Planetary Surfaces:

1. What define planetary surfaces geologically?
2. What controls the evolution of planetary surfaces?
3. How do surface-shaping processes scale across planetary bodies of different sizes, compositions, and physical states?
4. How do surface-shaping processes interact? What is the role of feedback?

1. What define a planetary surface geologically?

They have the following main features:

(a) Surface morphology.

(b) Composition that may vary spatially rather than be a constant.

(c) Surface age that may also vary spatially.

(d) Surface texture.

Wrinkled surface contains three different types of age information: age of the surface, age of rock formation, and age of geologic events (deformation and erosion)

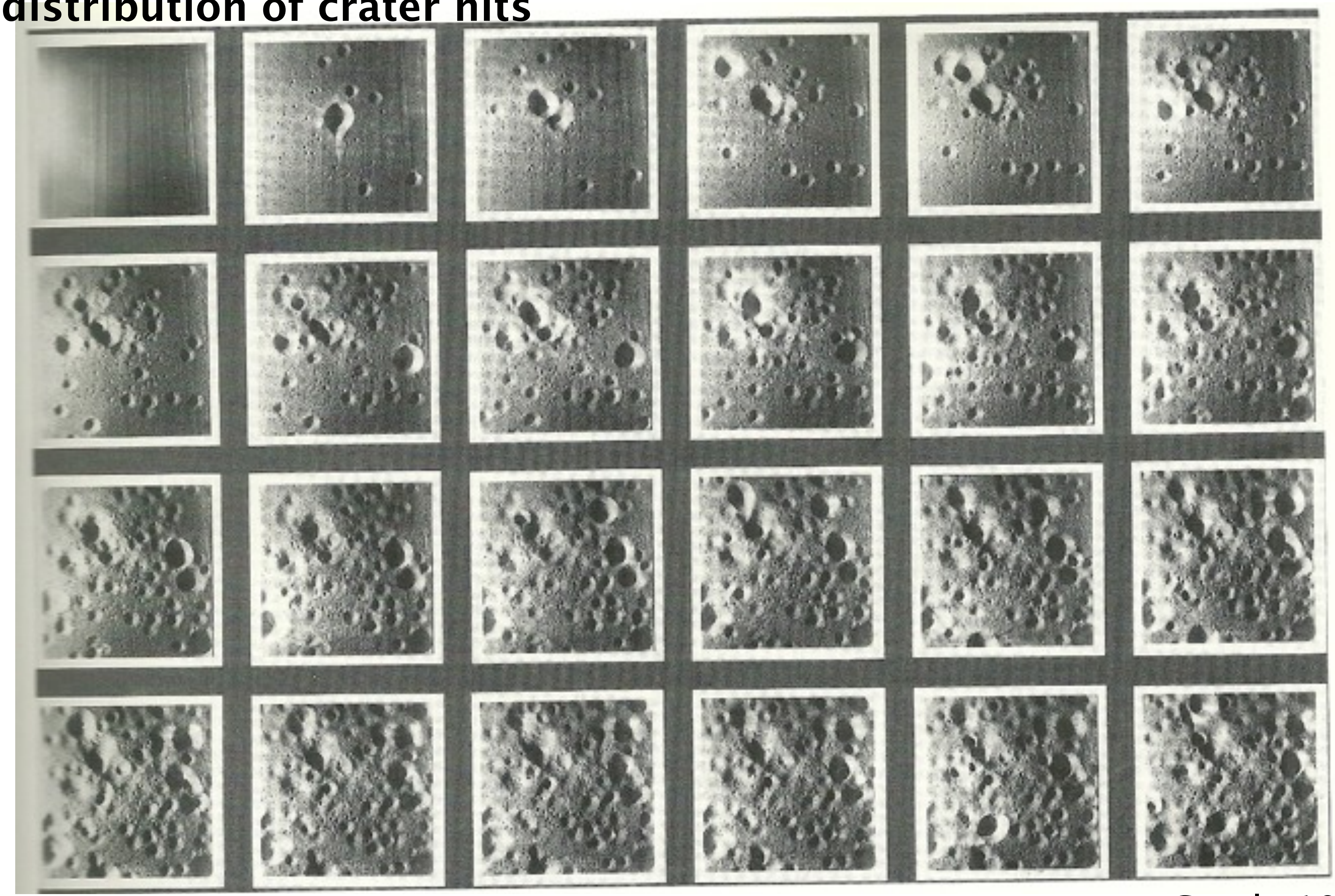
Surface morphology: wrinkle ridge

Composition: basaltic lava on Moon

Age: Assume the emplacement of lava is instantaneous, we can use a single age to represent this surface. If the surface is partially modified by episodic lave emplacement, the surface, depending on the extent that we are considering, may have variable ages.

Texture: cratered, irregular ridges

Relative surface ages using crater counting: assuming instantaneous formation of the surface and uniform distribution of crater hits



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Crater Counting Statistics in Log-Log plot

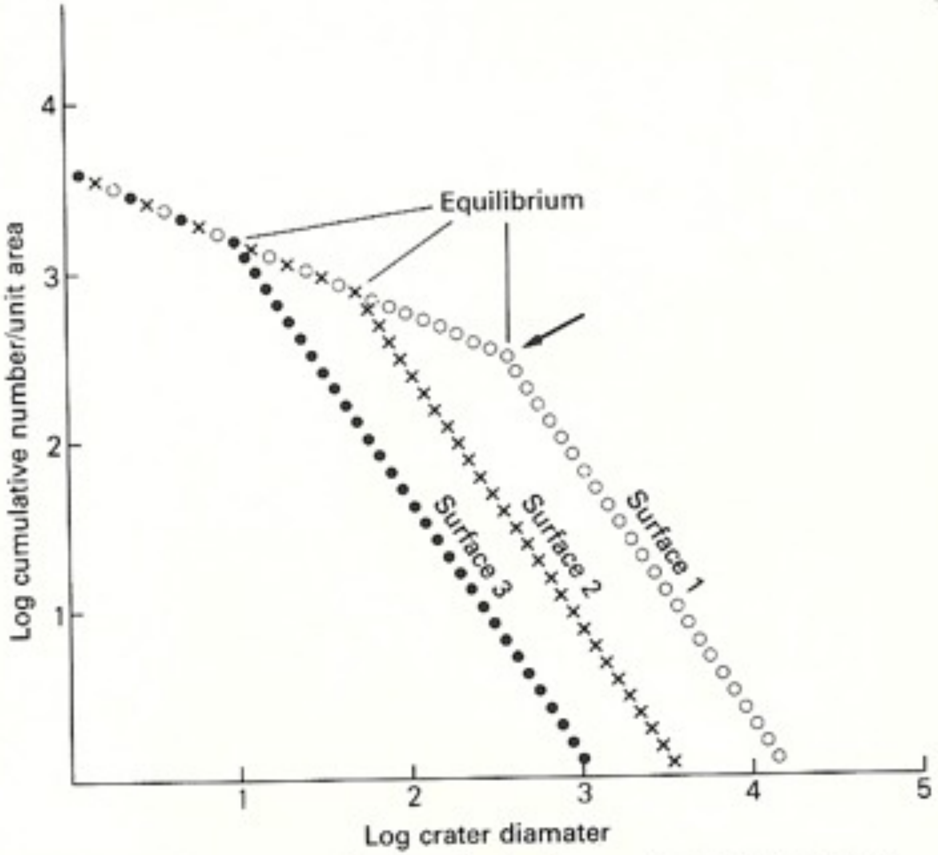

Figure 1.7(b) Stylized diagram indicating method of displaying crater statistics in which cumulative numbers of craters in given size ranges are plotted per unit surface area. "Break" in slope (arrow) marks upper crater size that has reached "equilibrium" (see Gault 1970). As a surface "ages", the total number of craters shifts toward larger sizes and greater numbers; surface 1 (oldest), surface 3 (youngest), assuming all craters are of impact origin.

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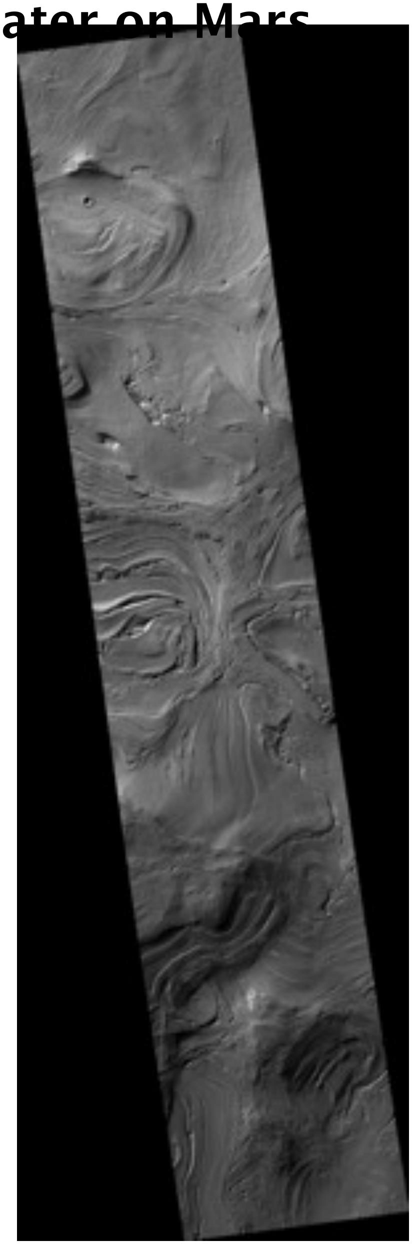
A surface marked by folded layers from the basin floor of the Hellas Crater on Mars



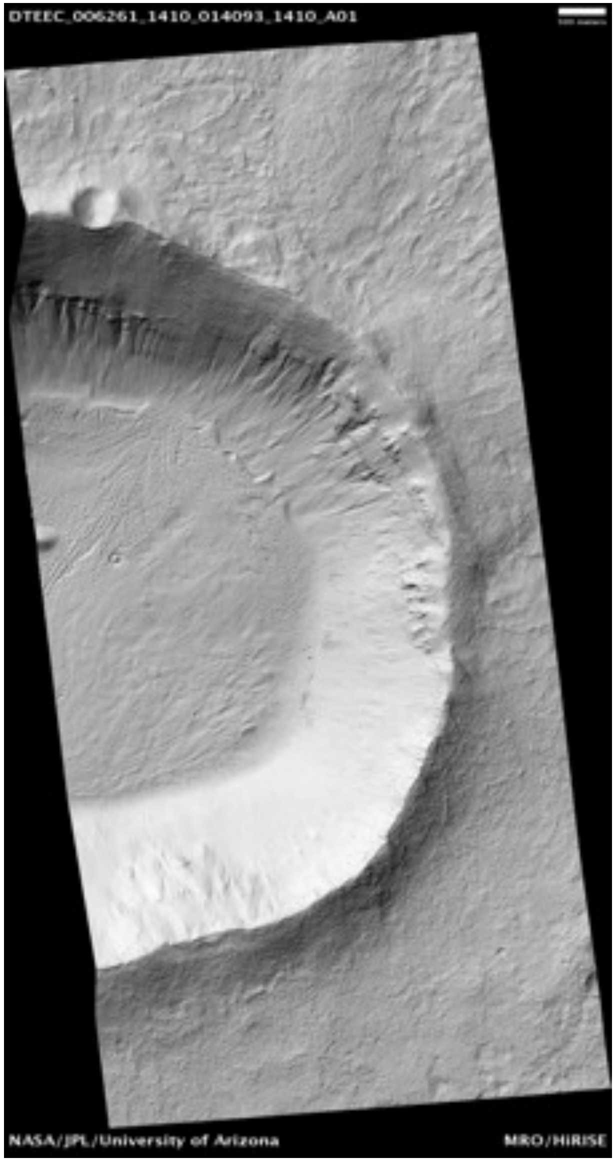
Morphology: Highly controlled by folded layered units; resisting units form ridges while weak units form depressions.

