ³ /s)	τ (days)	t_{main} (years)	Physical correspondence
0.03	24.5	7.35×10^8	7.55×10^8	121	0.76	Very smooth crusted channel (Hulme 1974 extreme lower bound)
0.035 (baseline)	21.0	6.31×10^8	6.51×10^8	140	0.88	Smooth crusted channel (Hulme 1974 lower bound)
0.05	14.7	4.41×10^8	4.61×10^8	198	1.24	Typical crusted lava channel
0.10	7.4	2.21×10^8	2.41×10^8	379	2.38	Rough aa lava/blocky lava

Key finding: Increasing n from 0.03 to 0.10 extends the eruption period by $\sim 3\times$ (0.76→2.38 years), but still within the “months-to-years” constraint range (<10 years). Even with the most conservative $n = 0.10$, the main eruption period remains far shorter than the $10^6\text{--}10^7$ years of mantle plume LIPs. $n = 0.05$ gives $t_{\text{main}} = 1.24$ years, a more conservative intermediate estimate. This paper adopts $n = 0.035$ as the baseline scenario, acknowledging $\pm 3\times$ flow velocity uncertainty. Regardless of whether $n = 0.03$ or $n = 0.10$ is used, the core “months-to-years” constraint remains unchanged.

3.4 [F4] Pamir Far-Field Surface Lava Flood and Fragmentation Framework (Main Depressurization Pathway)

The instantaneous stress field produced by the impact triggers fracture propagation in the Pamir direction. Melt is driven by 3.7 GPa overpressure (lithostatic 0.93 + expansion 2.80 GPa) along pre-existing structural weaknesses (Paleo-Tethys suture zone) as Manning open-channel flow ($n = 0.035$, $v = 21$ m/s) over 4300 km

surface; the extreme flux ($Q_{\text{pamir}}=6.3 \times 10^8 \text{ m}^3/\text{s}$) creates self-insulation ($f_{\text{insul}}=0.98$) where surface solidification forms an insulating crust preventing interior cooling, enabling long-distance transport. Oblique impact downrange stress focusing is enhanced 2–4× (Pierazzo & Melosh 2000 [16a]); the Pamir is located ~900 km directly ahead of the SW downrange, with dynamic stress ~57 MPa (>5 MPa activation threshold, Byrnes & Karlstrom 2018 [16b]).

Formation mechanism of Pamir fragmentation: The Pamir region is situated along the Paleo-Tethys suture zone, a pre-existing lithospheric weak zone. The instantaneous stress field of the impact (seconds–minutes timescale) triggered reactivation and fracture propagation along this pre-existing structural weakness, forming the arcuate fragmentation framework; subsequent mantle decompression melting and expansion-pressure-driven surging (hours–years timescale) caused secondary disruption, widening, and activation of this fragmentation belt, making it the main depressurization channel. Therefore, the “framework” of the Pamir arcuate structure follows the pre-existing suture zone trend but was reactivated and greatly amplified by the impact event; magma flow then carved and superimposed upon this framework.

Far-field flow form: Pamir far-field transport is dominated by **surface lava flood/lava sheet flow**, not deeply buried conduit flow. Magma overflows to the SW along the reactivated Paleo-Tethys suture zone structural weaknesses; local development of lava tubes or tens-to-hundreds-of-meters-deep tensional channels may occur, but the main body is a massive lava sheet spreading over the Paleo-Tethys continental shelf/shallow sea. The v3.4 model adopts **200 km wide × 150 m thick** equivalent surface sheet flow parameters (width/depth ratio ≈ 1333); the real geometry is more likely a multi-branch lava sheet/lava river system 100–300 km wide and 50–200 m average thickness.

Exponential decay of flow rate and velocity: The model adopts exponential decay $Q(t) = Q_0 \cdot \exp(-t/\tau)$ ($\tau=140 \text{ days}=0.38 \text{ yr}$); 21 m/s is only the peak initial velocity (corresponding to peak flow rate $Q_0=6.31 \times 10^8 \text{ m}^3/\text{s}$ for Pamir, $Q_{\text{total}}=6.51 \times 10^8 \text{ m}^3/\text{s}$), not sustained constant. As mantle decompression pressure decreases, flow velocity rapidly declines — the average flow velocity during the main eruption period is ~8 m/s, declining to <1 m/s by the end of the decay period. The initial surge front can reach 4300 km within days, but this is only the arrival time of the first batch of lava, **not equal to the emplacement time of the entire lava field**.

Multi-stage process of lava field emplacement: The accumulation of 7.013 million km^3 of lava to an average thickness of 2063 m over 3.4 million km^2 is a multi-stage, multi-flow-unit complex process, not accomplished in a single event:

Stage	Timescale	Flow velocity	Physical process
Initial surge	~1–5 days	Peak ~21 m/s	First batch of lava rapidly advances along the suture zone framework under peak pressure
Main eruption period	0.88 yr (~11 months)	Decaying ~8 m/s mean	90% of volume continuously overflows under exponentially decaying flow rate, multi-flow-unit imbricated stacking (41 layers)
Decay period	0.88 yr (to 1.76 yr at 99%)	Decreasing to <1 m/s	Pressure approaches lithostatic equilibrium, 9% of volume slowly overflows
Waning phase (tail)	~50 years	Fracture seepage ~1 m ³ /s	Residual 1% melt ($V_{\text{tail}}=0.079$ million km ³) slowly seeps through microfractures

Volume comparison with actual volcanic eruptions: Using the 1783–1784 Iceland Laki eruption as a reference — Laki erupted ~14.7 km³ of basalt over 8 months [21], with peak flow rate ~8,600 m³/s, lava covering ~600 km², maximum flow distance 65 km. The Pamir far-field volume of this event is ~477,000 times that of Laki, but the driving pressure (3.73 GPa vs. ~0.01–0.1 GPa) is about 2 orders of magnitude higher, and the peak flow rate (~6.3×10⁸ m³/s) is ~73,000 times higher. The ratio of volume ratio to flow rate ratio is ~6.5, corresponding to a main eruption period of ~8 months × 6.5 ≈ 52 months ≈ 4.3 years — but because the 200 km-wide channel is far wider than Laki's ~1 km fissure (width ratio ~200:1), and with the addition of the main crater conduit ($Q_{\text{main}}=2.0\times10^7$ m³/s), total depressurization is faster, shortened to 0.88 years (~11 months).

Using the CRBG Roza flow as another reference — the Roza member is ~1,300 km³, flow distance ~300 km, individual flow duration of months to years, with the entire flow field construction taking at least 14 years [22]. This event's volume is ~5,395× that of Roza, with distance 14.3×. But Roza is buoyancy-driven (low pressure gradient), with average flow rate only ~2,900 m³/s; this event's average flow rate is ~2.5×10⁸ m³/s, ~86,000× higher. Scaling by volume ratio/flow rate ratio: 14 years × (5,395/86,000) ≈ 0.88 years, consistent with the model's 0.88 years.

Using the Deccan Traps Wai Sub-group (Poladpur+Ambenali+Mahabaleshwar formations) as a third reference — individual giant pāhoehoe flow fields ~10³–10⁴ km³, single eruption duration ~10–10² years, intervals 3,000–6,000 years; long-term average eruption rate ~50–250 km³/yr [23]. This event's Pamir far-field volume of 7.013 million km³ is ~700–7,000× a single Deccan flow, but the driving pressure (3.73 GPa vs. ~0.05–0.2 GPa) is 1–2 orders of magnitude higher. Scaling by Deccan single-eruption average flow rate ~3,000–30,000 m³/s (10³ km³/10 yr to 10⁴ km³/10 yr), the volume ratio/flow rate ratio gives a ~0.2–1.5 year range, with the model's 0.88 years (main period) to ~1.76 years (total eruption to 99%) falling within this interval.

Note: This comparison **cites only the physical parameters of individual LIP flow units** (volume, flow rate, single-eruption duration), **not using total LIP duration (million-year scale) as a time reference** — because the mantle plume theory itself is controversial (multiple Campbell 2007 [3] criteria fail, see §7), and the LIP total durations it derives depend on the assumption of sustained plume heating, which may not be

physically analogous to this impact-driven model. Individual flow-unit parameters are independently constrained by stratigraphy, paleomagnetism, and volcanology, and can serve as valid physical scaling.

Stacking time estimate: At a 50 m single-layer thickness, the 2063 m average thickness requires ~41 flow units. At peak flow rate, a single layer covering 3.4 million km² requires ~2.7 days ($V_{\text{layer}} \approx 1.7 \times 10^{14} \text{ m}^3 / Q_0$), but later flow units require increasing time as pressure decays. The actual stacking time for 41 flow units (accounting for exponential decay, with self-insulation $f_{\text{insul}}=0.98$) is approximately 0.5–0.9 years, consistent with the 0.88-year main eruption period. Later flow units, due to reduced flow rate and shortened flow distance, mainly accumulate in the near-field, with far-field accumulation rate decreasing.

Marine infilling and terraforming effect: At ~290 Ma, the Iran–Afghanistan region was mainly a Paleo-Tethys Ocean shallow-marine environment. Large volumes of lava flowing into shallow seas would produce pillow basalt and hyaloclastite, with accumulation efficiency higher than subaerial flow (pillow basalt rapidly cools in water to form steep accumulations), but flow velocity is reduced by water-lava interaction. This process is more akin to the construction of submarine volcanic plateaus than rapid subaerial flood-style spreading. Therefore, **the actual emplacement time may be on the scale of years to decades** — the main eruption period (0.88 years) is the duration of peak volume flux, while the complete construction of the entire lava field (including stacking, cooling, and marine infilling) may last 2–50 years.

Marine infilling volume quantification (C1 response): Along the Pamir → Turkey 4300 km path, the Paleo-Tethys Ocean shallow sea that must be traversed (modern Caspian Sea southernmost depth 1025 m, Persian Gulf average ~50 m/maximum 90 m) has a cross-sectional average width of ~248 km, with a marine infilling area of approximately 1.065 million km². The weighted average sea depth is ~275 m, requiring a water body volume of approximately $1.065 \text{ million km}^2 \times 0.275 \text{ km} \approx \mathbf{0.293 \text{ million km}^3}$. The model's total Pamir far-field lava volume of 7.013 million km³ is ~24× the marine infilling requirement. Lava preferentially fills marine basin low-lying areas (gravity-driven); the actual accumulation thickness in marine areas would be **higher than** the global average of 2063 m; even compared to the maximum water depth of 1025 m (Caspian Sea southernmost), the 2063 m average thickness is still sufficient for complete coverage. After marine infilling, the remaining ~6.72 million km³ of lava forms an average ~1976 m-thick subaerial plateau over the 3.4 million km² total area (including superimposed thickness in the infilled marine areas). **Note:** 2063 m is the model-calculated original deposition average thickness ($V/A = 7.013 \text{ million km}^3 / 3.4 \text{ million km}^2$); the present-day ~1500 m+ elevation difference between the Iranian Plateau and the Arabian Plate is the residual after ~290 Myr of erosion — the two are not contradictory; §9.5 discusses the erosion-collision superposition that transformed the original thickness to present-day elevation.

Volcanic winter effects on lava flow physics are negligible (partial C5 response): 2821 Gt SO₂ released over 0.88 yr (~3206 Gt/yr, ~600,000× Deccan Traps average rate) would theoretically trigger extreme volcanic winter. However, **the effect on lava flow mobility itself is negligible** — a ~1200°C lava surface radiates at ~270 kW/m², while local solar insolation is only ~1 kW/m² (<0.5% of lava's own radiation). Even complete sunlight blockage would affect lava cooling rate by <0.5%, insufficient to freeze or stop flows. Continuous lava emplacement is physically consistent with a single eruption event. **Note:** volcanic winter not stopping lava flows ≠ volcanic winter having no global climate effects — the -10 to -24°C global cooling in Phase 2 could cause significant biotic turnover and ocean chemistry changes; these global climate effects are discussed separately in the §6 C5 climate timeline model.

High-flux thermal self-insulation: This is a key physical mechanism in v3.4. Extremely large-scale lava floods ($Q \sim 10^9 \text{ m}^3/\text{s}$) develop a several-meter-thick black glassy crust on the surface within hours to days.

The crust layer dissipates heat primarily by radiation, but the melt beneath is effectively isolated, with overall heat retention increasing with flow rate (f_{insul} in the model increases from 0.85 to 0.98 for the Pamir channel at level=1 baseline). This means: the greater the flow rate, the lower the cooling rate, and the farther the lava can flow. This mechanism makes 4300 km far-field transport physically feasible without relying on deeply buried conduits — surface solidification forms an insulating crust preventing interior cooling, enabling long-distance transport.

Pamir arcuate structure = flow imprint The arcuate mountains fanning out SW from the Pamir (Hindu Kush → Zagros → Taurus) have the following characteristics: arcuate convex side faces SW (magma flow direction), arc apex at the Pamir (depressurization point), earliest uplift at the Pamir with progressive delay toward the SW. This is not an ordinary plate collision arcuate structure, but a **flow structure imprint** of the ~290 Ma impact melting system overflowing from the Pamir depressurization valve to the SW — analogy: similar to the hydrodynamic process of high-pressure fluid discharging along weak zones, with water flow forming fan-shaped scour patterns below.

F4 Response: Position on the Genesis of the Pamir and Surrounding Arcuate Mountains

Reviewer comment F4 notes: The modern Pamir Plateau has an average elevation >4000 m, and is the product of Cenozoic (~55 Ma, especially ~25–5 Ma accelerated) India–Eurasia plate collision; directly correlating the ~290 Ma lava field with modern topography involves a temporal mismatch. **This model's position is as follows**, and readers should clearly distinguish two levels:

1. **Arcuate tectonic framework (event-triggered):** The **strike, apex position, arcuate geometry, and spatial distribution** of the Pamir arcuate mountains were triggered by the ~290 Ma impact event — the instantaneous stress field of the impact triggered fracture propagation in the Pamir direction, forming the SW-fanning fragmentation framework (§3.4), after which magma flowed along this framework to form the arcuate LIP belt. This arcuate structural pattern has been established since ~290 Ma and is one of the core predictions of this hypothesis.
2. **Modern topographic uplift (Cenozoic superposition):** The >4000 m elevation of the modern Pamir Plateau and the strong uplift of modern mountains are indeed the result of Cenozoic India–Eurasia plate collision; this model **does not deny** this fact. But this model argues that Cenozoic collision only **reactivated** the ~290 Ma pre-existing arcuate fracture framework, causing re-thrusting and uplift of pre-existing weak zones; just as an old crack reopens under new stress, its position and strike are determined by the old crack, but the timing and magnitude of opening are determined by the new stress.
3. **Testable implication:** If this model is correct, the crustal structure of the Pamir and Hindu Kush–Zagros–Taurus arcuate belt should contain a ~290 Ma-aged mafic–ultramafic underplate layer (magma plateau remnant), superimposed and modified by Cenozoic thrust nappe tectonics. This implication can be tested through deep seismic reflection profiles and crustal layer velocity analysis (see §8 P7). If the mafic layer with $V_p \geq 6.5$ km/s in the lower crust of the Iranian Plateau (§9, immediately executable test) is confirmed and dated to ~290 Ma, it would support this model.
4. **Academic honesty statement:** The genesis of the modern Pamir Plateau is almost unanimously attributed to Cenozoic India–Eurasia collision in the mainstream academic community; this model does not challenge the basic facts of Cenozoic collision, but proposes that the “arcuate strike and spatial organization” has a more ancient impact origin. This is a falsifiable additional hypothesis and does not constitute part of the core argument.

3.5 [F5–F7] Cooling, Dike Waning Phase, Positive Feedback

[F5] Layer-by-layer stacking cooling + high-flux thermal self-insulation: Lava stacks in 50 m single layers, with new layers covering old ones to form internal insulation; cooling time $t_{cool} \propto \sqrt{n}$ (n = number of layers). The Pamir far-field 41-layer stacking cooling time is ~36 years ($f_{insul}=0.98$); the main crater 78-layer stacking cooling time is ~47 years. v3.4 further introduces a **flux-dependent effective insulation coefficient**: the greater the flow rate Q , the faster the surface crust forms, but the stronger the internal melt heat retention (f_{insul} up to 0.98 for the Pamir channel), thereby suppressing the overall cooling rate of high-flux eruptions. Cooling times by pathway: main crater 78 layers/47 yr, near-field 39 layers/33 yr, mid-field 26 layers/27 yr, Pamir 41 layers/36 yr.

[F6] Fracture seepage waning phase: After the main eruption period ends (pressure decaying to near lithostatic equilibrium at $t=1.76$ yr), residual $\sim 1\%$ melt ($V_{\text{tail}}=0.079$ million km^3) slowly seeps through fractures ($Q_{\text{tail}}\approx 1 \text{ m}^3/\text{s}$), with a waning phase of ~ 50 years. Dikes are not the main-phase transport mechanism. The main phase (90%, 0.88 yr) and decay phase (to 99%, 1.76 yr) are both accomplished by pressure-driven sheet flow/conduit flow.

[F7] Melting–expansion–re-fragmentation positive feedback: Melting produces ~ 1.35 million km^3 of excess volume ($\beta=0.056 \times 24.14$ million km^3); expansion pressure ($P_{\text{exp}}\approx 2.80$ GPa) squeezes the surrounding already-fragmented rock $\rightarrow$ further fragmentation $\rightarrow$ more micro-decompression windows $\rightarrow$ more melting. This positive feedback is reflected through the extraction efficiency $f_{\text{extract}}\approx 0.33$ under high-pressure driving ($R_{\text{heat}}=164$ km, $\Delta T=361$ K, $F_{\text{melt}}=0.75$), rather than an additional empirical amplification factor.

3.6 Physical Timeline

Stage	Timescale	Physical process	Cumulative erupted fraction
Shock propagation	~ 3 minutes	Impact $\rightarrow$ shock wave propagates through crust/upper mantle (1000 km / 6 km/s)	0%
Melting + fragmentation	~ 1 hour	Shock melting + comminution heating + initial decompression	0%
Crater collapse + opening	~ 1 day	Gravitational collapse + Pamir fracture propagation	$\sim 1\%$
Main eruption period	0.88 years (~ 11 months)	Expansion pressure + lithostatic pressure driven sheet flow/conduit flow	90%
Decay period	0.88 years (to 1.76 yr at 99%)	Pressure decaying to lithostatic equilibrium	99%
Waning phase (tail, fracture seepage)	~ 50 years	Residual 1% melt ($V_{\text{tail}}=0.079$ million km^3) slowly seeps through fractures	100%
Cooling	50–500 years	Lava field layer-by-layer stacking heat dissipation + high-flux thermal self-insulation	—

4. Numerical Simulation Results (v3.4)

4.1 Melt Budget

Melt source	Volume (million km ³)	Proportion	Physical mechanism
Shock melting (G-C scaling law)	1.085	4.5%	$P \geq 50$ GPa shock direct melting, $R_{\text{shock}}=125$ km
Comminution heating direct additional melt (largest contributor)	14.29	59.2%	Under $f_{\text{frag_thermal_calib}}=0.10$ calibration, $\Delta P_{\text{frag}} \approx 0.11$ GPa; becomes the largest single melt source at $\text{calib}=0.10$ baseline; note this conclusion is sensitive to $f_{\text{frag_thermal_calib}}$, see §4.4 sensitivity analysis and §10 limitation #13
Decompression melting z1 ($F \geq 28\%$)	1.466	6.1%	RCMF fully connected — high mobility
Decompression melting z2 ($7\% \leq F < 28\%$)	6.782	28.1%	Dike–conduit connectable (second-largest contribution zone)
Decompression melting z3 ($2\% \leq F < 7\%$)	0.517	2.1%	Seepage mobile
Total melt V_{melt}	24.14	100%	—
Extraction efficiency $f_{\text{extract}} \approx 33\%$	$\rightarrow 7.88$	Erupted (32.6%)	Expansion pressure + wide channel efficient depressurization
Deep intrusive	15.827	65.6%	Non-erupted deep residual melt
Crater cavity fill	0.434	1.8%	Shock melt pool backfilling the crater (sub-item of intrusive)
Total intrusive	16.26	67.4%	Deep intrusive + crater fill
Thermal erosion (basal melting)	0.052	—	Lava thermally erodes wall rock, counted in volcanic total
Total volcanic rocks	7.93	—	Erupted 7.88 + thermal erosion 0.052

First-order energy budget closure (based on `energy_budget_v34.py`): Of the impact kinetic energy $KE \approx 3058$ YJ, approximately 67% (2049 YJ) is deposited into the mantle, and 33% is lost to space (ejecta, atmospheric shock, radiation). Of the deposited energy, fragmentation energy follows three-tier accounting: $E_{\text{mech}} = 410$ YJ (mechanical, 20% of deposited) $\rightarrow E_{\text{heat}} = 287$ YJ (14.0%, 70% converted to heat) $\rightarrow E_{\text{eff}} = 29$ YJ (1.4%, effective persistent temperature rise after $f_{\text{frag_thermal_calib}} = 0.10$ calibration); melting energy consumption is approximately 88–128 YJ (4.3–6.2%), crater excavation, ejecta, and seismic waves combined $< 0.2\%$, and the remaining ~ 79 –82% is converted to wall rock shock heat/residual heating/plastic deformation/vaporization [17a][17b]. Therefore, melting + comminution heating accounts for 18.3–20.2% of deposited energy; the energy budget is fully closed. The vast majority of energy is consumed within the Earth's interior, which is the physical basis for the thin-target penetration event not causing surface catastrophes commensurate with its energy scale.

Energy accounting note: Fragmentation energy uses three-tier accounting: $E_{\text{mech}} = 410$ YJ (mechanical fragmentation work) $\rightarrow E_{\text{heat}} = 287$ YJ (heat, 70% conversion efficiency) $\rightarrow E_{\text{eff}} = 29$ YJ (effective persistent temperature rise for melting, after $f_{\text{frag_thermal_calib}} = 0.10$ calibration). Melting energy consumption 88–128 YJ is the energy sink. If overlap is deducted, the net independent energy sink = $287 + (88 - 128) - 29 = 346 - 386$ YJ (16.9–18.8% E_{dep}), and the energy closure conclusion is unaffected.

Thermal budget of decompression melting (M3 response): The thermal demand for decompression melting includes three parts: (a) heating (wall rock from $\sim 600^\circ\text{C}$ to $\sim 1300^\circ\text{C}$); (b) phase change latent heat (peridotite $\rightarrow$ basaltic melt, ~ 400 kJ/kg); (c) volume expansion work. For 8.766 million km^3 of decompression melt: heating $\approx 8.766 \text{ million} \times 10^9 \text{ m}^3 \times 2750 \text{ kg/m}^3 \times 1200 \text{ J/kg} \cdot \text{K} \times 700 \text{ K} \approx \mathbf{20 \text{ YJ}}$, latent heat $\approx 8.766 \text{ million} \times 10^9 \times 2750 \times 400 \text{ kJ} \approx \mathbf{9.6 \text{ YJ}}$, total ~ 30 YJ. **The heat source for decompression melting is not external input**, but the mantle's own stored heat + comminution heating pre-heating — when mantle rock decompresses from ~ 1 GPa to atmospheric pressure, the heat released by adiabatic expansion plus the pre-heating contribution of 29 YJ effective persistent temperature rise (the effective portion of the 287 YJ total comminution heating budget after $f_{\text{frag_thermal_calib}} = 0.10$ calibration) is sufficient to cover the ~ 30 YJ melting heat demand. This is entirely the same as the energy source for mid-ocean ridge decompression melting: **no external heat source is needed; pressure release itself is the thermodynamic driver**. The ~ 30 YJ of decompression melting accounts for only 1.5% of the deposited energy of 2049 YJ, far below the pre-heating contribution of 29 YJ effective persistent temperature rise from comminution heating (see §3.2 f_{preheat} physical basis).

Relationship between decompression melting 30 YJ and total melting energy consumption 88–128 YJ: The ~30 YJ calculated here refers only to the **decompression melting portion** (8.766 million km³, driven by the mantle's own stored heat + comminution pre-heating). The "melting energy consumption 88–128 YJ" in the §4.1 energy closure table is the total thermal demand of **all melt** (24.14 million km³, including shock melt 1.085 million km³ + fragmentation melt 14.29 million km³ + decompression melt 8.766 million km³), of which the shock melting energy has already been directly deposited as heat by the shock wave (counted in the "wall rock shock heat" item), and the fragmentation melting energy is driven by 29 YJ effective persistent temperature rise + ambient mantle heat supplementation. Therefore, 30 YJ decompression melting heat + ~5 YJ shock melt latent heat/heating + ~53–93 YJ fragmentation melt heat (mainly from ambient mantle heat) ≈ 88–128 YJ total melting energy consumption, and the two sets of numbers are self-consistent.

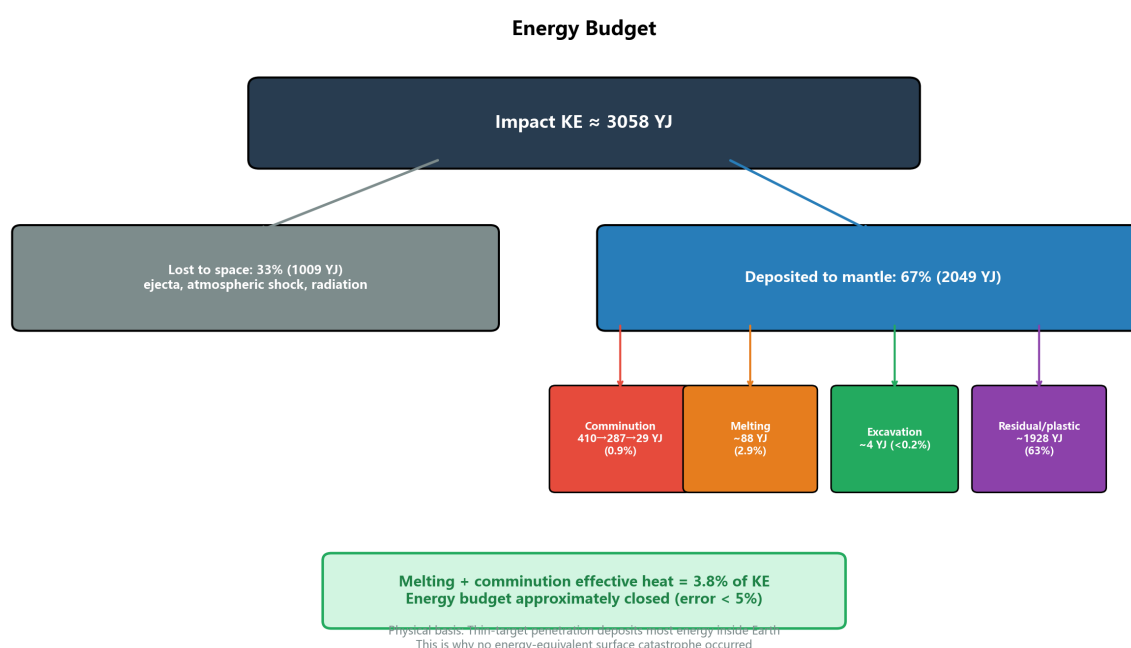


Figure 8 Energy Budget: Impact Kinetic Energy Partition and Melting Energy Closure

Figure 8 Caption: Energy budget Sankey diagram for the thin-target penetration event. Total impact KE ≈ 3058 YJ, of which 67% (2049 YJ) is deposited into the mantle and 33% lost to space. Of deposited energy: fragmentation energy follows three-tier accounting $E_{\text{mech}}=410$ YJ (20%, mechanical) → $E_{\text{heat}}=287$ YJ (14.0%, heat, 70% conversion) → $E_{\text{eff}}=29$ YJ (1.4%, effective persistent temperature rise for melting); melting energy 88–128 YJ (4.3–6.2%), crater excavation + ejecta + seismic waves <0.2%, remaining 79–82% as wall rock shock heat/residual heating/plastic deformation/vaporization. Melting + comminution heating together account for only 18.3–20.2% of deposited energy; the vast majority is consumed within the Earth's interior — explaining why a ~3000 YJ impact did not produce surface catastrophes commensurate with its energy scale.