Composition: Sediments (many infer it as glacial deposits and glacier-related deformation)

Age: Difficult to estimate if one does not understand what the operating geologic processes that shaped the surface.

Texture: Strongly controlled by the bedrock structures

Detailed view of a simple crater basin



Morphology: crater pit, rim, and surrounding plains.

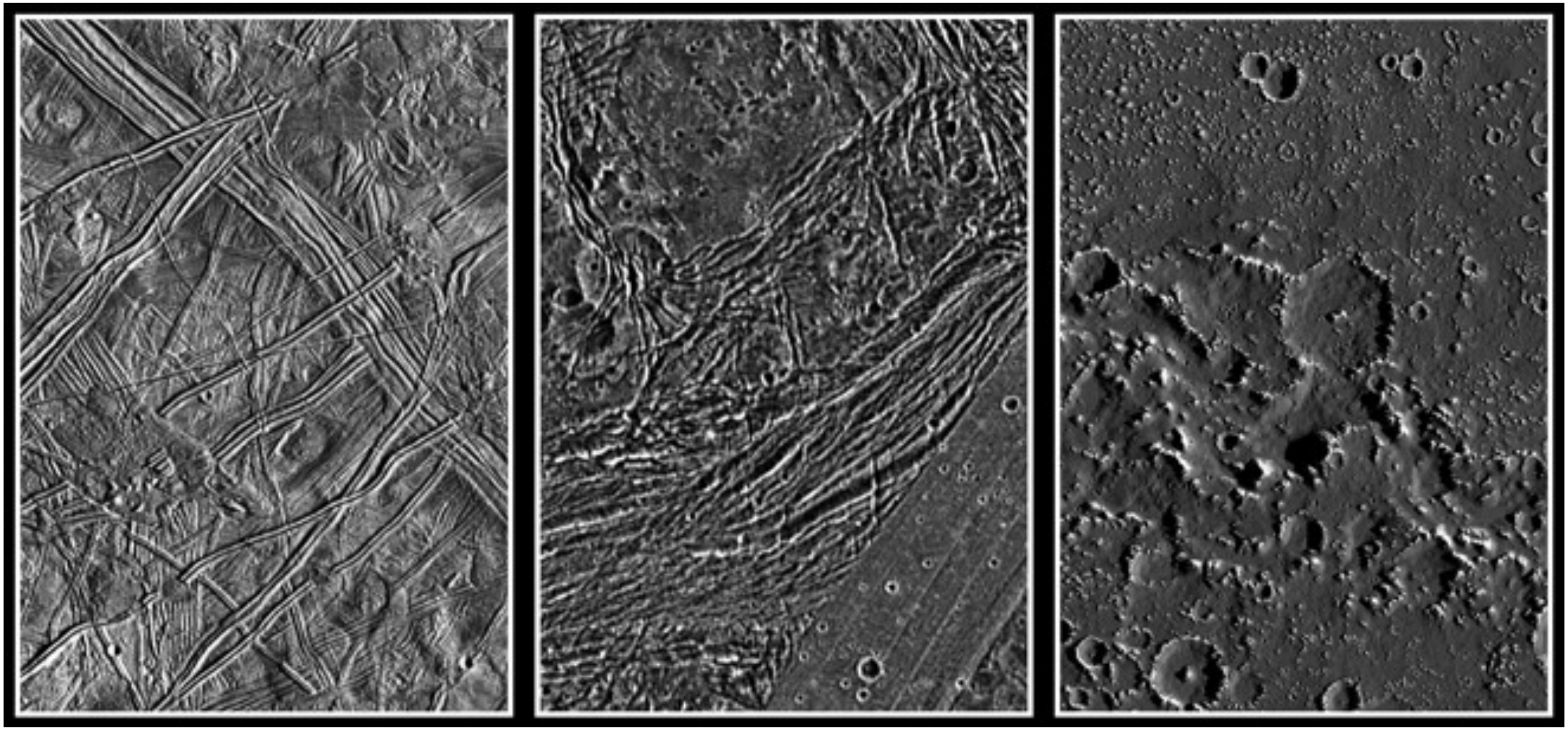
Composition: Layered sedimentary rocks or volcanic flows.

Age: The simple surface rimming the crater may have highly diachronous ages, as the surface deposits appear to be dust, which may have been periodically moblized and thus reset partially or complete the surface.

Texture: Layers can be surface on the crater walls; fine-grained materials draping most of the landscape.

Examples of planetary surfaces defined by their textures, with a decrease in the effect of tectonic forcing

Europa, Ganymede, and Callisto: Surface comparison at high spatial resolution



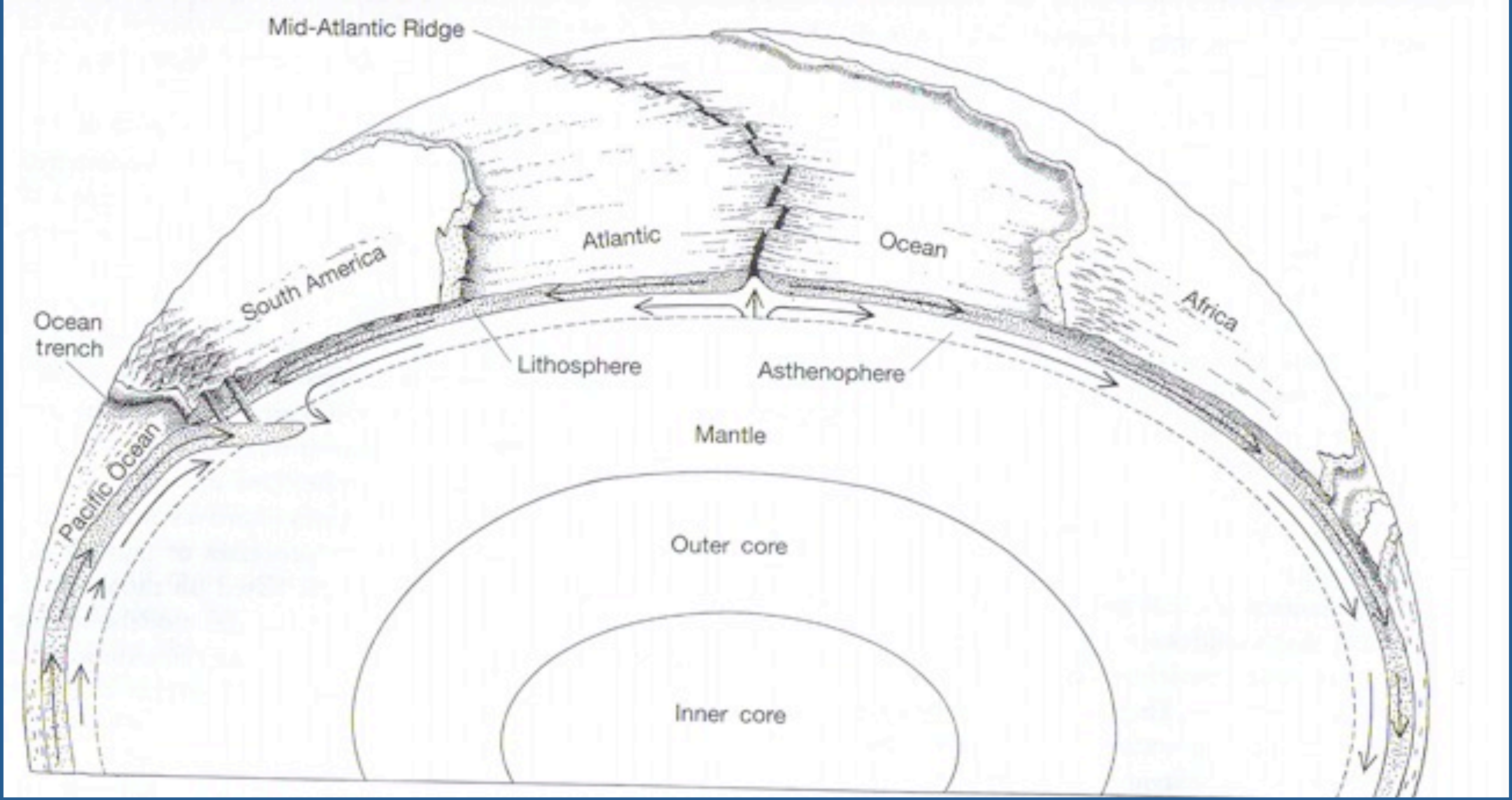
2. What controls the evolution of planetary surfaces?

(1) Tectonic processes driven by mantle convection, tidal forcing, and thermal stress (mainly internally induced for large bodies and in some cases for small bodies).

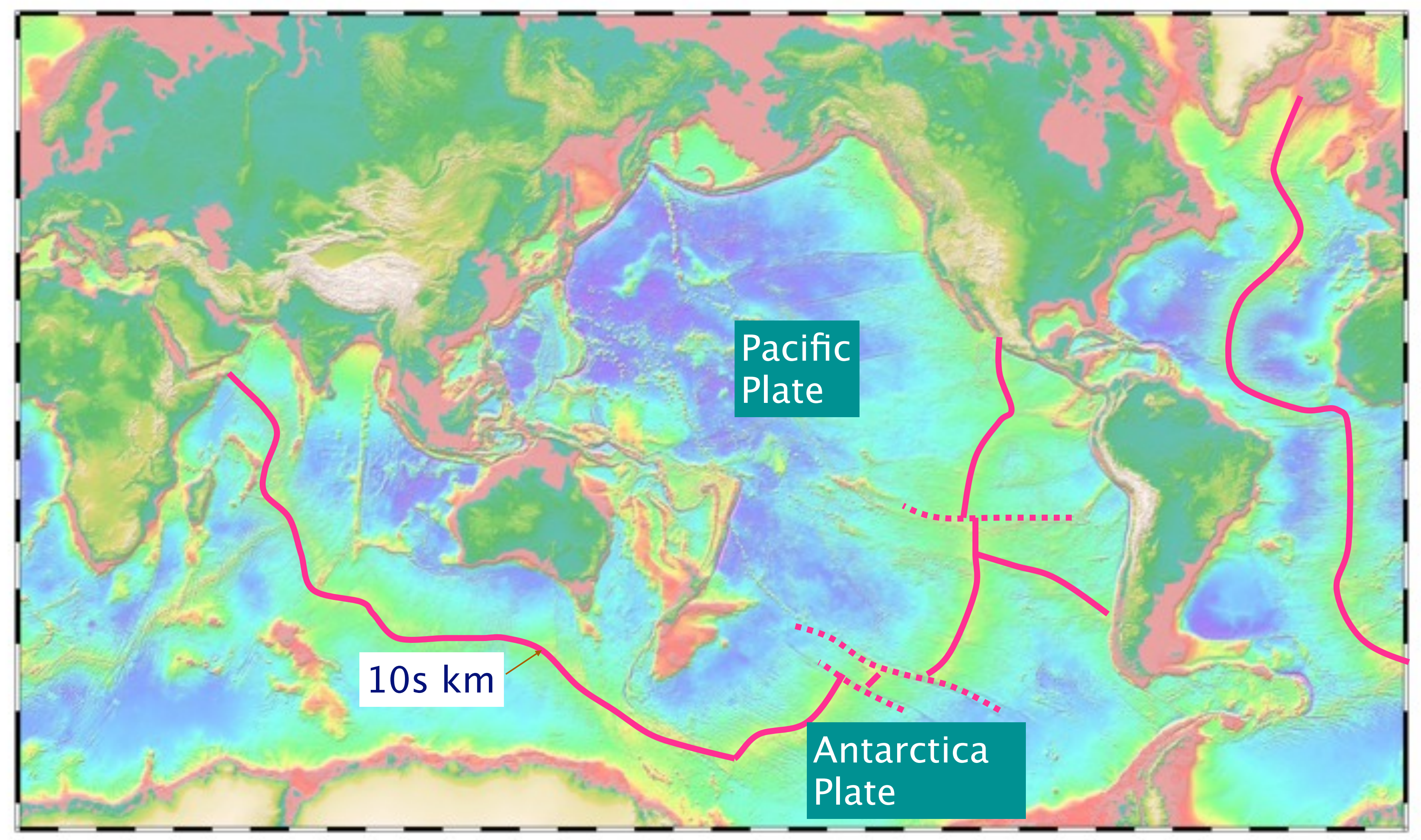
(2) The presence and lack of atmosphere (determining the erosion rates)

(3) Internally driven and externally induced processes: volcanic eruptions, wind, flowing water, moving glacier, weathering, biological activities, mass wasting processes and cratering.

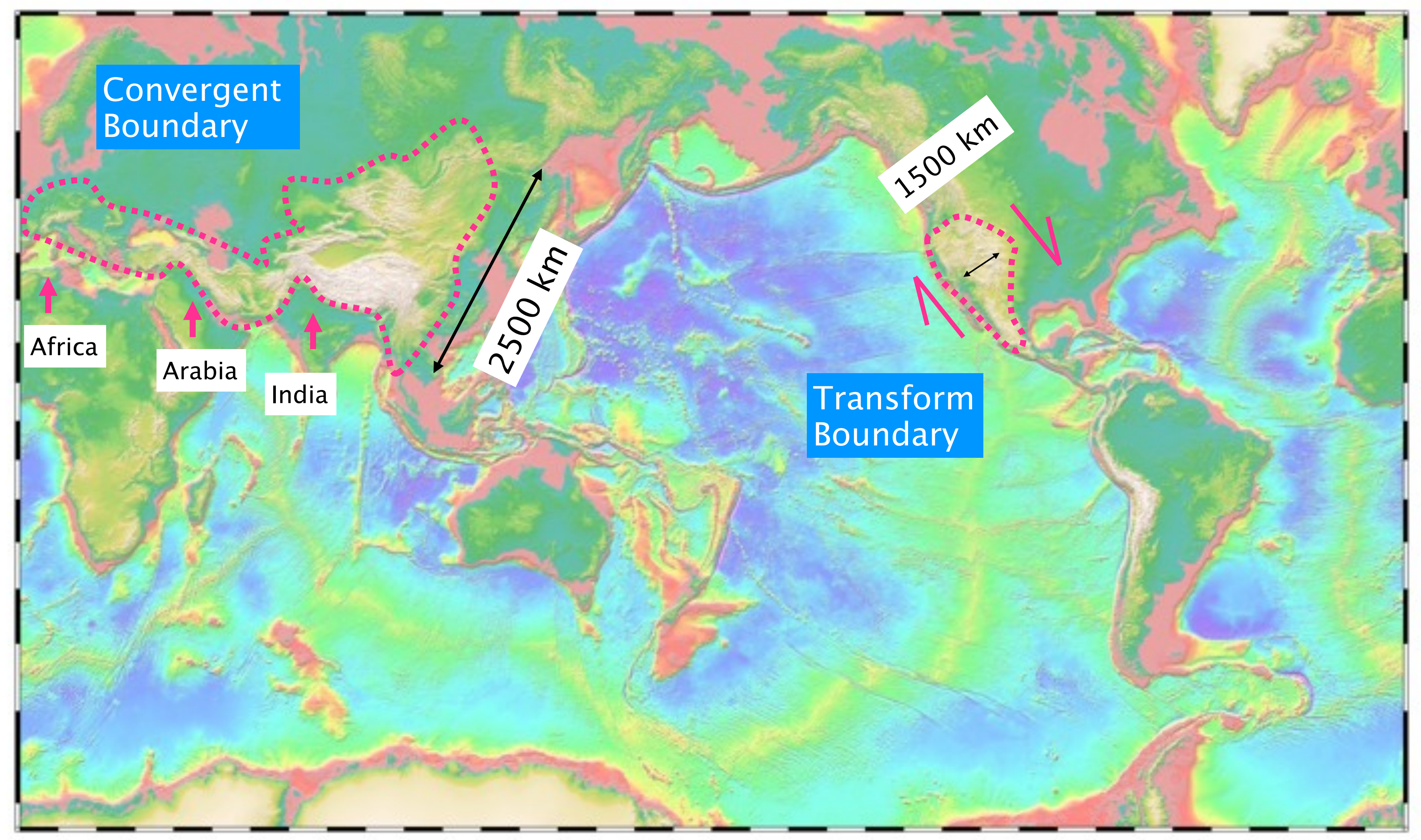
Earth has two modes of tectonics: discrete plate tectonics in the oceanic regions and distributed deformation in the continental areas (plate tectonics is an incomplete description of Earth's crustal deformation).



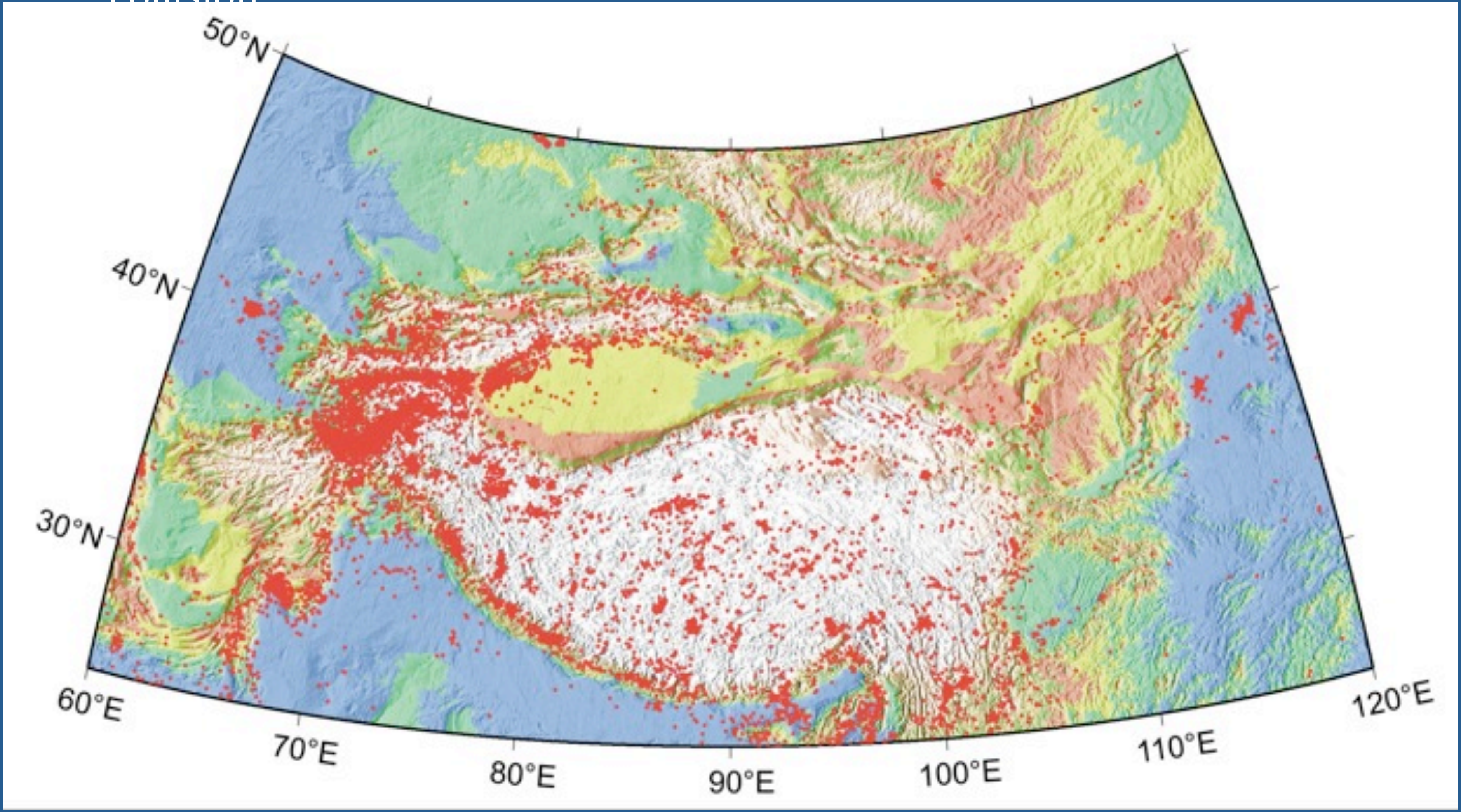
(1) Sharp plate boundaries in the oceanic domain



(2) Diffuse deformation in the continental domain



Best example is the Himalayan–Tibetan orogen induced by the India–Asia collision



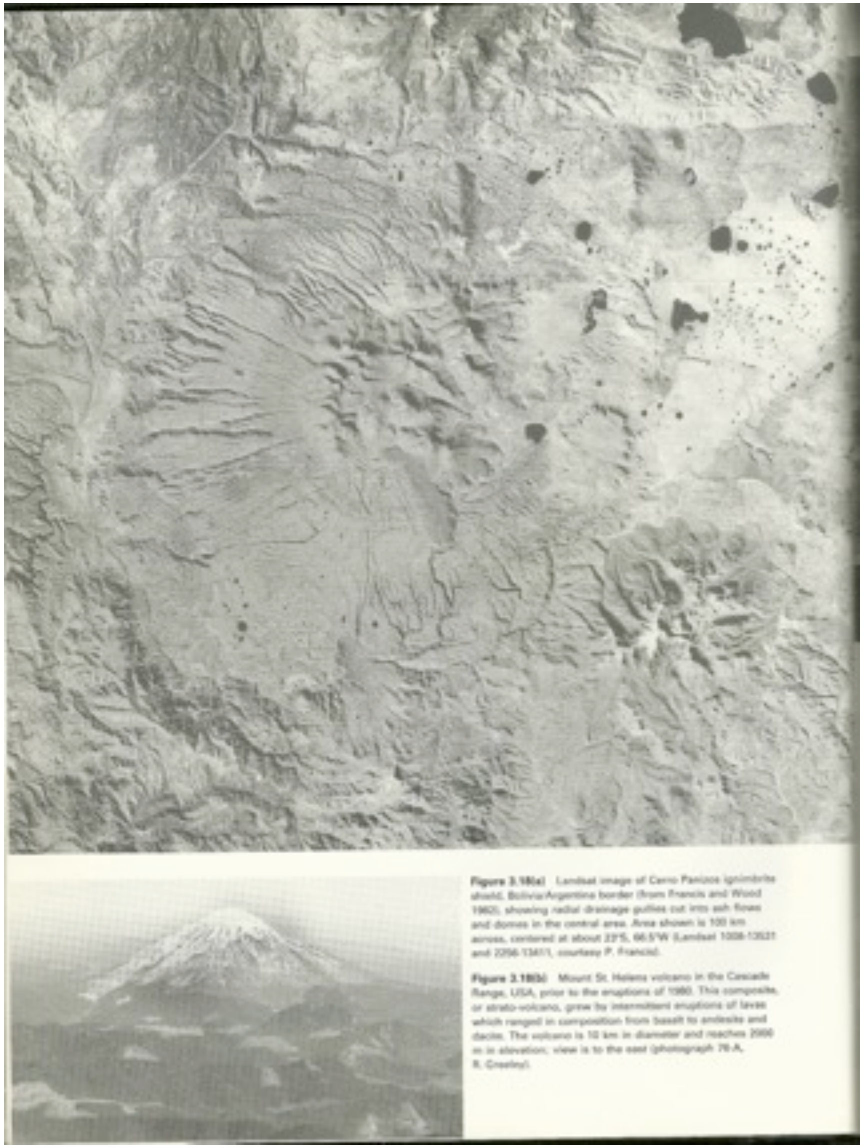


Figure 3.18(a) Landsat image of Cerro Panizos ignimbrite shield, Bolivia/Argentina border (from Francis and Wood 1982), showing radial drainage gullies cut into ash flows and domes in the central area. Area shown is 100 km across, centered at about 22°S, 66.5°W (Landsat 1008-13521 and 2256-13411, courtesy P. Francis).