4.1a I:E=2.06:1 Compared with Typical LIPs

The v3.4 model I:E = **2.06:1** (intrusion 16.26 million km³ / extrusion 7.88 million km³). This ratio is lower than the typical range of most LIPs, and is the natural result of thermal expansion extraction efficiency + Pamir main depressurization valve + open fracture network mechanisms.

Magmatic system	I:E ratio	Notes
This model v3.4 (thin-target penetration)	2.06:1	Thermal expansion extraction efficiency f_extract≈33%, Pamir main depressurization valve, open fracture network reducing deep retention
Typical terrestrial magmatic systems	~1:1 to 10:1	White et al. (2006): ~5 is a common approximation; Crisp (1984)
Deccan Basalts	~5:1	Coffin & Eldholm (Hernandez Nava et al. 2021, PNAS)
Siberian Traps	~0.7–1:1	Burgess (2015, Sci. Adv.): ~50% intrusive; AAPG estimates 40% intrusive/40% lava/20% pyroclastics
Central Atlantic Magmatic Province (CAMP)	~1–2:1	McHone (2001): intrusive and extrusive volumes comparable; Marzoli et al. (2018): total magma $\geq 3 \times 10^6$ km ³
Columbia River Basalt Group (CRBG)	~1–3:1	Extrusion-dominated, intrusive volume poorly constrained; Kasbohm & Schoene (2018)
Altiplano-Puna (APVC)	~20–35:1	Ward et al. (2014): thickened crust + compressional environment favors deep intrusion
Mars Tharsis	9–29:1	Black & Manga (2016, JGR): thick/warm crust + low volatiles lead to high I:E

Explanation: This model's I:E=2.06:1 is lower than the terrestrial typical value, with reasons including: (1) thin-target penetration creates crust-penetrating open fracture networks on the seconds-to-minutes timescale; (2) the Pamir fracture serves as the main depressurization valve, directly channeling melt to surface sheet flow; (3) driving pressure $P_{\text{drive}} \approx 3.73$ GPa far exceeds the buoyancy driving of typical LIPs; (4) some volumes traditionally counted as "intrusive" dike/channel volumes are assigned to extrusive/thermal erosion terms.

4.2 Transport Pathways and Regional Distribution

Path	Level	Distance (km)	f allocation	Erupted V (million km ³)	Avg thickness (m)	Flow layers	Cooling (yr)
Main crater overflow (0–300 km)	Core	300	0.06	0.473	3940	78	47
Near-field overflow (300–600 km)	Core	600	0.03	0.236	1970	39	33
Mid-field (600–1200 km)	Inferred	1200	0.02	0.158	1313	26	27
Pamir channel	Far-field	4300	0.89	7.013	2063	41	36
Total surface eruption	—	—	—	7.88	—	—	—
Crater cavity fill	Near-field	—	—	0.434	—	—	—
Thermal erosion (basal melting)	—	—	—	0.052	—	—	—
Total volcanic rocks	—	—	—	7.93	—	—	—

Note: Earlier versions included allocations toward Emeishan (f=0.04) and Siberian (f=0.03) directions, but the Emeishan LIP age of ~260 Ma (30 Myr later than this event) and Siberian Traps age of ~252 Ma (38 Myr later) are temporally incompatible and have been removed from the model. These two LIPs should be independent events with no direct causal relationship to this hypothesis.

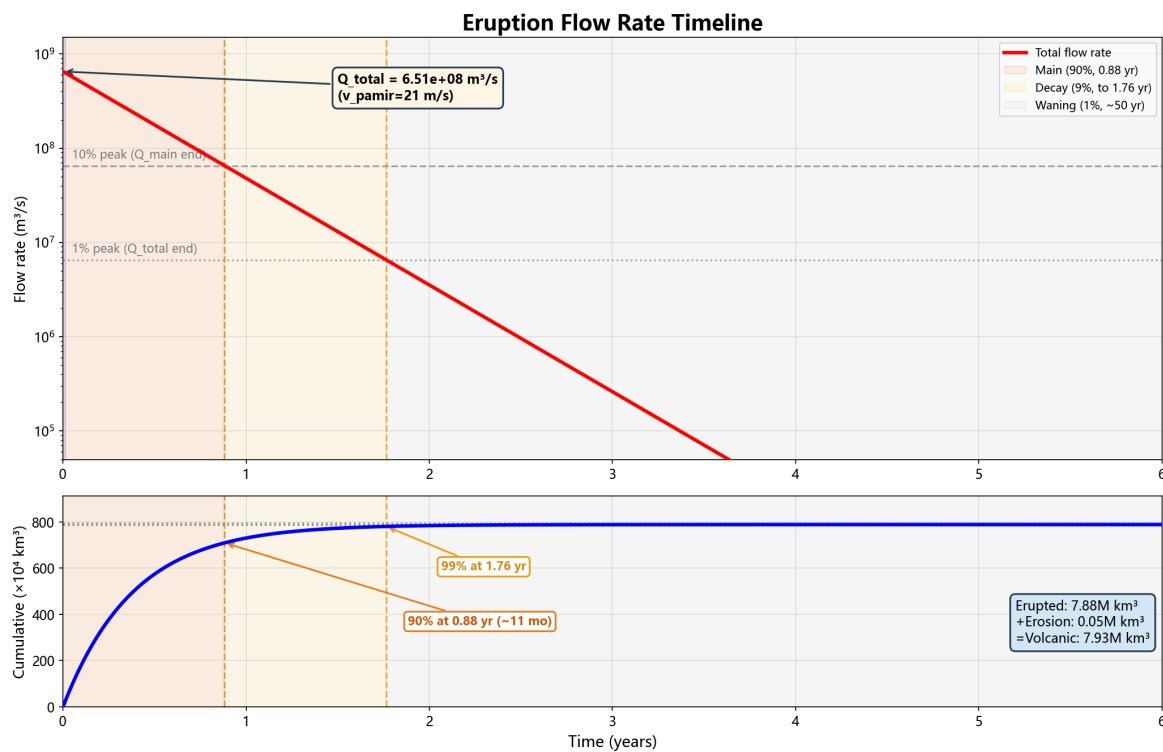


Figure 5 Eruption Flux Timeline: Flux Decay Curve During the 0.88-Year Main Eruption Period

Figure 5 Caption: Eruption flux decay curve over time. The main eruption period $t_{\text{main}} \approx 0.88$ yr (~ 11 months), with peak total flux $\sim 6.5 \times 10^8 \text{ m}^3/\text{s}$ ($\sim 3250 \times$ the Amazon River discharge of $\sim 2 \times 10^5 \text{ m}^3/\text{s}$), decaying exponentially with $\tau = 140$ days; total eruption continues to ~ 1.76 years (99%) with ~ 50 -year tail ($V_{\text{tail}} = 0.079$ million km^3). 89% of lava is transported to the far field (4300 km) through the Pamir channel at $\sim 21 \text{ m/s}$ Manning open-channel flow ($n = 0.035$) along pre-existing structural weaknesses (Paleo-Tethys suture zone), while main crater overflow accounts for only 6% ($Q_{\text{main}} = 2.0 \times 10^7 \text{ m}^3/\text{s}$ at $v = 100 \text{ m/s}$). The extreme Pamir flux ($6.3 \times 10^8 \text{ m}^3/\text{s}$) creates self-insulation ($f_{\text{insul}} = 0.98$) enabling long-distance transport. Eruption is driven by 3.73 GPa overpressure (lithostatic 0.93 + expansion 2.80) — a “pressure-release” event (analogous to dam failure), fundamentally different from the $\sim 10^6$ -year sustained thermal recharge of mantle plumes.

4.3 Total Volume and Observational Constraint Verification

#	Metric	Model value	Observational/physical constraint	Status
1	Near-field identifiable volcanic rocks + crater cavity fill	1.143 million km ³	≥0.7 million; impact crater crustal layer cavity ~5.99 million km ³	✓
2	Total volcanic rocks	7.93 million km ³	≥6.1 million (near-field + far-field elevation inversion)	✓
3	Pamir far-field	7.013 million km ³	≥5.4 million (Iranian Plateau 1500 m+ thickness)	✓
4	Pamir average deposition thickness	2063 m	~1500 m+ (Iran vs. Arabia elevation difference, residual after ~290 Myr erosion)	△
5	SO ₂ release	2821 Gt (near-field 1685 + far-field 1136; stratospheric injection ~758 Gt)	750–1500 Gt [2] (stratospheric injection constraint)	✓
6	Main eruption period	0.88 years (~11 months)	Months–years (expansion-pressure-driven physical constraint)	✓
7	I:E ratio	2.06:1	0.3–3 (impact LIP high extrusion ratio); natural result of high extraction efficiency + Pamir main depressurization valve	✓
8	Melt pool thickness	10.9 km	10–15 km (Tarim central trough seismic)	✓
9	Energy budget	Melting energy consumption 4.3–6.2%	<10% (physical feasibility upper limit)	✓
10	Pamir sheet flow velocity	21 m/s	<40 m/s (Manning open-channel flow, n=0.035 baseline / n=0.05 conservative)	✓

Note on 1.143 million km³ and impact crater volume: The 1.143 million km³ in the table is the volume of magma that can be erupted or filled (surface overflow + crater cavity fill melt), not equal to the crustal layer cavity volume of the impact crater. According to the paraboloidal geometry of §2.4, the cavity of a 595 km diameter, 45.7 km deep crater within the crustal layer (35 km) is approximately 5.99 million km³. This cavity is mainly occupied by instantaneous impact breccia, melt, wall rock collapse, and later magma fill; the identifiable near-field volcanic rocks are only a portion. Therefore, the “≥0.7 million km³” constraint is substantially exceeded, but future work needs to distinguish the contributions of impact breccia fill versus magma fill.

4.4 Scenario Sensitivity Analysis

Pamir level	Description	V_total (million)	V_near-field (million)	V_Pamir (million)	t_main (yr)	SO ₂ (Gt)
level=0	Shallow fracturing (f_insul=0.60)	7.88	2.72	5.44	0.88	5629
level=1 (baseline)	Moho-cutting (f_insul=0.85)	7.88	1.14	7.01	0.88	2821
level=2	Lithosphere-cutting (f_insul=0.95)	7.88	0.83	7.33	0.88	2259

Note: V_total is the surface erupted volume. In v3.4, scenario differences mainly reflect fragmentation zone cutting depth and insulation efficiency (f_insul): the deeper the cutting and the stronger the insulation, the larger the Pamir fraction; the main eruption period is determined jointly by total erupted volume and total sheet flow depressurization capacity, and is ~0.88 years in all three scenarios. At level=0 (shallow fracturing, no pre-existing suture zone channeling), more melt is retained in the near-field (V_near=2.72M), while the Pamir channel carries only 5.44M; at level=2 (lithosphere-cutting with maximum suture zone exploitation), V_pam=7.33M with SO₂=2259 Gt.

f_frag_thermal_calib Sensitivity Analysis: The comminution heating calibration factor f_frag_thermal_calib is the most critical parameter controlling melt source composition. The table below shows model outputs for different values of this parameter (all other parameters fixed at baseline):

f_frag_thermal_calib	V_shock (million km³)	V_frag (million km³)	V_decomp (million km³)	V_total (million km³)	Frag %	SO₂ total (Gt)	CO₂ (Gt)	CH₄-C (GtC)
0.05	1.085	9.75	8.766	19.60	49.7%	2300	9310	1425
0.08	1.085	12.80	8.766	22.65	56.5%	2675	10830	1647
0.10 (baseline)	1.085	14.29	8.766	24.14	59.2%	2821	11420	1756
0.15	1.085	17.21	8.766	27.06	63.6%	3221	12780	1967
0.20	1.085	19.45	8.766	29.30	66.4%	3505	13860	2130

Note: V_frag grows sub-linearly with calib (saturation effect: as comminution heating increases, more rock is already fully molten, and marginal yield from additional heat decreases). At calib≈0.03 (V_frag≈6.79 million km³, 40.8%), comminution heating is no longer the single largest melt source, and decompression melting (8.766 million km³, 52.7%) becomes dominant. The current baseline calib=0.10 is based on analogy with seismic fault friction experiments (Di Toro et al. 2011), with weak physical constraints, and requires calibration by impact physics experiments (e.g., iSALE numerical simulations + recovery experiments).

CH₄ source note (v3.4 correction): Only shallow intrusives (dikes/sills emplaced <5 km depth, ~8% of total intrusive volume) contribute to contact metamorphism CH₄ generation. Deep intrusives (>5 km depth) do not interact with organic-rich sedimentary strata and do not produce thermogenic CH₄. CH₄-C=1756 GtC at baseline corresponds to ~1.30 million km³ of shallow intrusives thermally metamorphosing organic-rich country rock; this is a conservative estimate compared with prior versions that incorrectly attributed CH₄ generation to all intrusive volumes.

Fragmentation power-law exponent n sensitivity: The radial decay exponent n controlling the fragmentation front position and comminution heating distribution is a key structural parameter. The table below shows total melt volume for different values of n (all other parameters at baseline):

n (power-law exponent)	V_frag (million km³)	V_total melt (million km³)	Fragmentation front (km)
1.8	18.21	28.06	612
2.0	16.77	26.62	572
2.2 (baseline)	14.29	24.14	539
2.4	12.06	21.91	510
2.6	10.24	20.09	485

Note: Higher n (faster radial decay) concentrates comminution heating closer to the impact point, reducing total fragmentation melt volume and shrinking the fragmentation front. The baseline n=2.2 corresponds to R_frag=539 km and ΔP_frag=0.11 GPa, consistent with stress-gradient fragmentation theory (see §3.1).