Figure 3.18(b) Mount St. Helens volcano in the Cascade Range, USA, prior to the eruptions of 1980. This composite, or strato-volcano, grew by intermittent eruptions of lavas which ranged in composition from basalt to andesite and dacite. The volcano is 10 km in diameter and reaches 2950 m in elevation; view is to the east (photograph 78-A, R. Greeley).

Volcanism is an important process in shaping planetary surface (we will learn how to scale lava transport distance and shape of volcanoes for planetary bodies of different sizes.

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Lava flow channel on Moon (mysterious source at the "snake head")

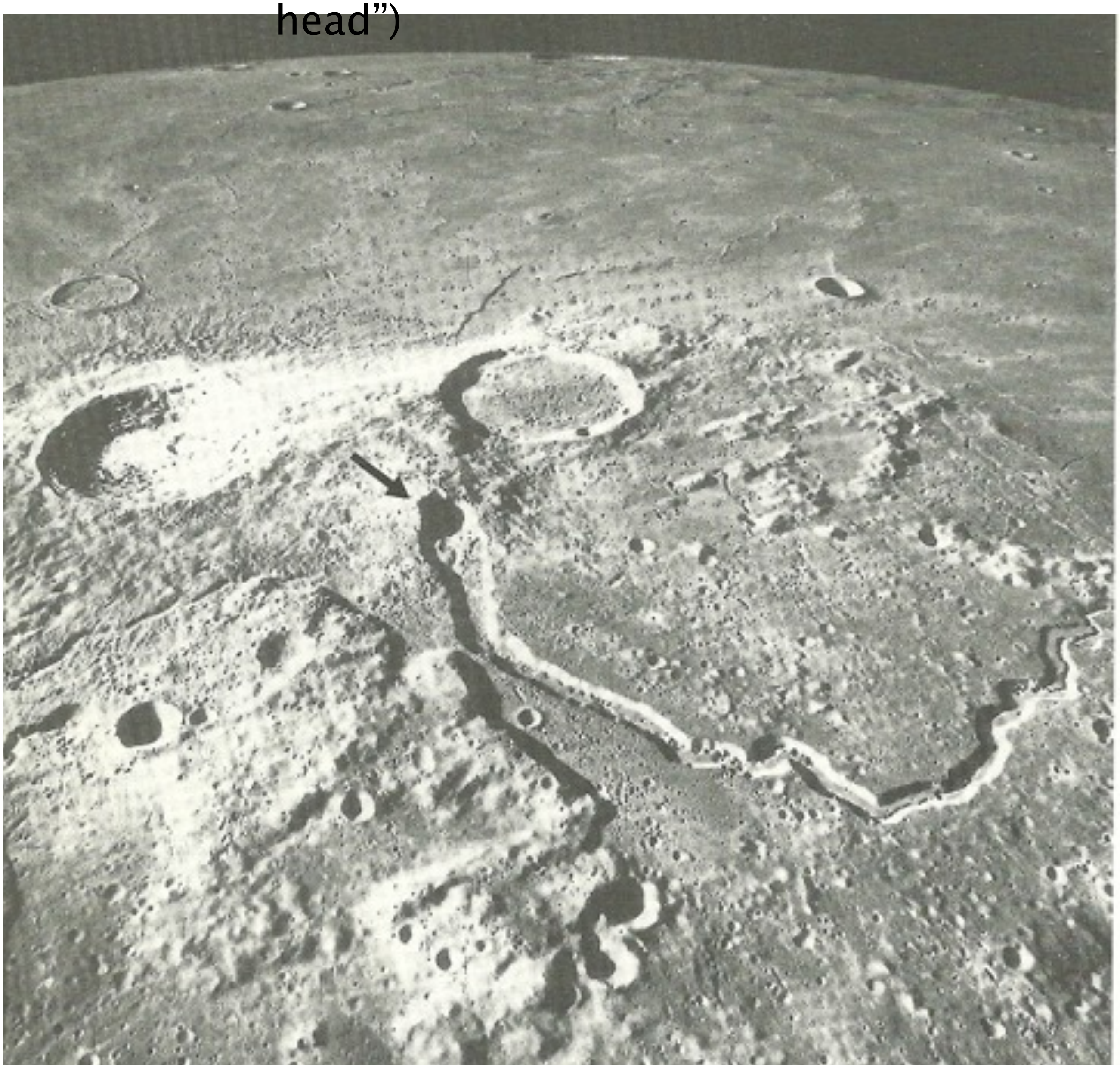

Figure 4.15 Oblique view taken by the Apollo 15 astronauts of the Aristarchus plateau and Schröter's Valley. The 40 km fresh impact crater in the upper left is Aristarchus. Schröter's Valley is more than 250 km long and exceeds 10 km in width. This rille originates in the "Cobra head" (arrow) and is considered to be a tectonic feature controlled by faulting combined with a volcanic flow channel which served as a vent for lavas emptying into Oceanus Procellarum to the upper right. The Aristarchus plateau is one of the major volcanic centers on the Moon. Illumination from the left (Apollo 15 AS15-M3-2611).

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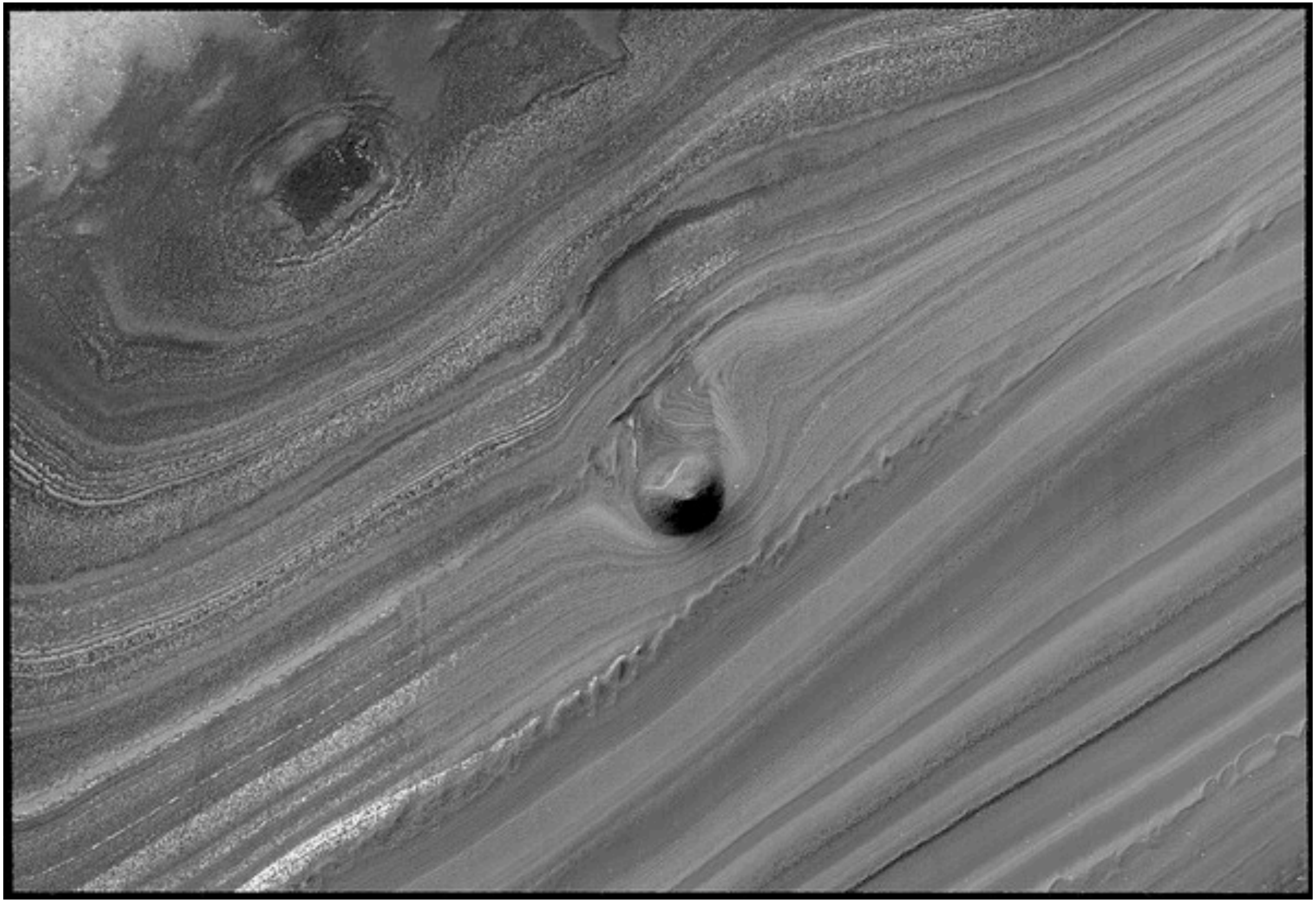
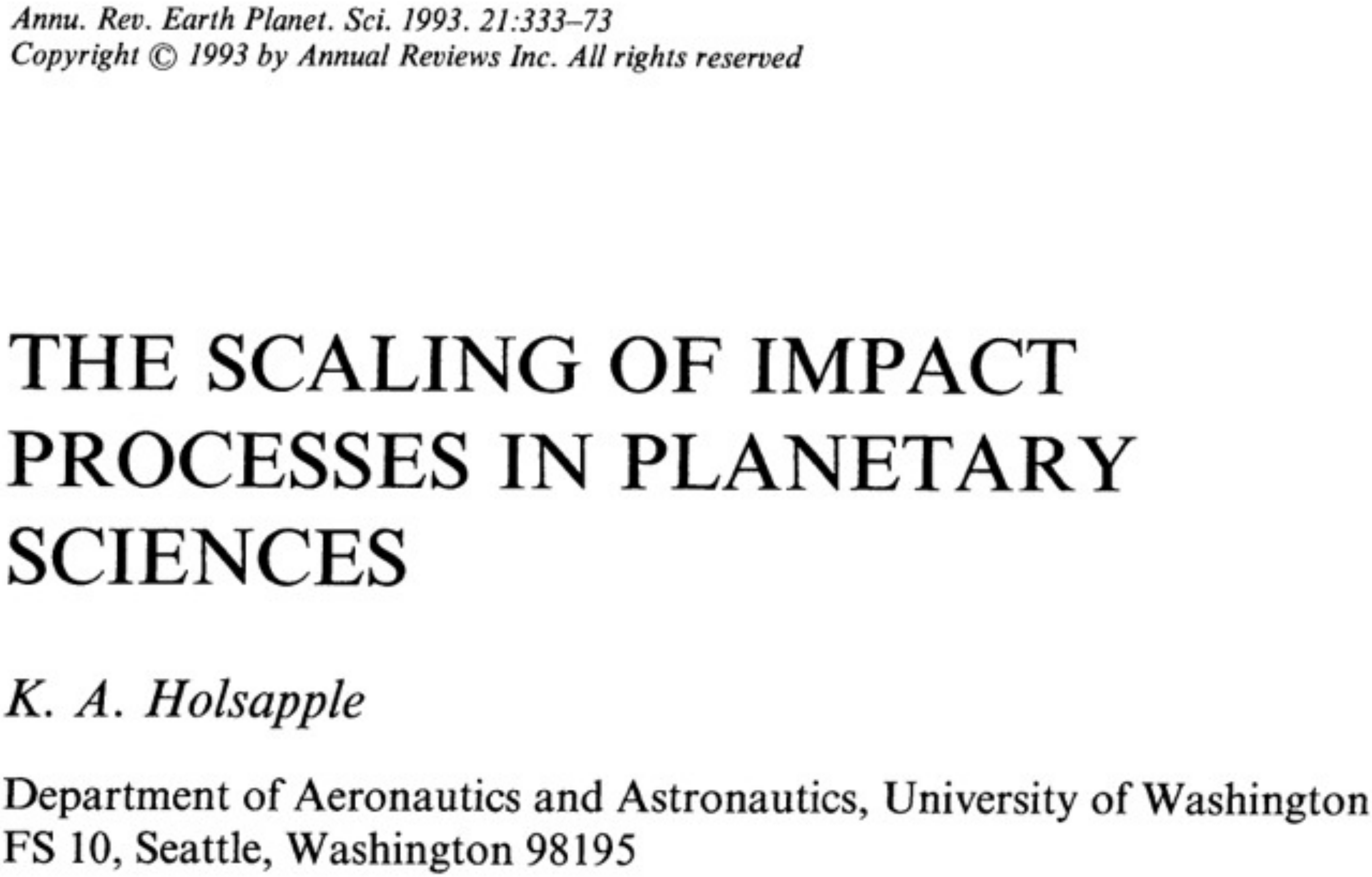
Impact crater in the northern polar region of Mars, partially buried glacier flow



3. How do surface-shaping processes scale across planetary bodies of different sizes, compositions, and physical states?

(We will use crater-size scaling laws to illustrate this point)

(The presentation below is mostly based on the paper below)



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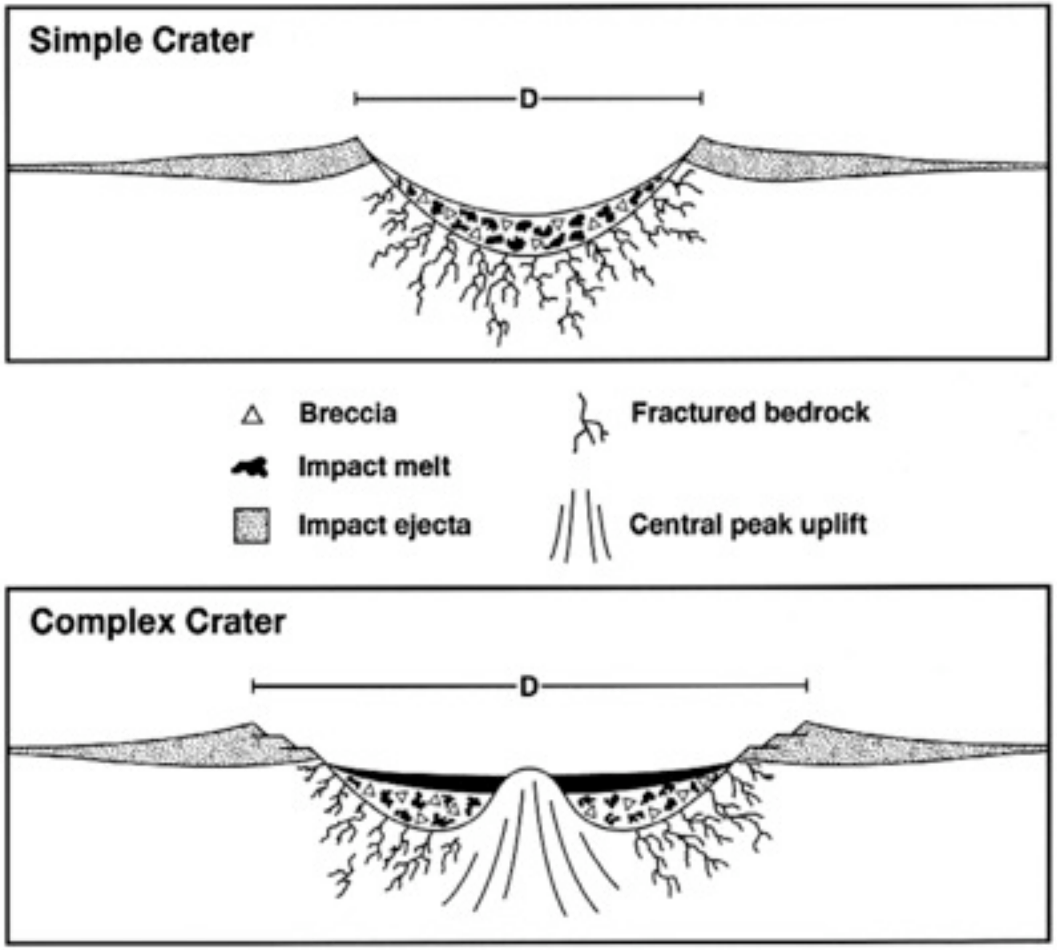
THE SCALING OF IMPACT PROCESSES IN PLANETARY SCIENCES

K. A. Holsapple

Department of Aeronautics and Astronautics, University of Washington
FS 10, Seattle, Washington 98195

Processes of impact cratering: contact, compression, decompression, and the passage of the shock wave (i.e. explosion); all occur within a few tenths of a second for a large impact.

Simple and Complex (peak-ring) Craters

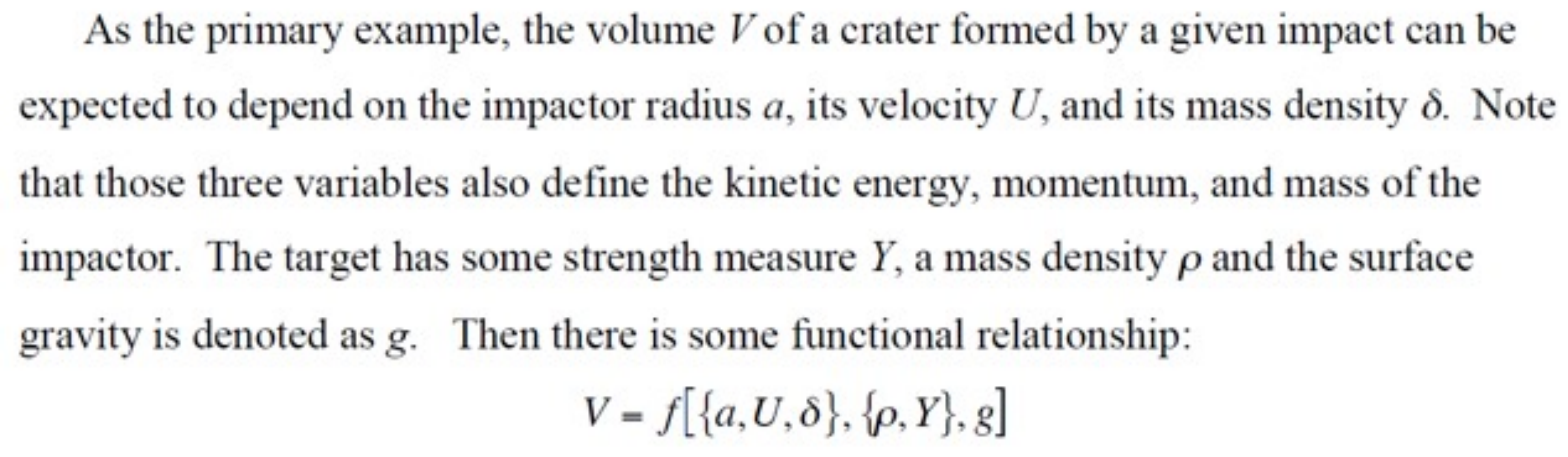


The fundamental goal of scaling is to find a relationship that the radius of a crater is a function of impact size and impact velocity:

$$R = f(a, U).$$

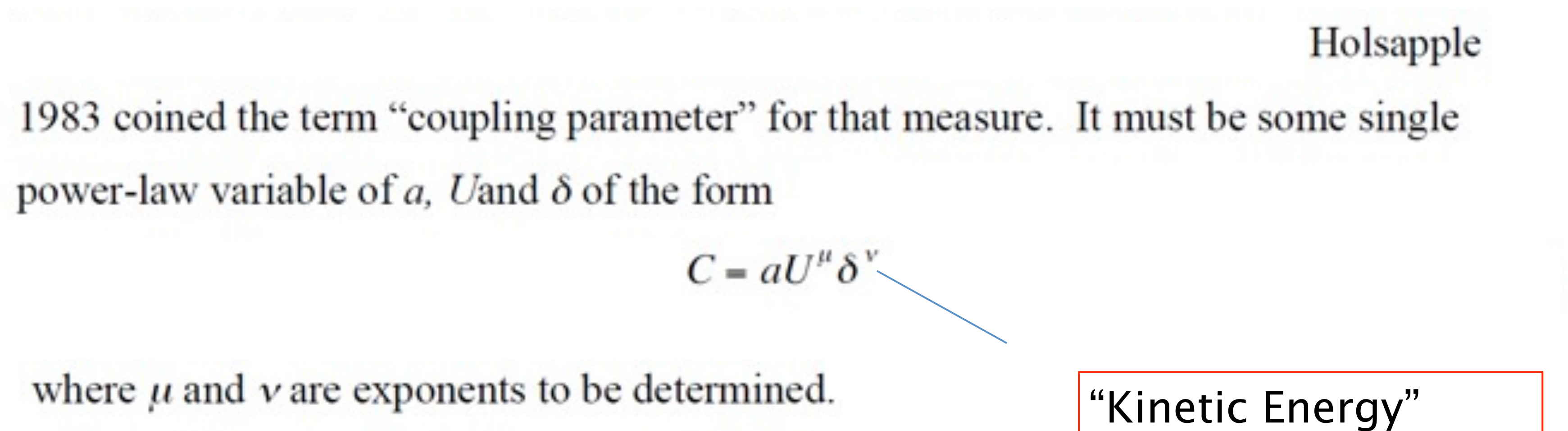
R, radius of crater
a, radius of impactor
U, impact velocity

Crater Scaling Relationship based on Point Source and Vertical Impact



As the primary example, the volume V of a crater formed by a given impact can be expected to depend on the impactor radius a, its velocity U, and its mass density δ. Note that those three variables also define the kinetic energy, momentum, and mass of the impactor. The target has some strength measure Y, a mass density ρ and the surface gravity is denoted as g. Then there is some functional relationship:

$$V = f\left[\{a, U, \delta\}, \{\rho, Y\}, g\right]$$



Holsapple 1983 coined the term "coupling parameter" for that measure. It must be some single power-law variable of a, U and δ of the form

$$C = aU^{\mu}\delta^{\nu}$$

where μ and ν are exponents to be determined.