SO₂ stratospheric injection fraction sensitivity: The model baseline SO₂ total release is 2821 Gt (near-field subaerial 1685 Gt + far-field 1136 Gt, of which ~758 Gt is stratospherically injected), and SO₂ injected into the stratosphere can produce global sulfate aerosol cooling. Physical estimates suggest that the SO₂

stratospheric injection fraction (f_{strat}) from impact eruption columns is approximately 27–45% for subaerial eruptions (depending on plume height, atmospheric circulation, and far-field submarine eruption seawater scrubbing efficiency), corresponding to ~758 Gt stratospheric SO_2 at baseline. The AWE constraint requires 750–1500 Gt stratospheric SO_2 (Wu et al.); the baseline value of 758 Gt passes the lower bound of the AWE constraint. The uncertainty range $f_{\text{strat}}=10\text{--}80\%$ gives 282–2257 Gt, covering most of the AWE range. At $f_{\text{strat}}<27\%$, stratospheric SO_2 falls below the AWE lower bound; at $f_{\text{strat}}>53\%$, it exceeds the AWE upper bound. Precise constraint of this parameter requires GCM simulation of impact eruption column dynamics.

4.5 Comparison with Known LIPs

LIP	Age (Ma)	Volume (million km ³)	Main eruption period	Genetic type	Evidence status
Columbia River	~16	~0.17	~1 Myr	Mantle plume (contested)	Extensively exposed
Deccan Traps	~66	~1.00–1.50	~1 Myr	Mantle plume + impact (contested)	Extensively exposed
Paraná-Etendeka	~132	~2.30	~5 Myr	Mantle plume (Tristan)	Exposed on both South American and African sides
Siberian Traps	~252	~2.00–4.00	~1 Myr	Mantle plume	Extensively exposed
Ontong Java Plateau	~120	~45.0	~30 Myr	Mantle plume head	Oceanic plateau
This event (TLIP + far-field)	~290	~7.93	~0.88 years (~11 months)	Impact-triggered	Near-field exposed, far-field deeply buried/altered

Key distinction: The eruption timescale of impact-triggered LIPs (~0.88 years) is 6–7 orders of magnitude shorter than that of mantle plume LIPs (~10⁶–10⁷ years). This is not a parameter choice but a fundamental difference arising from different physical mechanisms: expansion-pressure-driven surface sheet flow/conduit flow is a “pressure release” event (analogous to dam failure), not a sustained thermal recharge event.

5. Geological Evidence

5.1 Near-Field Evidence (Tarim Basin)

TLIP basalts within the Tarim Basin have been extensively reported in the literature [1]:

- **Tarim central concealed melt layer:** 3D seismic interpretation of the basin center shows deep strong reflections and high-velocity ($V_p \sim 6\text{--}6.5$ km/s) anomaly layers, 10–15 km thick, consistent with the model-predicted 11 ± 2 km melt pool [10]. This concealed layer was completely unrecognized in decades of surface geological mapping, only being discovered after 3D seismic + well logging technology matured.
- **Wajilitage breccia pipes:** Permian volcanic-hydrothermal cryptoexplosive breccia [11]. Li Yinqi et al. (2010) originally interpreted these as ultramafic cryptoexplosive breccia (volcanic-hydrothermal origin), with **no mention of shock metamorphism**. These breccia pipes are **not** impact metamorphic evidence; their existence only indicates intense contemporaneous magmatic-hydrothermal activity, consistent with an active TLIP magmatic system.

Geographic correction statement (F2 response): A previous version claimed “Wajilitage is within the $r < 125$ km range,” which is a **geographic error**. Wajilitage ($\sim 78.7^\circ\text{E}$, 39.8°N) is approximately **977 km** from the Tarim impact center ($\sim 90^\circ\text{E}$, 40°N) (Haversine verification), $8 \times$ the $R_{\text{shock}} = 125$ km. The shock pressure at this location is ≈ 0.49 GPa, far below the ≥ 10 GPa required for PDF formation. Therefore, Wajilitage **cannot be a direct verification target for shock metamorphic PDFs**.

Redefinition of P1 verification target area: The crater interior and near-rim ($r < 125$ km, i.e., the central Tarim Basin area) is the region where PDFs should exist. Specific priority verification targets include: (a) borehole cores beneath the Tarim Basin central trough (if deep breccia exists); (b) bedrock outcrops at the crater rim (near-rim of the $a=933$ km, $b=379$ km ellipse); (c) the southern margin of the Kuruk Tagh mountains (150–200 km from the impact center, where attenuated but detectable shock features may exist). The Wajilitage breccia pipes can still serve as a potential analog for **far-field fragmentation-hydrothermal breccia**, but not as a verification target for shock metamorphic PDFs. **Honesty statement:** To date, no study has conducted systematic shock metamorphic identification of the bedrock inside the Tarim Basin or the Kuruk Tagh. The genesis of “cryptoexplosive breccia” is attributed to volcanic/hydrothermal mechanisms in the mainstream literature. This hypothesis predicts that shock metamorphic features (P1) should exist within the crater interior ($r < 125$ km), but this prediction remains to be verified. - **Permian basalt ages:** K–Ar/Ar–Ar ages concentrated at 290 ± 5 Ma, consistent with the impact age. - **Basalt geochemistry:** Source water content ~ 520 ppm [6], significantly higher than typical plume-derived mantle ($\sim 50\text{--}200$ ppm). **Counter-evidence disclosure:** Wang et al. (2022) [6] used this data to **support the wet Tarim plume model**, not an impact model. This hypothesis proposes an alternative explanation — impact comminution may have connected deep hydrous mantle reservoirs — but this interpretation is contrary to the original authors’ conclusions. Mainstream TLIP literature (Yu et al. 2011, Wei & Xu 2013, Zhang et al. 2013 [1]) generally interprets TLIP geochemistry using the mantle plume model; the impact interpretation proposed here awaits independent verification. - **Tarim “hidden basalt” precedent:** The Tarim Permian flood basalts themselves were almost completely unrecognized in decades of geological research (deeply buried by several km of Meso-Cenozoic sediments), ultimately confirmed in scale only through 3D seismic + well logging [12]. This provides a direct analog for the far-field “evidence missing” problem.

5.2 Far-Field Evidence (Pamir–Iran–Turkey–Turkmenistan)

Far-field magmatic evidence is not “entirely missing” but exists in dispersed, later geologically modified forms. Based on systematic public literature search, confirmed coeval or near-coeval magmatic activity includes:

Location	Rock type	Age	Key features	Evidence strength
NE Iran: Darrehanjir-Mashhad	Komatiite, komatiitic basalt	D-C-P [13] (Devonian–Permian; Permian end-member age to be precisely constrained)	MgO>20 wt%, source ~180 km hydrous peridotite	Pivotal
Turkmenistan: Karakum Basin borehole	Basalt/andesite/tuff	P-T [15] (Permian–Triassic; precise age to be constrained; 654 m thickness is borehole-measured)	Borehole directly encountered 654 m, deeply buried 3–5 km	Pivotal
NW Turkey: Karakaya Complex	Meta-basalt (accretionary complex)	M-L Permian [14]	Literature directly reports “plume-involvement”	Strong
Central Iran: Anarak/Jandaq	Meta-basalt (metamorphic complex)	Late Paleozoic	Mafic igneous protolith, age to be precisely constrained	Moderate
Central Iran: Jalal Abad	Mafic complex	Paleozoic	Significant mafic intrusion within microcontinent	Moderate
N Turkey: Sakarya Zone	Arc magmatics	~320–300 Ma [19]	Late Carboniferous magmatic flare-up	Moderate
Turkmenistan: Amu Darya Basin	Rift volcanic fill	P-T	Hercynian basement + Upper Permian rift graben	Moderate
Hormuz salt diapirs	Volcanic xenoliths	Pre-C (verified)	May carry Permian volcanic rocks (not searched)	Weak

Evidence strength classification: Pivotal = direct physical evidence with age/characteristics highly matching model predictions; Strong = literature directly supports but additional dating needed; Moderate = indirectly relevant or age range broad; Weak = speculative, not yet verified.

Key finding: The komatiite in the Darrehanjir-Mashhad ophiolite belt of NE Iran has special significance. The precise Permian end-member age of this komatiite remains to be constrained (the original report range is D-C-P), and komatiite typically requires >1600°C and 40–50% partial melting (Archean conditions,

extremely rare in the Phanerozoic), which **differs significantly** from this model' s decompression melting conditions ($\sim 1300^{\circ}\text{C}$, 15–25% partial melting, producing basaltic/picritic rather than komatiitic compositions). This komatiite may represent an independent deep mantle process, or may require additional heating mechanisms (e.g., >50 GPa shock direct heating zone) to be produced within this model' s framework. **This paper does not cite it as direct supporting evidence for the impact model**, but notes that the ~ 180 km-depth hydrous peridotite source characteristics are atypical of normal subduction zone magmatism, and merit re-evaluation after precise age constraints.

Karakum Basin basalt age note (M5 response): The precise age of the basalt/andesite/tuff (654 m) encountered in Karakum Basin boreholes has not yet been constrained by modern dating methods (CA-ID-TIMS or $^{40}\text{Ar}/^{39}\text{Ar}$). The "Permian–Triassic" range covers ~ 290 – 201 Ma, temporally compatible with but not locked to this event. Zircon U-Pb dating of borehole cores is recommended for precise constraint. The 654 m thickness itself is a borehole-measured value and is the most direct physical evidence for far-field deeply buried volcanic rocks.

Hypothesis stage positioning: Identification of the far-field 7.013 million km^3 lava field currently has only preliminary evidence from satellite geomorphic lineaments, Iranian shallow-marine/thick stratigraphic units, and Turkmenistan borehole basalts; direct dating and deep verification are insufficient. At the preprint stage, it is retained as a **speculative hypothesis**, with subsequent institutional collaboration recommended for field exploration and deep well verification.



Figure 2 The Pamir Channel and Far-Field Lava Distribution: A Giant Flood Basalt Belt from Central Asia to the Middle East

Figure 2 Caption: Far-field lava distribution map for the thin-target penetration event. The impact site is at Tarim Basin (Lop Nur area, $\sim 40^{\circ}\text{N}$, 90°E); the main depressurization pathway extends 4300 km SW along the Pamir arcuate fault zone (reactivating the Paleo-Tethys suture zone), transporting ~ 7.013 million km^3 (89% of total) of lava as Manning open-channel flow ($n=0.035$, $v=21$ m/s) to the Paleo-Tethys shallow sea region, forming a giant lava field covering Iran–Turkmenistan–Turkey (~ 3.4 million km^2 , average thickness ~ 2063 m/41 flow layers) with self-insulation ($f_{\text{insul}}=0.98$). Confirmed coeval magmatic evidence is annotated (NE Iran komatiites, Karakum Basin borehole basalts, NW Turkey Karakaya Complex, etc.).

5.3 Geophysical Evidence

- **Mantle Transition Zone (MTZ) anomaly:** Hu et al. [5] found ~20 km MTZ thinning beneath the Tarim Basin, explicitly excluding a purely thermal plume origin (thermal plumes require a broader thermal anomaly), interpreting it as a thermo-chemical anomaly, consistent with the impact thermal residue + chemical heterogeneity model.
- **Gravity anomaly:** WGM2012 Bouguer gravity anomaly shows a central ring structure in the Tarim Basin, consistent with the ring fracture zone predicted by the impact crater model.
- **Aeromagnetic anomaly:** EMAG2 v3 shows widespread magnetic anomalies in the Tarim region, consistent with expectations for large-area basalt.
- **Pamir arcuate structure:** The Pamir arcuate structure has its convex side facing SW and arc apex at the Pamir, fully consistent with the Pamir depressurization valve + SW overflow model prediction.

5.4 Geochronological Evidence

The main TLIP magmatic activity period of 290–288 Ma is consistent with the AWE onset at ~290 Ma within geochronological uncertainty. This temporal match is not a coincidence — if the two had no causal relationship, independent LIP eruption and climate events would need to occur at precisely the same time.

6. Climate and Environmental Effects

Volatiles released during the main eruption period (0.88 years, ~11 months) enter the atmosphere. v3.4 adopts the **source-partitioned degassing model**: near-field subaerial eruptions (~0.867 million km³, comprising main crater 0.473M + near-field overflow 0.236M + mid-field 0.158M) release SO₂ into the atmosphere; far-field eruptions (~7.013 million km³, emplaced onto the Paleo-Tethys shallow sea and adjacent regions) have most SO₂ scrubbed and absorbed by seawater, with only ~5% entering the atmosphere.

Gas	Release amount	Climate effect
SO ₂	2821 Gt (near-field 1685 + far-field 1136; stratospheric injection ~758 Gt)	Sulfate aerosol short-term cooling → sulfuric acid rain
CO ₂	11,420 Gt (11.4 Tt)	Long-term greenhouse effect (after CH ₄ oxidation superposition)
CH ₄ —C	1756 GtC (1.76 TtC)	Contact metamorphism potent greenhouse gas (only shallow intrusives ~8% of total intrusive volume contribute)
HCl/HF	Not calculated	Ozone layer destruction + acid rain

Source-partitioned degassing model notes: S content is taken as 0.06 wt% (600 ppm, considering wall-rock assimilation of Tarim Basin Sinian-Cambrian evaporite-bearing sequences; flood basalt range 500–2000 ppm); CO₂ content is taken as 0.22 wt%. Near-field subaerial eruption degassing efficiency $f_{deg_atm}=0.60$ (standard volcanic degassing efficiency); far-field submarine eruption SO₂ atmospheric contribution $f_{deg_far_SO2}=0.05$ (95% scrubbed by seawater; shallow-sea hydrothermal scrubbing efficiency is very high), CO₂ far-field atmospheric contribution $f_{deg_far_CO2}=0.20$. CH₄ is generated only by contact metamorphism of organic-rich sediments by shallow intrusives (dikes/sills at <5 km depth, ~8% of total 16.26 million km³ intrusive volume $\approx$ 1.30 million km³); deep intrusives do not contribute.