"Kinetic Energy" form inspired by both physics and experiments

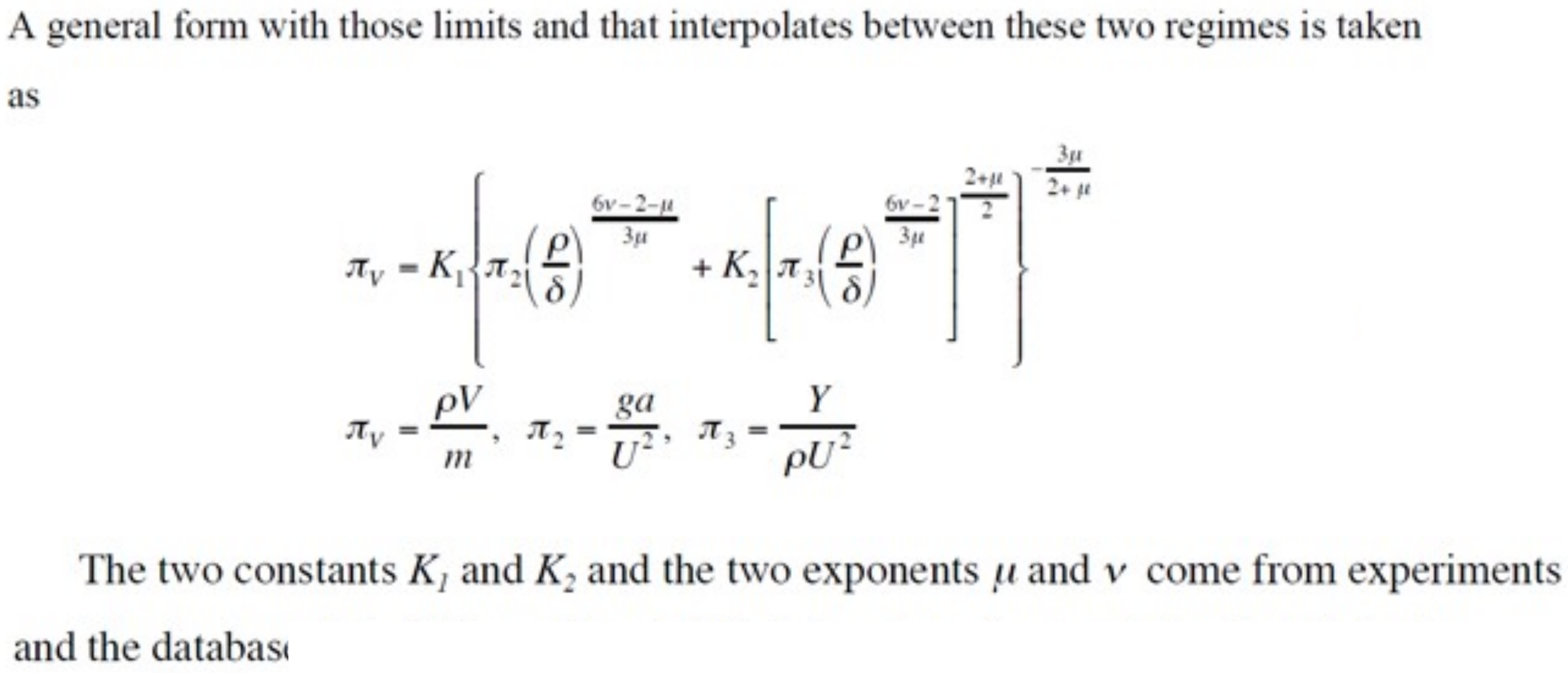
Holsapple (1983) defined two regimes for point source approximation:

1. Strength Regime (smaller craters, i.e., radius < 1 km)

$$V \propto \frac{m}{\rho}\left(\frac{\rho U^2}{Y}\right)^{\frac{3\mu}{2}}\left(\frac{\rho}{\delta}\right)^{1-3\nu}$$

2. Gravity Retime (large craters, i.e., radius > 1 km)

$$V \propto \frac{m}{\rho}\left[\frac{ga}{U^2}\right]^{\frac{-3\mu}{2+\mu}}\left(\frac{\rho}{\delta}\right)^{\frac{2+\mu-6\nu}{2+\mu}}$$



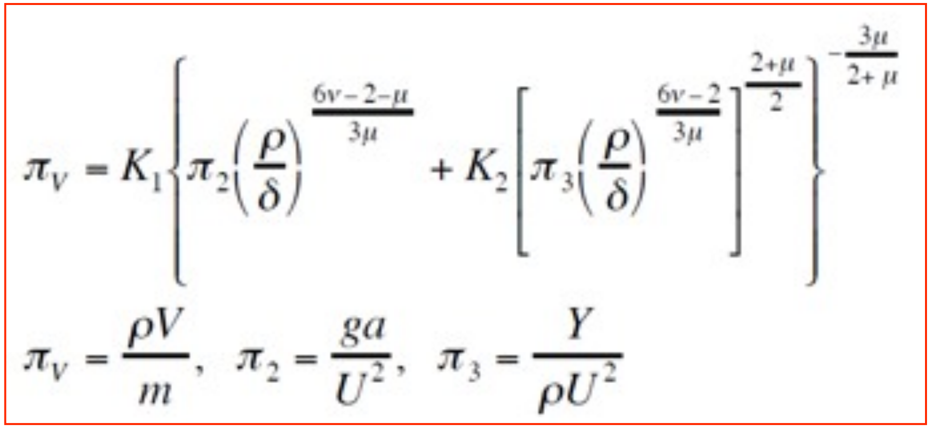
A general form with those limits and that interpolates between these two regimes is taken as

$$\pi_V = K_1\left\{\pi_2\left(\frac{\rho}{\delta}\right)^{\frac{6\nu-2-\mu}{3\mu}} + K_2\left[\pi_3\left(\frac{\rho}{\delta}\right)^{\frac{6\nu-2}{3\mu}}\right]^{\frac{2+\mu}{2}}\right\}^{-\frac{3\mu}{2+\mu}}$$

$$\pi_V = \frac{\rho V}{m}, \quad \pi_2 = \frac{ga}{U^2}, \quad \pi_3 = \frac{Y}{\rho U^2}$$

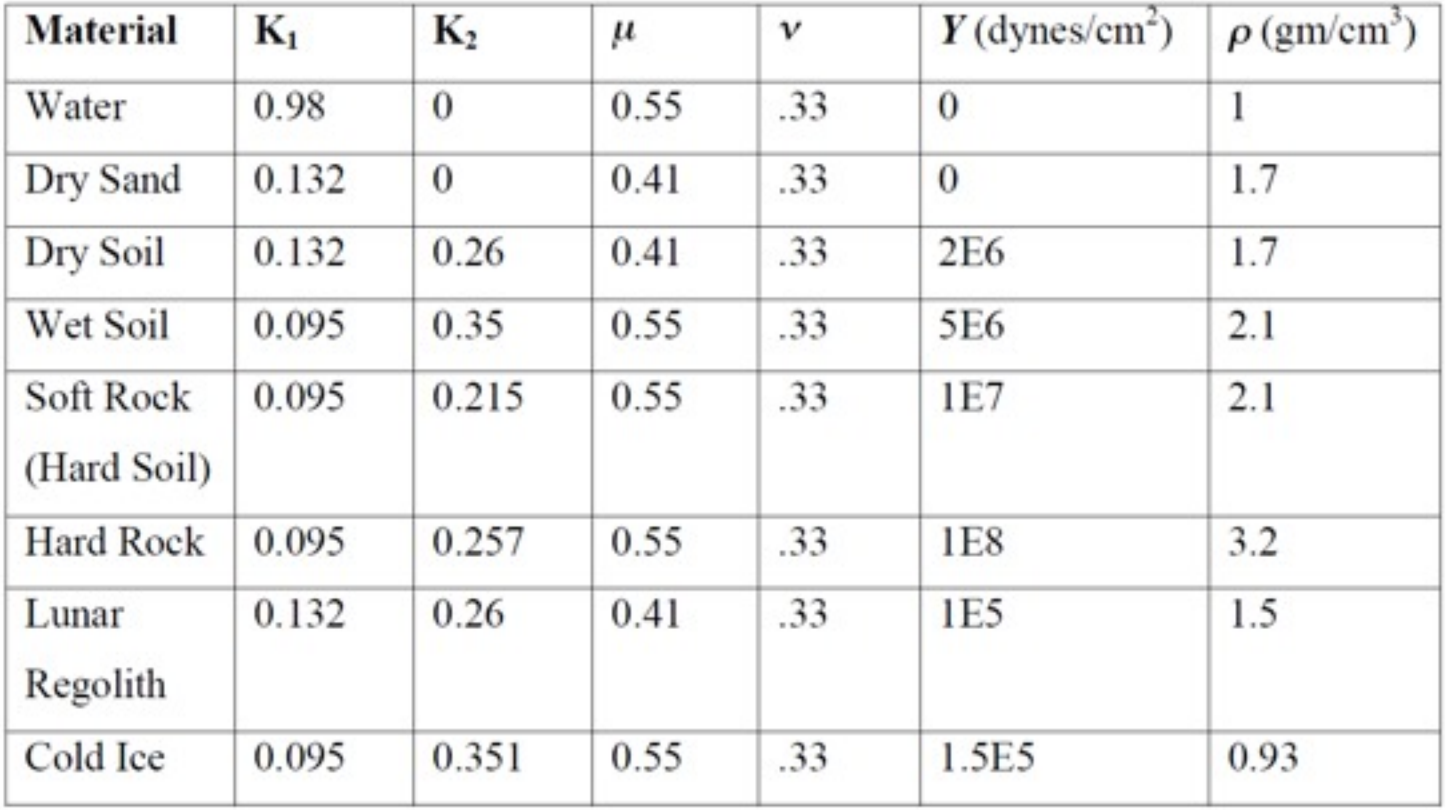
The two constants K_1 and K_2 and the two exponents μ and ν come from experiments and the databas

The above form is generally referred to as “the π-scaling law”