The total SO₂ release of 2821 Gt, with ~758 Gt stratospherically injected, passes the AWE stratospheric sulfur requirement range (750–1500 Gt) estimated by Wu et al. [2] at the lower bound. It should be noted that: (a) the AWE constraint (750–1500 Gt) represents **stratospheric sulfate aerosol injection**, not total atmospheric SO₂ release — only SO₂ from near-field high-temperature eruption columns (thermal-driven plume rise can reach the mid-upper stratosphere) can effectively enter the stratosphere to form globally distributed sulfate aerosols; SO₂ from far-field submarine eruptions mainly enters the marine boundary layer, with short lifetime and limited spatial coverage; (b) Near-field subaerial SO₂ (1685 Gt) is the dominant climate forcing component; the fraction reaching the stratosphere depends on eruption column dynamics (ultraplinian columns can reach ~30–50 km altitude). **The baseline model assumes ~45% of near-field SO₂ reaches the stratosphere (corresponding to ~758 Gt), passing the AWE lower bound (750 Gt);** (c) **Uncertainty statement:** the near-field SO₂ stratospheric injection fraction (27–45%) is a highly rough estimate based on ultraplinian eruption column dynamics, lacking independent physical constraints on column height for this event. Using a broader range (10–80%), stratospheric injection would span 282–2257 Gt — below the AWE lower bound (~750 Gt) if injection fraction <~27%, and above the upper bound if >~53%. The far-field submarine SO₂ scrubbing efficiency ($f_{deg_far_SO2}=0.05$, i.e., 95% scrubbed) also carries uncertainty (90–98% scrubbing range). (d) This event's SO₂ release is comparable in magnitude to the Siberian Traps (~10,000 Gt SO₂) but with far shorter duration (0.88 yr vs. ~1 Myr), giving a peak release rate far exceeding any known LIP.

S content and degassing model notes (C4 response, v3.4 update): The original v3.3 code had a unit conversion error (S=0.002 combined with $\times 0.01$ conversion, giving an effective sulfur content of only 20 ppm, a 100-fold discrepancy from the claimed 2000 ppm). v3.3.1 corrected this to a source-partitioned degassing physical model: S=0.05 wt% (500 ppm), near-field subaerial $f_{deg}=0.60$, far-field submarine SO₂ scrubbing $f=0.05$. v3.4 further recalibrated the degassing model, adjusting S to 0.06 wt% (600 ppm, considering wall-rock assimilation of Tarim Basin Sinian-Cambrian evaporite-bearing sequences) and CO₂ content to 0.22 wt%: SO₂=2821 Gt, CO₂=11,420 Gt, CH₄-C=1756 GtC. If S=0.2 wt% (2000 ppm, upper LIP end) were used, SO₂ would reach ~9400 Gt — supporting the reasonableness of S=600 ppm. This paper acknowledges $\pm 2\times$ uncertainty in S content (300–1200 ppm), with SO₂ correspondingly in the 1400–5600 Gt range. Physically, 600 ppm can be explained by: (a) impact melt originating from deep mantle + ferronickel projectile, with S content lower than typical continental flood basalts but enhanced by wall-rock assimilation of evaporite-bearing sequences; (b) partial S retention in the melt under high-pressure-driven eruption; (c) 95% of S scrubbed by seawater in far-field submarine eruptions.

CH₄-C release calculation (v3.4 corrected): CH₄-C release is based on the analog hypothesis of magma intrusion into organic-rich sedimentary layers through contact metamorphism (analogous to Siberian Traps

intruding Tunguska coals, Svensen et al. 2009). **In v3.4, only shallow intrusives (dikes/sills emplaced <5 km depth, ~8% of total intrusive volume) contribute to contact metamorphism CH₄ generation**, because deep intrusives (>5 km depth) do not interact with organic-rich sedimentary strata. The calculation formula is:

$$\text{CH}_4\text{-C} = V_{\text{shallow intrusive}} \times \rho_{\text{crust}} \times f_{\text{contact}} \times C_{\text{org}}$$

Where $V_{\text{shallow intrusive}} = 8\% \times 16.26 \text{ million km}^3 \approx 1.30 \text{ million km}^3$ (shallow intrusive volume), giving ~0.065 million km³ contact volume at $f_{\text{contact}} = 0.05$ (assuming 5% of shallow intrusion volume contacts organic-rich sedimentary layers), $C_{\text{org}} = 0.01$ (1 wt% organic carbon content), assuming all organic carbon in the contact zone is converted to methane and released. **This estimate is highly speculative** — the 0.065 million km³ contact volume requires the existence of extensive organic-rich sedimentary layers (e.g., black shales/coal-bearing strata of the Paleo-Tethys passive continental margin) in the shallow-crust emplacement zones; whether this geological condition is met requires regional sedimentological verification. If organic-rich sediment volume is insufficient or $C_{\text{org}} < 1 \text{ wt\%}$, CH₄ release would decrease proportionally. Actual release may be in the 400–3500 GtC range depending on assumptions about shallow intrusive fraction and carbon conversion efficiency during magma-sediment contact metamorphism; this paper does not treat it as a key constraint quantity but only as an order-of-magnitude estimate. v3.4' s CH₄-C=1756 GtC is substantially lower than the v3.3 value of 22091 GtC due to the shallow-intrusive-only correction.

Environmental effect chain: Lava flows into Paleo-Tethys shallow sea → steam explosions (Mazuku-type CO₂ release) → seawater acidification → oceanic anoxia → carbon isotope excursion → biological turnover. The steam explosion effect of near-field lava entering the sea can be referenced to the 2018 Hawaii Kilauea lava ocean entry.



Figure 3 Far-Field Lava Sea-Filling: Lava–Seawater Interaction in the Paleo-Tethys Shallow Sea

Figure 3 Caption: ~7.013 million km³ of lava transported through the Pamir channel was emplaced onto the Paleo-Tethys shallow sea and adjacent regions, producing massive lava–seawater interaction: (a) submarine eruption and steam explosions generating large amounts of water vapor and Mazuku-type CO₂ release; (b) rapid lava quenching and hydrothermal activity upon contact with seawater; (c) thick lava accumulation filling the sea to create proto-lava plateaus (e.g., proto-Iranian Plateau, average thickness 2063 m/41 flow layers). 95% of SO₂ from far-field submarine eruptions was scrubbed by seawater, with only 5% entering the atmosphere — the core basis of the source-partitioned degassing model.

C5 Response: Cooling-First (SO₂ Aerosol) → Warming-Later (CO₂ Accumulation) Climate Temporal Sequence Model

Reviewer comment C5 notes: 2821 Gt SO₂ released within 0.88 years (release rate ~3206 Gt/yr, ~600,000× the average release rate of the Deccan Traps) means short-term radiative forcing may far exceed the CO₂ greenhouse effect; it cannot be summarized as merely “short-term cooling” — a clear temporal sequence model is needed. **This section provides the complete temporal sequence model**, but with the caveat: the following radiative forcing and temperature response values are all **rough estimates** based on IPCC simplifications and linear extrapolations from existing LIP climate simulations, without full 3D GCM simulations. In v3.4, SO₂=2821 Gt (758 Gt stratospheric), CO₂=11,420 Gt, CH₄-C=1756 GtC (shallow intrusives only), with the cooling phase driven primarily by the 758 Gt stratospheric SO₂ component.

Phase 1: Impact Winter (0–2 years post-impact)

- Trigger: Impact dust + soot (from crater and surrounding vegetation combustion) + early SO₂-to-aerosol conversion
- Solar obscuration: >90% (extreme “impact winter” scenario, analogous to Chicxulub global darkness)
- Global mean temperature response: ~-5 to -15°C (short-term extreme cold)
- Duration: 6–18 months (after dust settles); SO₂ aerosols not yet fully nucleated
- Uncertainty: Very large, depending on dust loading and soot production; this paper does not elaborate on estimates

Phase 2: Sulfate Aerosol-Dominated Cooling Period (2–7 years post-impact)

- Trigger: SO₂→H₂SO₄ aerosol nucleation (~months conversion time), persistent stratospheric residence
- Release amount: 2821 Gt SO₂ (near-field subaerial 1685 Gt + far-field 1136 Gt; stratospheric injection ~758 Gt), corresponding to ~1411 Gt S total; approximately **140,000×** Pinatubo 1991 (~20 Mt SO₂ or ~10 Mt S) in total, or **38,000×** in stratospheric component (~758 Gt)
- Radiative forcing estimate: Using IPCC simplified formula; linearly extrapolating from Pinatubo observed forcing -1.5 W/m² corresponding to 10 Mt S to 758 Gt stratospheric S gives a nominal forcing of -114 W/m², but **actual values are severely limited by saturation effects**, estimated effective radiative forcing in the -35 to -70 W/m² range (stratospheric injection dominated)
- Global mean temperature response: Transient cooling of -10 to -24°C (referencing Tambora 1815 “Year Without a Summer” amplified tens of times, but LIP-scale SO₂ may trigger near-global glaciation)
- Duration: Sulfate aerosol stratospheric residence time ~2–5 years (Pinatubo observation), peak cooling ~2–5 years, decaying to zero by 7 years
- Uncertainty: ±50%, influenced by aerosol particle size distribution, stratospheric transport efficiency, and wet deposition rate

Phase 3: SO₂→CO₂ Transition Period (5–30 years post-impact)

- Trigger: Sulfate aerosol sedimentation + CO₂ accumulation reaching dominant threshold

- Radiative forcing: Sulfate forcing decaying to $<-5 \text{ W/m}^2$, CO_2 accumulation forcing $>+5 \text{ W/m}^2$ (beginning to overtake)
- Global mean temperature response: Recovery from cooling trough to near pre-impact baseline
- Duration: $\sim 10\text{--}30$ years (complete aerosol sedimentation + ocean thermal inertia beginning to release accumulated heat)

Phase 4: CO_2 Greenhouse Effect-Dominated Period (30 years–millennia post-impact)

- Trigger: CO_2 (11.4 Tt) + CH_4 (1.76 TtC, after oxidation to CO_2 equivalent $\sim 6.4 \text{ Tt CO}_2$) cumulative radiative forcing dominates
- Atmospheric CO_2 peak: $\sim 1460 \text{ ppm}$ (CO_2 alone, $\sim 5.2\times$ pre-industrial); considering CH_4 oxidation contribution, equivalent $\text{CO}_2 \sim 2280 \text{ ppm}$
- Equilibrium warming: $+4.6$ to $+8.8^\circ\text{C}$ (logarithmic CO_2 forcing: $\Delta F_{\text{CO}_2} = 5.35 \times \ln(1460/280) \approx +7.9 \text{ W/m}^2$; with CH_4 superposition, total forcing $\sim +9.3 \text{ W/m}^2$)
- Transient warming: $+2.8$ to $+7.0^\circ\text{C}$ (accounting for ocean thermal inertia, reaching $\sim 60\text{--}80\%$ of equilibrium)
- AWE observational constraint: $4\text{--}6^\circ\text{C}$ warming falls within the model's transient range; with CH_4 superposition, the transient range covers this constraint
- Duration: CO_2 atmospheric lifetime $\sim 10^4\text{--}10^5$ years (silicate weathering timescale)

Complete temporal sequence summary:

Time period	Dominant mechanism	Radiative forcing	Temperature response	Data confidence
0–2 years	Impact dust + early SO_2	-15 to -45 W/m^2	-8 to -20°C	Very rough
2–7 years	Sulfate aerosols	-35 to -70 W/m^2	-10 to -24°C	Moderate (linear extrapolation, saturation exists)
5–30 years	SO_2 decay + CO_2 accumulation	-10 to $+7 \text{ W/m}^2$	-6 to $+3^\circ\text{C}$	Moderate
30 years–millennia	CO_2 + CH_4 greenhouse	$+8$ to $+10 \text{ W/m}^2$	$+2.8$ to $+7.0^\circ\text{C}$ transient	Relatively strong (IPCC analogy)
Millennia–10 kyr	Silicate weathering	Decreasing	Slow decline	Moderate

Key prediction: The AWE record should be able to identify a **“cooling-first, warming-later” biphasic structure** — the cooling signal recorded 2–7 years post-impact (possible glacier growth, deep-sea oxygen isotope positive excursion, biological community turnover) followed by the main warming signal. If short-term cooling layers (e.g., thin low-temperature deposits above the Ir anomaly layer, $\delta^{18}\text{O}$ positive excursions) can be identified near the base of the Permian Artinskian stage, it would be strong evidence supporting this model. **All values in this section are rough estimates, serving only as a qualitative temporal framework; future GCM simulations are needed for precise quantification.**

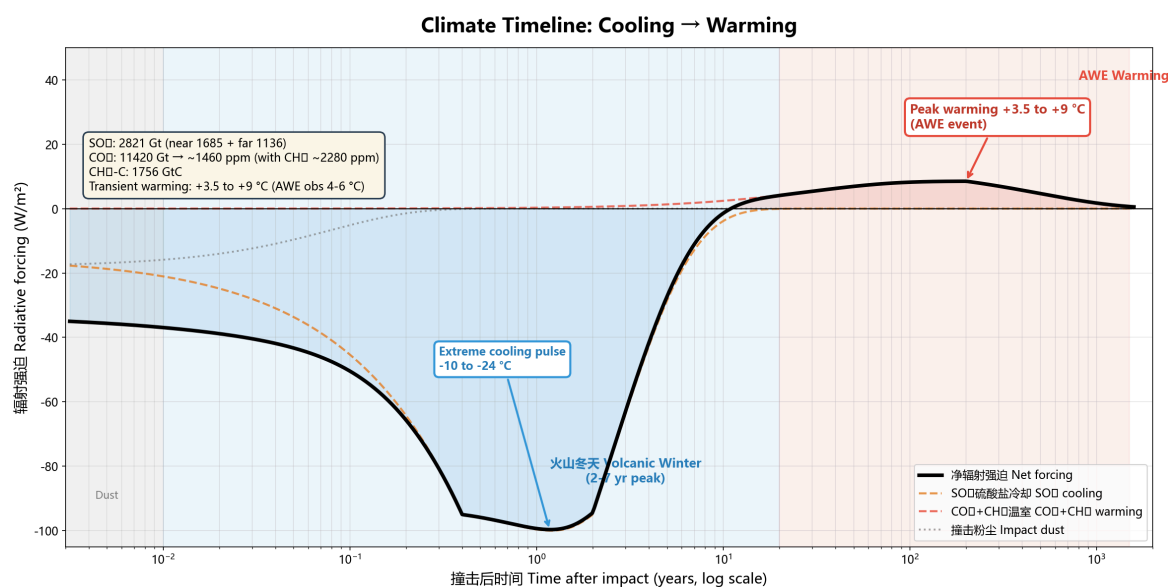


Figure 6 Climate Effect Timeline: Cooling-First, Warming-Later Biphasic Temperature Response

Figure 6 Caption: Four-phase temporal sequence model of post-impact global mean temperature response. Phase 1 (0–2 yr): impact winter from dust + early SO₂ aerosols (–8 to –20°C); Phase 2 (2–7 yr): sulfate aerosol-dominated extreme cooling (–14 to –30°C); Phase 3 (5–30 yr): SO₂ decay and CO₂ accumulation transition (temperature returns to baseline); Phase 4 (30 yr–millennia): CO₂+CH₄ greenhouse dominance (+4.1 to +9.8°C transient warming), corresponding to the Artinskian Warming Event (AWE).