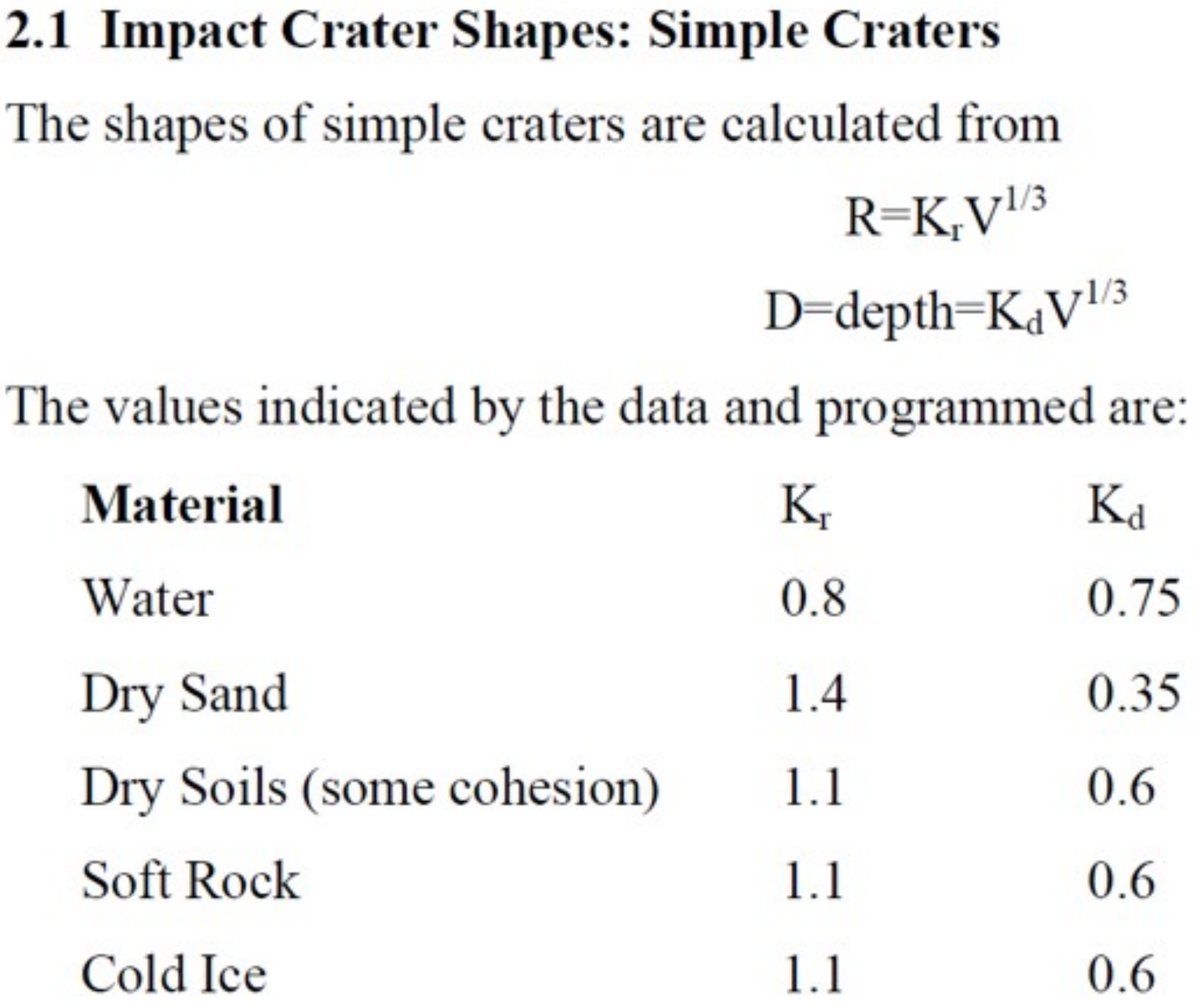


$$\pi_V = K_1\left\{\pi_2\left(\frac{\rho}{\delta}\right)^{\frac{6\nu-2-\mu}{3\mu}} + K_2\left[\pi_3\left(\frac{\rho}{\delta}\right)^{\frac{6\nu-2}{3\mu}}\right]^{\frac{2+\mu}{2}}\right\}^{-\frac{3\mu}{2+\mu}}$$

$$\pi_V = \frac{\rho V}{m},\quad \pi_2 = \frac{ga}{U^2},\quad \pi_3 = \frac{Y}{\rho U^2}$$



Material	**K_1**	**K_2**	μ	ν	Y (dynes/cm^2)	ρ (gm/cm^3)
Water	0.98	0	0.55	.33	0	1
Dry Sand	0.132	0	0.41	.33	0	1.7
Dry Soil	0.132	0.26	0.41	.33	2E6	1.7
Wet Soil	0.095	0.35	0.55	.33	5E6	2.1
Soft Rock (Hard Soil)	0.095	0.215	0.55	.33	1E7	2.1
Hard Rock	0.095	0.257	0.55	.33	1E8	3.2
Lunar Regolith	0.132	0.26	0.41	.33	1E5	1.5
Cold Ice	0.095	0.351	0.55	.33	1.5E5	0.93



2.1 Impact Crater Shapes: Simple Craters

The shapes of simple craters are calculated from

$$R=K_rV^{1/3}$$

$$D=\text{depth}=K_dV^{1/3}$$

The values indicated by the data and programmed are:

Material	K_r	K_d
Water	0.8	0.75
Dry Sand	1.4	0.35
Dry Soils (some cohesion)	1.1	0.6
Soft Rock	1.1	0.6
Cold Ice	1.1	0.6

Lab experiment based scaling law

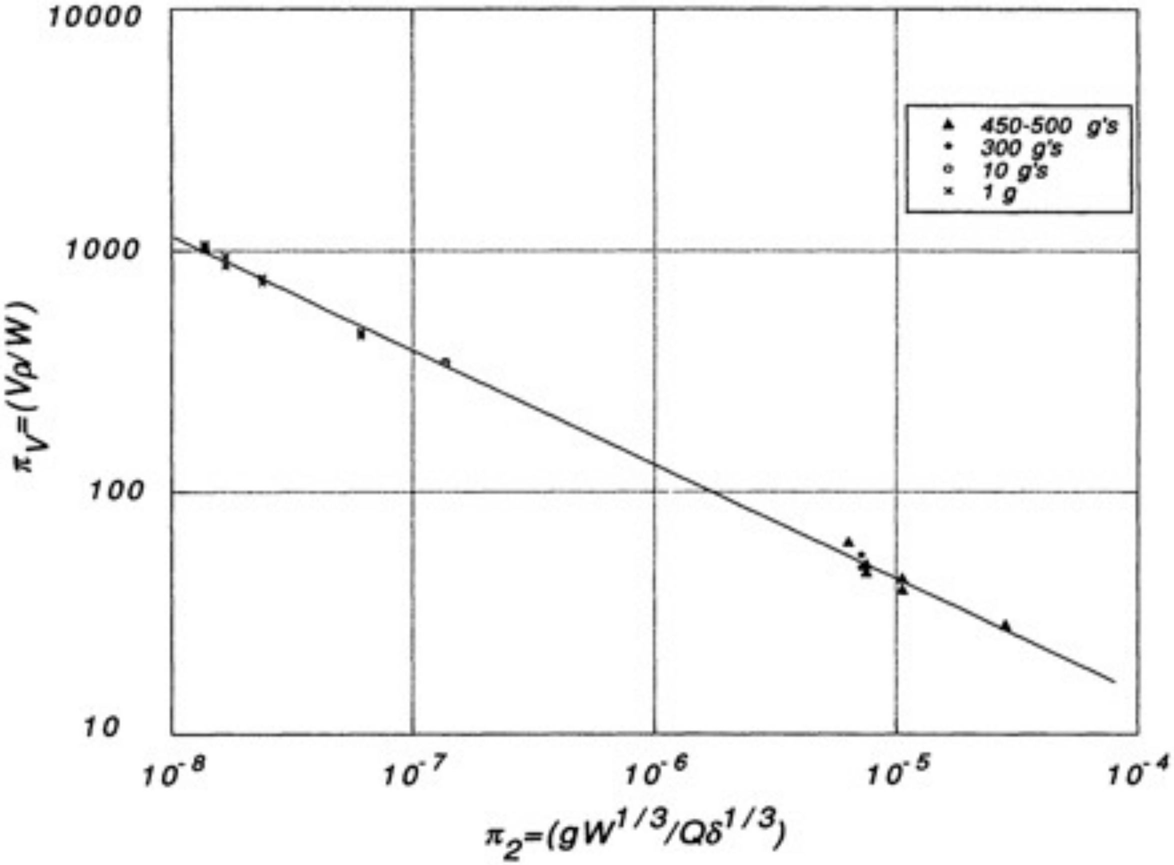


Figure 1 Explosive cratering results in a dry sand at normal and elevated gravity. The power-law fit extends at least four decades in the gravity-scaled size parameter, giving decreasing cratering efficiency with either increasing explosive size or increasing gravity.

Cratering-in-water experiments

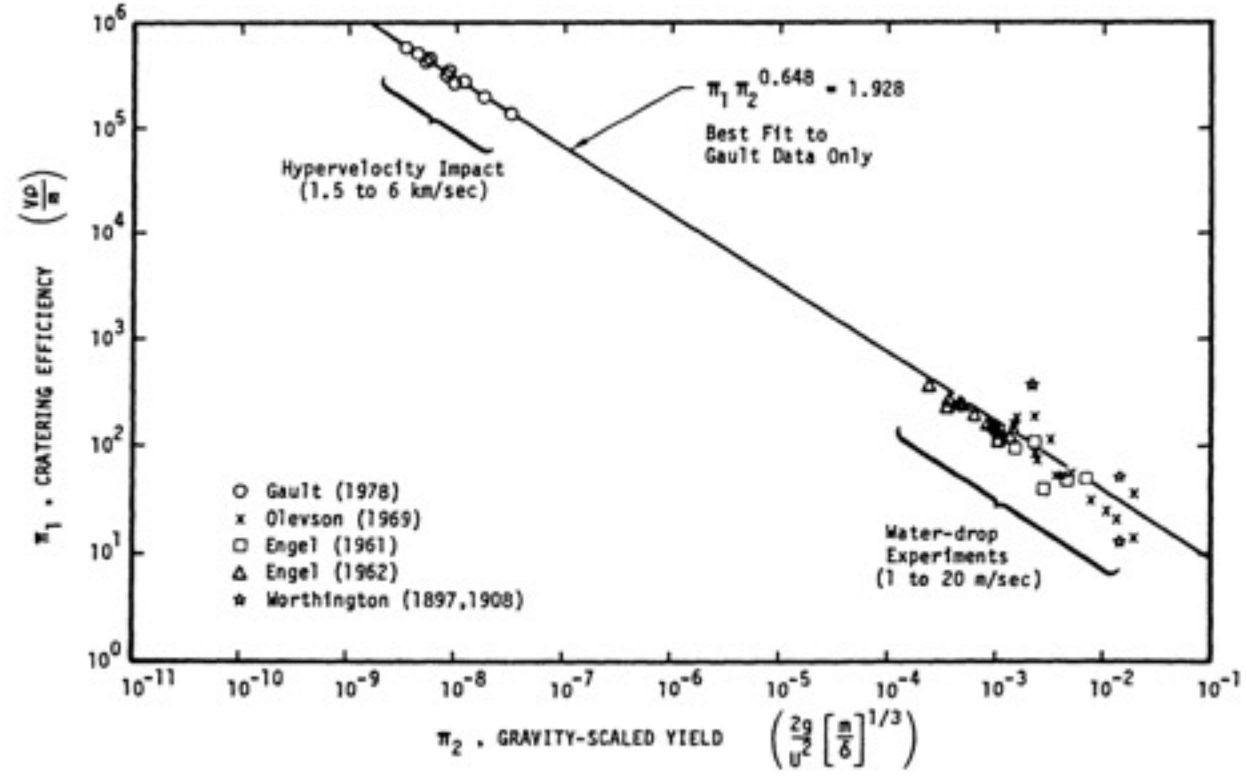


Figure 2 Data for cratering in water for velocities from 1 m/sec to 6 km/sec. A single power-law fit through the hypervelocity data goes right through the low speed data. Thus, a single power-law fit holds for over 8 decades of impactor radius (24 decades of mass), 4 decades in impact velocity, and 8 decades in gravity.