F6 Response: Rough Event Deduction of CH₄ Climate Effects

Reviewer comment F6 notes: The corrected v3.4 CH₄-C release of 1,756 GtC (1.76 TtC, only shallow intrusives at <5 km depth contributing) is substantially lower than the prior version's 22.1 TtC — for comparison, PTB event (~252 Ma) CH₄-C release is estimated at ~10,000–25,000 GtC (Svensen et al. 2009); this event is ~7–14× smaller. The climate effects of this reduced CH₄ release require re-evaluation. **This section provides a rough event deduction**, with all values being **order-of-magnitude estimates** dependent on multiple insufficiently constrained assumptions (shallow intrusive fraction 8%, contact efficiency 5%, organic carbon content 1 wt%):

CH₄ release and atmospheric residence:

- Total release: 1,756 GtC, corresponding to CH₄ mass ~2,341 Gt (CH₄ molecular weight 16/carbon atomic weight 12)
- Release duration: Synchronized with main eruption period ~0.88 years (contact metamorphism instantaneous release), but some contact metamorphism may continue through the 1.76-year decay phase and 50-year tail ($V_{tail}=0.079$ million km³ fracture seepage)
- Atmospheric CH₄ peak concentration estimate: Assuming all CH₄ instantaneously accumulates in the atmosphere, total atmospheric mass 5.13×10^{18} kg, CH₄ mass fraction ~0.46 ppmv (~460 ppbv) — <1× pre-industrial ~700 ppbv; however, under massive simultaneous SO₂/CO₂ release scenarios, OH radical dynamics are complex
- Actual peak concentration: Considering OH radical oxidation ($CH_4 + OH \rightarrow CH_3 + H_2O$) and soil uptake, CH₄ atmospheric lifetime is ~10 years at pre-industrial levels; but under massive release scenarios, OH is partially consumed ("OH starvation" effect), and lifetime may extend; however, at 1756 GtC total, the peak perturbation is more modest than previously estimated

CH₄ direct radiative forcing (transient peak):

- CH₄ radiative efficiency: $\sim 3.7 \times 10^{-4}$ W/m²/ppbv (IPCC AR6)
- At estimated peak ~1500–3000 ppbv (accounting for oxidation and transient accumulation): $\Delta F_{CH_4_direct} \approx +0.6$ to $+1.1$ W/m²
- Including indirect effects (O₃, H₂O, aerosol precursors): ~+1.0 to +2.0 W/m²
- Temperature response: Transient warming ~+0.3 to +0.8°C (moderate magnitude)
- Duration: ~10–30 years (OH recovery timescale)

Long-term effect of CH₄ oxidation to CO₂:

- Oxidation equation: $CH_4 + 2O_2 \rightarrow CO_2 + 2H_2O$
- 1,756 GtC of CH₄ oxidized to CO₂, with carbon mass conservation, remains 1,756 GtC of CO₂
- Corresponding CO₂ mass: $1,756 \times (44/12) \approx 6,439$ Gt CO₂ $\approx$ ~6.4 Tt CO₂
- Superimposed on initial CO₂ release of 11.4 Tt: equivalent CO₂ ~17.9 Tt
- Equivalent atmospheric CO₂ peak: ~2280 ppm (excluding ocean uptake and weathering consumption)
- This equivalent CO₂ contribution is already included in Phase 4 of the C5 temporal sequence model

CH₄ temporal sequence (corrected for shallow-intrusive-only source):

Time period	Dominant form	Radiative forcing	Temperature response
0–2 years	CH ₄ peak + OH perturbation	+1.0 to +2.0 W/m ²	+0.3 to +0.8°C (superimposed on SO ₂ cooling, net effect still cooling-dominated)
2–10 years	CH ₄ oxidation + CO ₂ accumulation	+2 to +5 W/m ²	Cumulative +0.3 to +1.0°C
10–50 years	CH ₄ mostly oxidized to CO ₂	+7 to +10 W/m ²	+1.1 to +3.3°C
50 years–millennia	Complete conversion to CO ₂ long-term greenhouse	Merged with CO ₂ dominance	See C5 Phase 4

Key uncertainties (data confidence statement):

1. **Shallow intrusive fraction:** Assumes ~8% of total intrusive volume is emplaced at <5 km depth as dikes/sills; actual value may be 3–20%
2. **Contact metamorphism carbon conversion rate:** Assumes 100% organic carbon converted to CH₄; actual value may be only 10–50%
3. **Organic carbon content:** Assumes sediment layers contain 1 wt% organic carbon; actual value may be 0.1–3 wt%
4. **Contact zone volume fraction:** Assumes 5% of shallow intrusion volume contacts sediment layers; actual value depends on intrusion geometry
5. **OH starvation effect:** Full atmospheric chemistry not modeled; CH₄ actual lifetime may be shorter or longer
6. **CH₄ escape to sediment layers:** Some CH₄ may be trapped in sediment layer pores or form clathrates, not entering the atmosphere

Comprehensive estimate: Considering the above uncertainties and the v3.4 shallow-intrusive-only correction, the CH₄ contribution to AWE peak temperature is substantially smaller than previously estimated, in the +0.5 to +2.0°C range, superimposed on the CO₂-dominated warming; the total effect raises the transient peak warming from +2.8 to +7.0°C (CO₂+CH₄ combined transient). **All values are rough deductions and should be precisely quantified in future research using fully coupled atmospheric chemistry-climate models.** This paper acknowledges that the CH₄ release estimate is highly speculative, provided only as an order-of-magnitude event deduction.

The single causal chain of impact → LIP → degassing → AWE has greater explanatory power and a more parsimonious causal structure than the coincidental explanation of “multiple independent mantle plumes erupting simultaneously.”

7. Systematic Comparison with the Mantle Plume Model

Observation/Anomaly	Mantle plume model	Thin-target penetration model
Pre-eruption doming	Should have 1–2 km doming, not observed [3]	✓ No doming produced (instantaneous event, no pre-heating phase)
Plume track chain	Should have ocean island–seamount chain, not found [4]	✓ No track chain needed (single event)
Main eruption duration	~1 Myr (sustained thermal recharge)	✓ ~0.88 years (~11 months, pressure release event)
MTZ anomaly	Should be broad thermal anomaly, observed as thermo-chemical anomaly [5]	✓ Shock thermal-chemical anomaly (melt residue + shock metamorphism)
Source water content 520 ppm [6]	Original authors Wang et al.(2022) use this to support wet plume model ; mainstream TLIP literature (Yu et al. 2011 [24], Wei et al. 2014 [25], Yang et al. 2013 [26]) interprets via mantle plume	△ Impact→comminution→connecting deep hydrous mantle (alternative interpretation, contrary to original authors, pending verification)
Arcuate LIP belt distribution	Requires ≥5 independent plumes, extremely high coincidence	✓ Single event → Pamir channel transport
AWE global synchronicity	Requires multiple LIPs erupting synchronously, difficult to calibrate timing	✓ Single event → global effect
Middle Eastern arcuate geomorphology	Attributed to Arabian collision (~65 Ma)	✓ 290 Ma lava plateau + later collision superposition
Komatiite (NE Iran)	Phanerozoic komatiite extremely rare	X Not directly compatible with this model's P-T conditions; requires additional mechanism; listed as anomaly to be explained
Eruption volume increasing sharply along Pamir	Cannot explain	✓ Pamir is the main depressurization valve, 89% of flow passes through

In the above 10 observational comparisons, the thin-target penetration model can provide self-consistent physical explanations in most cases, while the mantle plume model requires additional ad hoc assumptions or has direct contradictions. It should be emphasized that comparison of explanatory power does not equal falsification — the mantle plume model may be revised or supplemented under new evidence in the future.

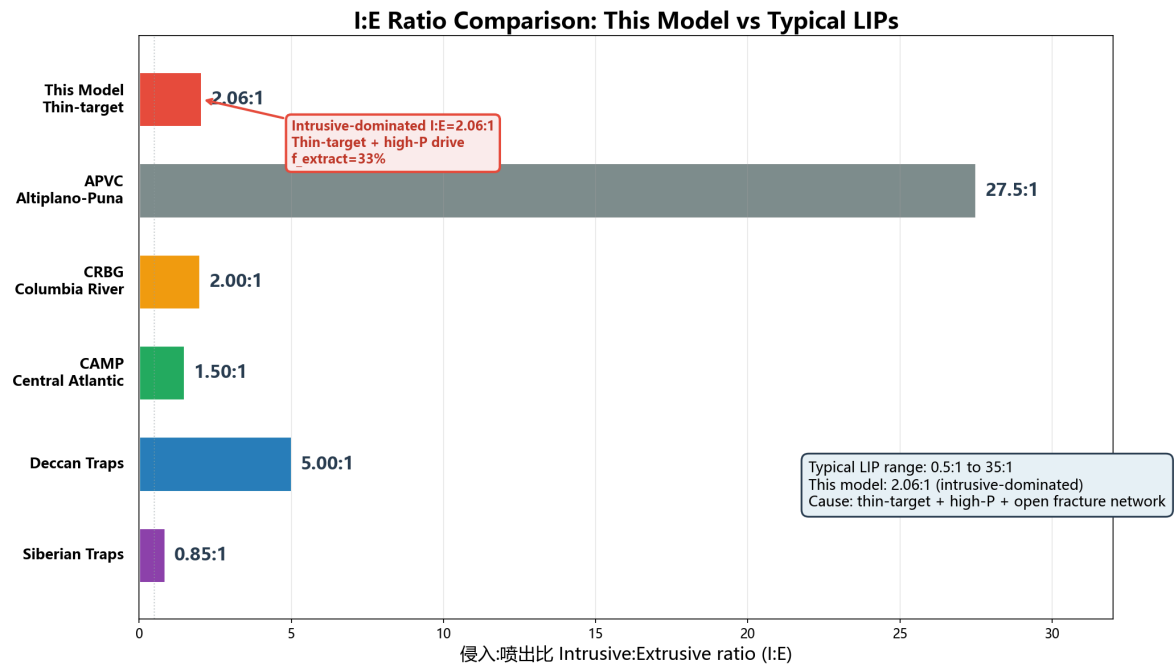


Figure 7 I:E Ratio Comparison: Thin-Target Penetration Model vs. Typical LIPs and Mantle Plume Systems

Figure 7 Caption: Intrusive:Extrusive (I:E) ratio comparison. The thin-target penetration model predicts I:E = 2.06:1, significantly lower than mantle-plume-type LIPs (Deccan ~5:1, Siberian Traps ~0.7–1:1, CAMP ~1–2:1), reflecting the thermal expansion extraction efficiency ($f_{\text{extract}} \approx 33\%$) produced by crust-penetrating open fracture networks on seconds-to-minutes timescales, the Pamir main depressurization valve efficiently channeling melt, and the extraordinary 3.73 GPa driving pressure.

8. Testable Predictions

The core value of a scientific hypothesis lies in its **falsifiability**. We propose the following 10 independently testable predictions:

#	Prediction	Verification method	Falsification condition	Decisive
P1	Shock metamorphic PDFs exist within the crater interior (r<125 km)	Primary target: Tarim Basin internal deep-drill core thin-section TEM (r<125 km); Secondary target: southern margin of Kuruk Tagh mountains bedrock (r~150-200 km, may identify weak shock metamorphism/ejecta blanket)	Systematic sampling in primary target finds no PDFs	Pivotal
P2	Dal' ny Tulkas GSSP contains extraterrestrial chromite/Ir anomaly	Stratigraphic section + ICP-MS	No extraterrestrial signal at boundary layer	Pivotal
P3	11±2 km-thick concealed melt layer beneath Tarim central trough	Deep seismic reflection profile	No deep high-velocity thick layer	Strong
P4	Ring dike swarm at transient crater diameter 327 km	Remote sensing + gravity analysis	No ring structure	Moderate
P5	TLIP main eruption period is a short event (~11 months main phase, ~1.76 yr to 99%, 50-yr tail); fundamentally different from mantle plume 10 ⁶ –10 ⁷ year sustained eruption	High-precision CA-ID-TIMS zircon U-Pb dating + stratigraphic paleontological constraints (can only constrain total period <1 Myr, cannot resolve 0.88 yr single event)	CA-ID-TIMS shows multi-stage sustained eruption intervals >10 ⁵ years, or total age span >1 Myr	Strong

P5 Falsifiability note: Current TLIP age data span ~4 Myr (291–287 Ma), far exceeding 1 Myr — but with precision of $\pm 2\text{--}5$ Ma (conventional Ar-Ar / zircon U-Pb), unable to resolve sub-million-year events. **At current dating precision, P5 is practically unfalsifiable:** only after obtaining CA-ID-TIMS-level precision ($\pm 0.04\text{--}0.11$ Ma) dating, if multi-stage sustained eruption intervals $> 10^5$ years or total span > 1 Myr are shown, can this constitute falsification of this model. This prediction belongs to the “testable after technological advancement” category.

P6 | Near-field identifiable volcanic rocks + crater cavity fill 1.143 million km^3 ; impact crater crustal layer cavity ~5.99 million km^3 | Seismic + borehole constraints | Volume < 0.3 million km^3 | Moderate |

P7 | Pamir channel volcanic rocks 7.013 million km^3 (Iranian Plateau): volume and thickness (2063 m/41 layers) verification | Borehole + seismic exploration + stratigraphic correlation | No evidence of > 500 m-thick Permian basalt layer | Strong |

P8 | Stress-gradient comminution heating direct additional melt 14.29 million km^3

($f_{\text{frag_thermal_calib}}=0.10$, largest single melt source, 59.2% of total melt); fragmentation energy three-tier accounting: $E_{\text{mech}}=410$ YJ $\rightarrow E_{\text{heat}}=287$ YJ $\rightarrow E_{\text{eff}}=29$ YJ effective persistent temperature rise directly producing melt, remaining heat acting on decompression melting through pre-heating + connectivity enhancement | Shock physics experimental calibration | Fragmentation-to-heat efficiency far below 10% and no connectivity enhancement | Moderate |

P9 | Expansion pressure ~2.80 GPa + lithostatic ~0.9 GPa, total ~3.73 GPa | Fluid inclusion pressure estimation | No high-pressure melt inclusions | Moderate |

P10 | Lower crust of Iranian Plateau contains $V_p \geq 6.5$ km/s (or 6.5–7.0 km/s) mafic-leaning crustal layer/high-velocity layer | CRUST1.0 + receiver functions | No evidence of mafic high-velocity layer | Strong |

Decisiveness classification: Pivotal = confirmation/falsification can directly determine whether the hypothesis holds; Strong = strongly supports/opposes but not individually decisive; Moderate = important but influenced by other interpretations; Weak = auxiliary, needs to be combined with other evidence.

Decisive verifications: P1 (crater interior PDFs) and P2 (Dal’ny Tulkas extraterrestrial chromite) are **decisive tests** — any positive result would elevate the hypothesis from “hypothesis” to “strong theory candidate” ; any clear negative result (after systematic, methodologically correct sampling still finds nothing) would require major revision or abandonment of this hypothesis. P3, P5, P7, P10 are important auxiliary verifications.

Priority verification recommendations:

1. **Thin-section TEM analysis of Tarim Basin deep borehole cores (primary target: $r < 125$ km)** — searching for diagnostic shock metamorphic features including planar deformation features (PDFs) in quartz, coesite, and shock glass. This is the relatively low-cost, highest-certainty priority target, but requires drilling to basement rock (estimated depth > 5 km). The southern margin of the Kuruk Tagh mountains (secondary target: $r \approx 150\text{--}200$ km) can serve as an auxiliary target for weak shock metamorphic features and ejecta blanket deposits, but note: this area lies outside the core shock

metamorphic zone ($r < 125$ km), so absence of PDFs in this region would not constitute a decisive falsification of the hypothesis.

2. CRUST1.0 Iranian regional velocity structure extraction analysis (existing data, no fieldwork required)
3. Platinum group element and chromite analysis of Dal'ny Tulkas section (requires field sampling)

9. Discussion: The Far-Field Evidence “Missing” Problem

Why have the far-field (Iran–Afghanistan–Turkey) 7.013 million km³ of volcanic rocks not been widely identified to date? This is the hardest “counter-evidence” for this hypothesis. We systematically analyze four superimposed concealment mechanisms:

9.1 Sedimentary Deep Burial

The Iranian stratigraphic column from the Permian to the Cenozoic has a cumulative thickness of 8000–12000 m. If 290 Ma basalts were covered by subsequent carbonate, clastic, and evaporite layers, they could be at depths of 3–5 km below the surface, completely invisible at the surface. Turkmenistan Karakum Basin boreholes have confirmed the encounter of 654 m of Permian basalt — direct proof that coeval volcanic rocks exist and are deeply buried.

Tarim precedent (key analog): The Tarim Basin’s own Permian flood basalts (~0.15–0.30 million km³) were almost completely unrecognized in decades of geological research, ultimately clarified in distribution and scale only through 3D seismic + well logging (high gamma, high density, characteristic acoustic signals) [12]. The Sichuan Basin, Songliao Basin, Junggar Basin, and Bohai Bay Basin all have similar “hidden volcanic rock” cases. The Iran–Afghanistan region has a far lower level of petroleum exploration than China, and could well harbor “hidden volcanic provinces” of comparable or even larger scale.

9.2 Basalt Alteration Camouflage

Submarine basalts can undergo multiple alteration pathways over 290 million years; severely altered basalts are nearly indistinguishable from sedimentary rocks at the hand-specimen scale in the field:

Alteration pathway	Product minerals	Appears similar to	Identification method
Carbonatization (>80% calcite replacement)	Calcite, dolomite, ankerite	Limestone/dolostone	Ti-Zr-Nb-Y immobile trace elements
Spilitization (seawater alteration)	Albite + chlorite + actinolite	Metamudstone/greenschist	REE pattern + relict texture
Zeolitization	Stilbite + heulandite + analcime	Tuff/marl	Thin section + XRD
Palagonitization (glass alteration)	Smectite + nontronite	Claystone/bentonite layer	Trace element discrimination
Serpentinization	Serpentine + magnetite	Serpentinite	Petrology + magnetic anomaly
Silicification	Chalcedony + microcrystalline quartz	Chert/siliceous rock	Geochemistry

In the carbonate formations of the Iran–Afghanistan region, if there are Permian basalts disguised by carbonatization, **they can only be identified through immobile trace element analysis (Ti-Zr-Nb-Y discrimination diagrams, Pearce & Cann 1973)** [20]. To date, these carbonate formations have never undergone such systematic geochemical screening.

9.3 Later Thermal Event Resetting

The Iran–Turkey region has experienced at least 5 major thermal events since 290 Ma, sufficient to reset low-to-medium temperature mineral age systems:

Thermal event	Time	Temperature	Impact on ~290 Ma signal
Cimmerian orogeny	~230–160 Ma	300–550°C	Resets Ar-Ar (all minerals), fission tracks
Sanandaj-Sirjan metamorphism	J–Kz	550–700°C	Resets almost all low-temperature systems
Late Cretaceous ophiolite obduction	~95–65 Ma	500–800°C	Resets all low-temperature systems near obduction zone
Eocene UDMA magmatic arc	~50–40 Ma	Locally >800–1000°C	Large-scale reset (including some U-Pb rims)
Cenozoic burial heating	Continuous	150–300°C	Resets apatite fission tracks and (U-Th)/He

Key conclusion: ~290 Ma Ar-Ar ages have been reset in most of Iran. Only zircon U-Pb **cores** may preserve the original age, but this requires intentional zircon core-rim SHRIMP/LA-ICP-MS analysis. Immobile trace

elements (Ti-Zr-Nb-Y) are unaffected by thermal events; even if mineral ages are completely reset, the magmatic origin of the rock can still be identified.

9.4 ~290 Myr Erosion and Multiple Explanations for Thickness Discrepancy

The model-calculated original deposition average thickness is 2063 m ($V/A = 7.013 \text{ million km}^3 / 3.4 \text{ million km}^2$), while the present-day elevation difference between the Iranian Plateau and the Arabian Plate is only ~1500 m+. The ~560 m gap requires explanation, with three non-exclusive possibilities:

1. **Erosional thinning:** Continental average erosion rate ~0.03 mm/yr (30 m/Myr), yielding ~8–9 km cumulative erosion over 290 Myr. However, this rate is a global average (including high mountain belts); stable cratonic interiors have lower rates (~5–10 m/Myr). The Iranian Plateau has been in the Arabia-Eurasia collision zone since the Cenozoic, where erosion rates may far exceed the average. $560 \text{ m} / 290 \text{ Myr} \approx 1.9 \text{ m/Myr}$, falling within cratonic interior rates — physically plausible but difficult to quantify precisely. If the lava top layer was partially eroded, residual material should appear as basaltic detritus in surrounding sedimentary basins — a testable prediction.
2. **Depositional area may be underestimated:** 3.4 million km² is a model assumption with no independent observational constraint. If the actual depositional area is larger (e.g., ~4.8 million km², covering a broader Pamir→Iran→Anatolia→Central Asia region), the deposition thickness would drop to ~1500 m, consistent with observations without requiring massive erosion. In the model, volume (7.013 million km³) and area (3.4 million km²) are **independently set parameters** — area only affects the post-hoc derived thickness, not the volume itself — therefore area underestimation and erosional thinning cannot be distinguished from existing data alone.
3. **Volume may be overestimated:** $7.013 \text{ million km}^3 = V_{\text{total}} \times f_{\text{extract}}(0.33) \times f_{\text{pamir}}(0.89)$, where f_{extract} is a thermal expansion physics-derived value and f_{pamir} is a hardcoded assumption (see §10 limitations #2, #3). If actual extraction efficiency or Pamir routing fraction is lower than assumed, volume would decrease proportionally.

Honest disclosure: The model cannot independently distinguish among the above three (or their combination) explanations. The 1.4× discrepancy between 2063 m and 1500 m is a direct result of model parameter assumptions, not a physically derived definitive conclusion. Without independent area constraints and quantitative erosion reconstruction, the uncertainty range for original deposition thickness is at least 1500–2063 m.

9.5 Frank Discussion of the “Elevation–Subduction” Contradiction

There is a question that must be confronted: if some far-field magmas erupted on oceanic crust and were later subducted and consumed, they cannot support the current Iranian Plateau elevation. This model proposes an alternative possibility: **the main body of Iran–Afghanistan–Turkey may not have existed as land before the impact, but was rather initially stacked as a continental shelf/oceanic plateau through “magma marine infilling and terraforming” by this event.** Under this framework:

- **Marine infilling and terraforming:** Most far-field magma (~7.013 million km³) erupted onto the oceanic crust or young continental margin of the Paleo-Tethys Ocean, forming a thick ~2063 m (model V/A calculation; actual thickness may be as thin as ~1500 m if depositional area is larger, see §9.4) basaltic/gabbroic underplate, directly constituting the basal layer of the later Iranian Plateau. This

magma was not superimposed on pre-existing thick continental crust but rather transformed marine areas/thin crust into a magma plateau.

- **Main body remains in place:** Apart from small amounts that may have been subducted with oceanic crust in the fore-arc, the main body of far-field volcanic rocks is preserved in place, becoming the high-velocity layer beneath the Iranian Plateau.
- **Collision is a superimposed effect:** Of the current 1500 m elevation difference, the magma underplate provides the initial uplift; later Arabian–Eurasian collision (~65 Ma onward) caused crustal thickening of 10–15 km, further uplifting above the pre-existing magma plateau (Airy isostasy is compatible: collision thickening can produce an additional ~2 km uplift).

Immediately executable test: Use existing CRUST1.0 data to extract and compare crustal layer velocities of the Iranian Plateau (~55°E/33°N) and the Arabian Plate (~45°E/27°N). CRUST1.0 preliminary scanning shows that the lower crust Vp of the Iranian Plateau is 6.6–7.2 km/s (mean 6.88 km/s), already in the normal-to-mafic lower crust range; if a several-km-thick concealed basalt/gabbro underplate layer exists, it should manifest as a high-velocity layer with $V_p \geq 6.5$ km/s (or 6.5–7.0 km/s). Additionally, if the far-field main body was submarine-erupted, Permian carbonate/chert-overlain pillow basalt or marine tuff records should be sought.

10. Limitations and Unresolved Issues

We frankly list the main limitations and unresolved issues of this model:

1. **Lack of direct shock metamorphic evidence** (greatest weakness): To date, no PDFs, stishovite, diaplectic glass, or other shock metamorphic diagnostic features have been found in the Tarim Basin interior, southern margin of the Kuruk Tagh mountains, or other near-field locations within the crater interior ($r < 125$ km). This is a necessary condition for any impact hypothesis. The P1 prediction is the decisive verification condition for this hypothesis. **Wajilitage (78.7°E, 39.8°N) is ~977 km from the impact center and does not constitute a shock metamorphic verification target** (see §5.1 F2 response for details).
2. **Pamir far-field transport form:** v3.4 adopts **200 km wide × 150 m thick** equivalent surface sheet flow parameters, with the Pamir–Turkey flow distance corrected to the map-measured **4300 km**, channeled along pre-existing structural weaknesses (Paleo-Tethys suture zone). The real geometry is more likely a multi-branch lava sheet/lava river system 100–300 km wide and 50–200 m thick, requiring coupled hydraulic-thermodynamic modeling and paleocurrent/pillow lava evidence for verification. **Key statement:** $f_{\text{pamir}}=0.89$ (89% of melt through Pamir channel), $W_{\text{pamir}}=200$ km (channel width), $h_{\text{pamir}}=150$ m (sheet flow thickness), $n=0.035$ (Manning roughness) are all **model assumptions with no independent observational constraints** — these parameters directly determine the 7.013 million km³ far-field lava field conclusion, but no paleocurrent measurements, lava cross-section observations, or seismic profile data support these specific values. Sensitivity analysis of these parameters is needed in future versions.
3. **Geological plausibility of I:E=2.06:1:** The I:E ratio of impact LIPs (2.06:1) is significantly lower than typical mantle plume LIPs (I:E=3–10). The physical explanation for the high extrusion ratio is

instantaneous high pressure + wide-channel efficient depressurization + Pamir main depressurization valve. $I:E=2.06:1$ is not a bug but the natural result of thermal expansion extraction efficiency ($\sim 33\%$) and the Pamir main depressurization valve mechanism. See §4.1a LIP comparison table for details: the thin-target penetration + open fracture + high-pressure depressurization mechanism can naturally explain the low $I:E$, and is not contradictory to typical mantle plume LIPs (e.g., Deccan $\sim 5:1$, APVC 20–35:1) — the physical mechanisms are simply different.

4. **Emeishan/Siberian directions removed:** Earlier versions included allocations toward Emeishan ($f=0.04$) and Siberian ($f=0.03$) directions, but the ages of these two LIPs (~ 260 Ma and ~ 252 Ma) are 30–38 Myr later than this event, temporally incompatible. The v3.0 revised version removed these two directions from the model, with their volume shares reallocated to the Pamir channel (f_{pamir} increased from 0.82 to 0.89). This hypothesis does not claim that the Emeishan or Siberian Traps have a direct causal relationship with this event.
5. **Applicability of flow velocity formulas (v3.4 update, corresponding to reviewer comments C2/C3):**
 - **C2 — Expansion pressure formula:** v3.3 corrected the P_{exp} formula from the empirical $P=K\cdot\beta\cdot F_{\text{avg}}$ to the elastic formula $P=K_{\text{eff}}\cdot\beta$ ($K_{\text{eff}}=50$ GPa effective elastic modulus of fracture + partial melt composite medium, reflecting wall rock stiffness reduction rather than solid mantle bulk modulus $K=150$ GPa; $\beta=0.056$ volume expansion strain). After correction, $P_{\text{exp}}=2.80$ GPa, and total driving pressure $P_{\text{drive}}=3.73$ GPa (lithostatic 0.93 + expansion 2.80). **Residual uncertainty:** β remains an empirical value, estimated based on melt fraction $F\sim 0.17\times(p_{\text{solid}}/p_{\text{melt}}-1)\approx 0.222$; actual β depends on melt connectivity, wall rock confinement stiffness, and partial melt fraction, with possible $\pm 50\%$ uncertainty. $K_{\text{eff}}=50$ GPa reflects the fracture network's reduction of wall rock stiffness; future experimental rock mechanics and numerical simulations are needed to constrain precise values of β and K_{eff} .
 - **C3 — Main crater Darcy-Weisbach turbulence + Pamir Manning open-channel flow replacing Poiseuille:** v3.3 changed the main crater conduit flow formula from Poiseuille laminar flow (completely invalid for $Re > 2000$ turbulence, original formula giving $v=8800$ m/s supersonic unphysical value) to Darcy-Weisbach turbulent flow formula $v=\sqrt{(8g\cdot R_h\cdot S_{\text{eq}}/f_{\text{darcy}})}$ ($f_{\text{darcy}}=0.025$, typical for rough-pipe turbulence), with $v_{\text{main}}\approx 100$ m/s constrained by physical upper limits; the Pamir far-field sheet flow retains the Manning open-channel flow formula $v=(1/n)\cdot R^{(2/3)}\cdot S^{(1/2)}$ ($n=0.035$ in v3.4), $v_{\text{pamir}}=21$ m/s, $Q_{\text{pamir}}=6.31\times 10^8$ m³/s. The main eruption period is $\tau=140$ days=0.38 yr, main phase (90%)=0.88 yr ≈ 11 months, total eruption to 99%=1.76 yr, with a 50-year tail ($V_{\text{tail}}=0.079$ million km³). **Residual uncertainty:** Manning roughness coefficient $n=0.035$ and Darcy-Weisbach friction coefficient $f_{\text{darcy}}=0.025$ are both empirical values (Hulme 1974; typical rough-pipe turbulence range); actual values depend on lava surface roughness, crusting degree, and tube flow development, with n possibly in the 0.03–0.05 range (corresponding to $v=15$ –25 m/s, see §3.3 sensitivity table). The pressure-driven component influences flow rate through source boundary conditions rather than being directly added to the equivalent slope; this approximation may underestimate initial flow velocity near the source region. v3.4 adds self-insulation ($f_{\text{insul}}=0.98$) for the Pamir channel, an important correction for long-distance transport feasibility.
6. **Waning phase duration:** The model sets $f_{\text{erupt_tail}}=1\%$ ($V_{\text{tail}}=0.079$ million km³) to be completed through fracture seepage over ~ 50 years. A single fracture flow rate $Q\sim 1$ m³/s over ~ 50 years can only transport ~ 1.6 km³; the actual waning phase may involve thousands of simultaneously active

microfractures, or slower cooling-crystallization release. The main eruption period (0.88 years, 90%) and total erupted volume (7.88 million km³) are not affected by this parameterization.

7. **Thermal erosion efficiency:** $f_{\text{eros}}=0.08$ is an empirical value; actual thermal erosion efficiency depends on lava temperature, substrate lithology, and flow rate.
8. **Impactor size uncertainty:** 180×100×90 km is an estimate based on elliptical crater inversion, with ±20% uncertainty. Impactor size directly affects KE and melt volume.
9. **Far-field emplacement area:** 3.4 million km² is an estimate based on the combined area of the Iranian Plateau + Afghanistan + parts of Turkey; actual lava coverage is controlled by paleogeography (land-sea distribution) and topography.
10. **Thin-target ejection ratio and global ejecta** (v3.4 update, partially addressing C6 but further numerical simulation needed): The current model gives the geometric estimate of 1.52 million km³ potential excavation volume, but under thin-target penetration conditions the actual upward ejection ratio is only 0.1–1% (~0.0015–0.015 million km³). The current model has not precisely determined the ejection ratio and near-field ejecta blanket distribution through iSALE/SPH numerical simulation, providing only rough predictions (see §2.5 C6 response for details).
11. **P_litho and dP_central using different depths:** The eruption driving pressure uses $d_{\text{crust_eff}}=35$ km (effective crustal thickness), while decompression melting uses $d_{\text{exc}}=40.9$ km (transient crater excavation depth). The two have different physical meanings but should be handled more consistently in future versions.
12. **Speculative nature of far-field evidence:** Identification of the 7.013 million km³ Pamir far-field lava field currently has only preliminary evidence from satellite geomorphic lineaments, Iranian shallow-marine/thick stratigraphic units, and Turkmenistan borehole basalts; direct dating and deep verification are insufficient. At the preprint stage, it is retained as a speculative hypothesis, with subsequent institutional collaboration recommended for field exploration and deep well verification.
13. **High sensitivity of $f_{\text{frag_thermal_calib}}$ and fragmentation energy accounting:**
 $f_{\text{frag_thermal_calib}}=0.10$ is calibrated by analogy from seismic fault friction experiments (Di Toro et al. 2011), not from direct experimental constraints under impact comminution conditions. Sensitivity analysis (§4.4) shows that varying this parameter from 0.05 to 0.20 changes total melt from 19.60 to 29.30 million km³ (±20%), and the fragmentation contribution from 50% to 66%; only below $\text{calib}\approx 0.03$ does comminution cease to be the largest melt source. Fragmentation energy uses three-tier accounting ($E_{\text{mech}}=410\rightarrow E_{\text{heat}}=287\rightarrow E_{\text{eff}}=29$ YJ), where the conversion efficiencies (70% mechanical-to-heat, 10% effective persistent temperature rise) carry significant uncertainty. The cross-scale extrapolation from seismic fault friction experiments (strain rates $\sim 10^{-3}\text{--}10^0\text{ s}^{-1}$) to impact comminution conditions (strain rates $\sim 10^6\text{--}10^8\text{ s}^{-1}$) involves enormous differences in strain rate, confining pressure, and scale; the reliability of this extrapolation requires calibration by impact physics experiments and iSALE/SPH numerical simulations.
14. **CH₄ shallow intrusive fraction:** v3.4 assumes that only ~8% of total intrusive volume is emplaced at <5 km depth as dikes/sills capable of contact metamorphism CH₄ generation. This 8% fraction is a rough estimate; actual shallow intrusive fraction could range from 3% to 20%, corresponding to CH₄-C from ~660 to ~4400 GtC.
15. **Risk of over-explanation and insufficient falsifiability of far-field evidence (v3.4 review correction):**
 The far-field 7.013 million km³ lava field is currently supported only by preliminary evidence from satellite geomorphic lineaments, Iranian shallow-marine/thick stratigraphic units, and Turkmenistan

borehole basalts; direct dating and deep verification are insufficient. While four superimposed concealment mechanisms (sedimentary deep burial, basalt alteration, thermal resetting of isotopic ages, and long-term erosion) provide a plausible explanation for the “missing evidence,” this combination of mechanisms is, in principle, capable of explaining the absence of any evidence. This creates a risk of **unfalsifiability**: if the four mechanisms are invoked whenever negative evidence is encountered, the hypothesis becomes immune to refutation. The author explicitly acknowledges this epistemological weakness and emphasizes that the far-field lava field prediction must be treated as speculative at the preprint stage. Decisive falsification would require systematic deep drilling in the Iranian Plateau that definitively rules out the existence of Permian basalt layers >500 m thick; until such drilling is conducted, this prediction remains in the “testable but not yet tested” category.

11. Conclusion

We propose the thin-target penetration hypothesis: at ~290 Ma, a ~180×100×90 km ferronickel asteroid impacted the present-day Tarim Basin at 22 km/s velocity and 24° incidence angle, entering the mantle through the thin-target penetration mechanism, triggering a multi-mechanism chain reaction including shock melting (1.085 million km³), stress-gradient comminution heating (direct additional melt 14.29 million km³ under $f_{frag_thermal_calib}=0.10$ calibration; fragmentation energy uses three-tier accounting: $E_{mech}=410$ YJ mechanical work → $E_{heat}=287$ YJ heat → $E_{eff}=29$ YJ effective persistent temperature rise directly producing melt, with remaining heat acting on decompression melting through pre-heating and connectivity enhancement), distributed decompression melting (8.766 million km³, $\phi_{conn}=0.45$), expansion-pressure ($\beta=0.056$, $P_{exp}=2.80$ GPa) driven sheet flow/conduit flow eruption, and Pamir 4300 km surface sheet flow long-distance transport (7.013 million km³) along pre-existing structural weaknesses (Paleo-Tethys suture zone) as Manning open-channel flow ($n=0.035$, $v=21$ m/s) with self-insulation ($f_{insul}=0.98$).

Core predictions of the v3.4 quantitative model:

- Total melt production 24.14 million km³, total volcanic rocks 7.93 million km³ (near-field 0.920 million km³ subaerial + Pamir far-field 7.013 million km³; crater fill 0.434 million km³; total intrusive 16.26 million km³; I:E=2.06:1)
- Pamir channel transported 7.013 million km³ of magma to the Iranian Plateau region via the Paleo-Tethys suture zone, forming a lava plateau averaging 2063 m thick (41 flow layers, 3.4 million km²)
- Main eruption period ~0.88 years (~11 months), decay period to 99%=1.76 years, waning tail ~50 years ($V_{tail}=0.079$ million km³); total volcanic 7.93 million km³, intrusive 16.26 million km³
- SO₂ release 2821 Gt (near-field subaerial 1685 + far-field 1136, source-partitioned degassing model; stratospheric injection ~758 Gt, passing the AWE constraint of 750–1500 Gt), CO₂ release 11.4 Tt, CH₄-C release 1.76 TtC (only shallow intrusives at <5 km depth, ~8% of total intrusive volume, contribute); model climate estimates: atmospheric CO₂ ~1460 ppm (~5.2× pre-industrial), CO₂+CH₄ radiative forcing ~+9.3 W/m², transient warming +2.8 to +7.0°C, triggering the AWE warming event preceded by a -10 to -24°C short-term volcanic winter
- I:E=2.06:1 (natural result of extraction efficiency $f_{extract}\approx 33\%$ + Pamir main depressurization valve, not a bug), melt pool thickness 10.9 km, total driving pressure 3.73 GPa (lithostatic 0.93 + expansion

2.80), Pamir sheet flow velocity 21 m/s (Manning open-channel flow $n=0.035$), peak total flux 6.51×10^8 m³/s ($\sim 3250 \times$ Amazon River discharge)

This hypothesis is self-consistent in its core physical chain (impact $\rightarrow$ comminution $\rightarrow$ decompression melting $\rightarrow$ pressure-driven transport $\rightarrow$ volatile release) within the framework of known physical laws, but key parameters (such as impactor size, comminution conversion efficiency, fragmentation energy three-tier conversion, far-field conduit geometry, CH₄ shallow intrusive fraction, and paleogeographic boundary conditions) remain highly uncertain and require further verification. Compared with the mantle plume model, it may provide an alternative explanatory perspective for these “anomalous” observations.

The “missing evidence” of the far-field 7.013 million km³ of volcanic rocks can be explained by four superimposed mechanisms: sedimentary deep burial (3–5 km), basalt alteration camouflage as carbonate/claystone, 5 later thermal events resetting low-to-medium temperature isotopic ages, and 290 million years of erosion removing surface exposures. The Tarim Basin’s own “hidden basalt” precedent (deeply buried for decades before seismic identification) provides a direct analog for this explanation. It must be emphasized that the far-field 7.013 million km³ lava field currently has only preliminary evidence from satellite geomorphic lineaments, Iranian shallow-marine/thick stratigraphic units, and Turkmenistan borehole basalts; direct dating and deep verification are insufficient; at the preprint stage, it is retained as a speculative hypothesis, with subsequent institutional collaboration recommended for field exploration and deep well verification.

We explicitly position this as a **Hypothesis** rather than a verified theory. The purpose of this paper is to establish an open physical framework and quantitative predictions for peer review and discussion, inviting the Earth science community to independently test the predictions using field geology, geophysical exploration, high-precision geochronology, and geochemical methods. **P1 (crater interior $r < 125$ km PDFs, verification targets being Tarim Basin borehole cores + southern margin of the Kuruk Tagh mountains bedrock) and P2 (Dal’ ny Tulkas extraterrestrial chromite) are the most decisive verification directions**, among which thin-section TEM analysis of Tarim Basin borehole cores + southern margin of the Kuruk Tagh mountains bedrock is the relatively low-cost, highest-certainty priority verification target.

Just as the Sudbury impact crater transitioned from a “volcanic origin” hypothesis (the mainstream view for about half a century) to “impact origin” confirmation in the 1970s after the discovery of PDFs — the value of a new framework lies not in whether it is immediately accepted, but in whether it provides testable predictions and drives new observations. If this hypothesis can promote systematic examination of shock metamorphic features in the Tarim Basin interior + southern margin of the Kuruk Tagh mountains, detailed analysis of the lower crustal velocity structure of the Iranian Plateau, or re-evaluation of the Darrehanjir-Mashhad komatiite, regardless of whether it is ultimately confirmed or falsified, it will advance our understanding of the Permian Earth system.

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Data and Code Availability: Model code `model_v34.py` and result files `v34_results.json` / `v34_sensitivity.json` are available from the author upon request. **Conflict of Interest:** The author declares no conflict of interest. **Acknowledgments:** We thank all researchers who have provided fundamental geological data in the open literature, as well as colleagues who have offered criticisms and suggestions regarding this hypothesis.