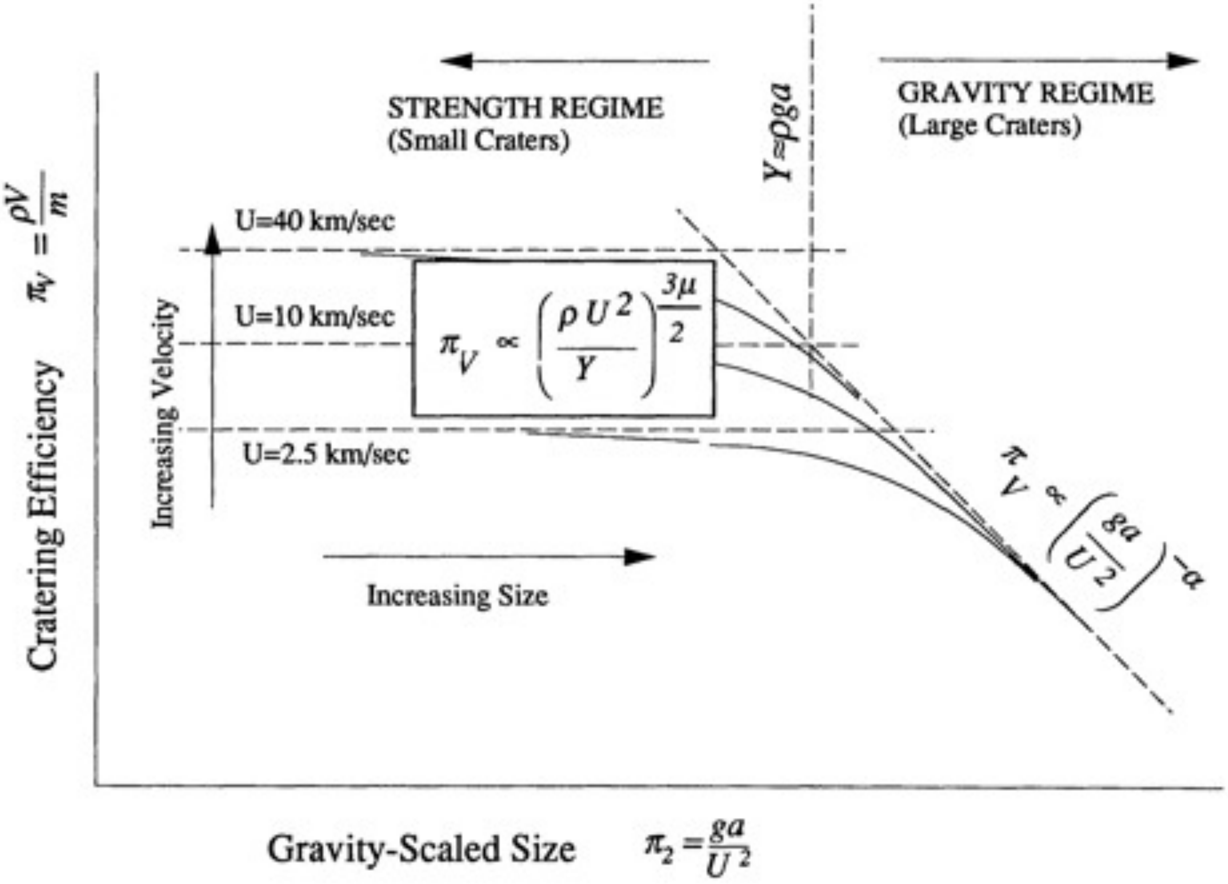


Figure 3 The regimes of cratering for a material with strength. In the strength regime the cratering efficiency depends on the impact velocity, but is independent of gravity-scaled size. For increasing size at a fixed velocity, there is a transition to the gravity regime in which the cratering efficiency has a power law decrease with increasing size. Most experiments in geological materials are by necessity in the strength regime.

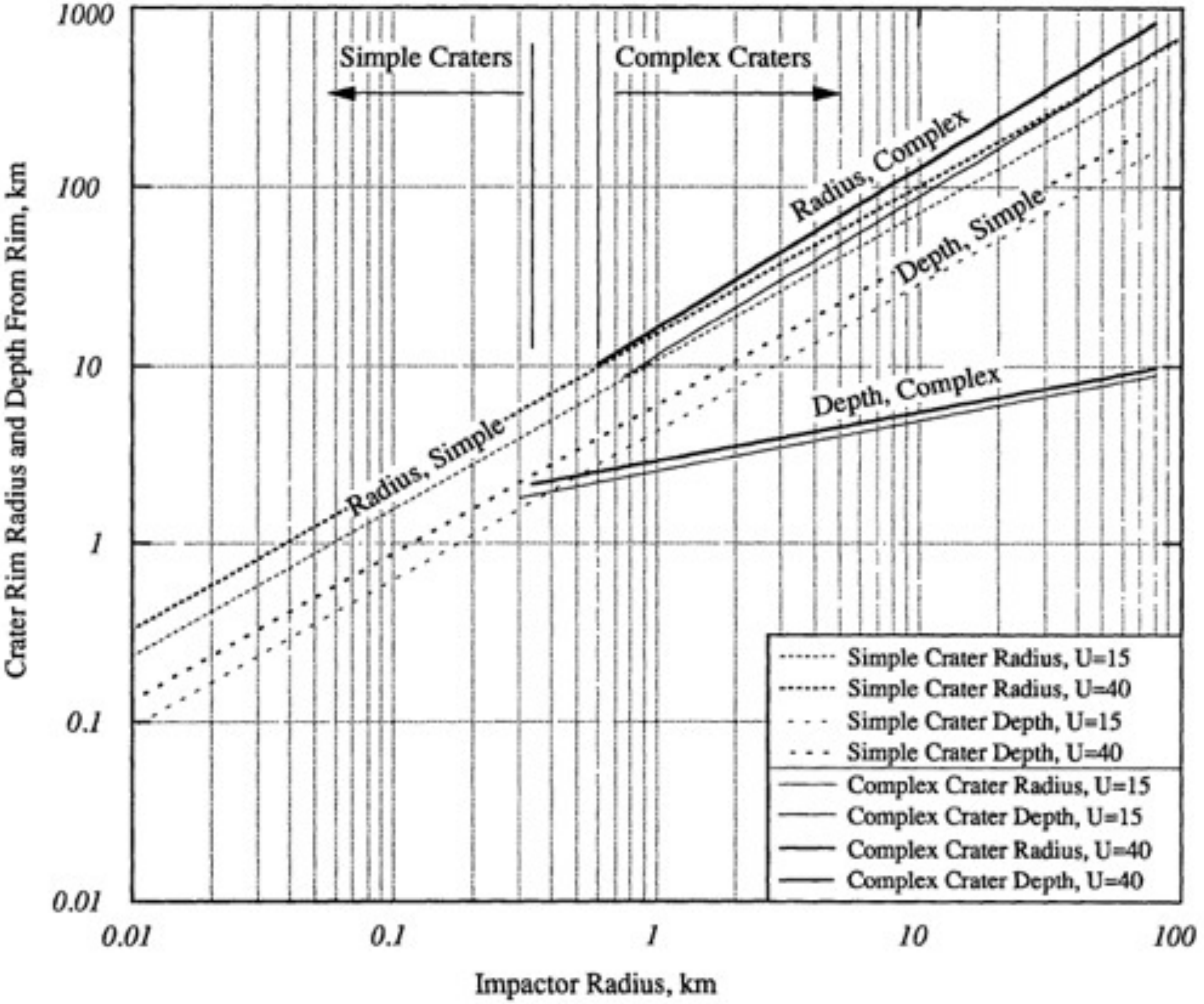


Figure 10 Crater radius and depth for lunar craters showing both the transitory simple crater and the final complex crater sizes.

Evacuation radius:

$$R_e = 7.8G^{-0.17}a^{0.83}U^{0.34}$$

Rim radius:

$$R_r = 10.14G^{-0.17}a^{0.83}U^{0.34}.$$

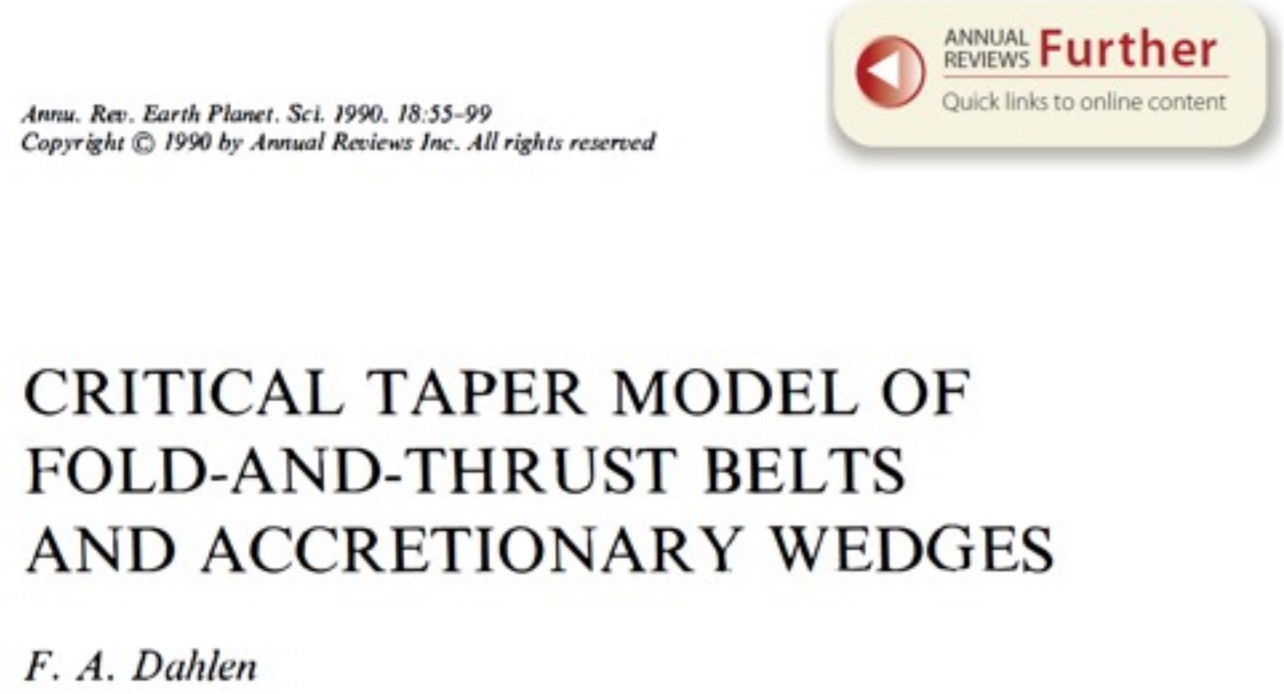
4. How do surface-shaping processes interact? What is the role of feedback?

(I will use Earth example to best illustrate this point)

The most prominent features on Earth are mountain belts. They are the best expression of tectonic forces driven by mantle convection. However, the mountain building processes are strongly coupled with Climate and Surface Processes.

To understand how the coupling works, one need to first understand how mountains were built. We will consider the simplest theory**, the Coulomb Critical Wedge Hypothesis.**

The lecture on the critical Coulomb wedge model for mountain building is mainly based on the following review paper, a classic!



Annu. Rev. Earth Planet. Sci. 1990. 18:55-99

CRITICAL TAPER MODEL OF
FOLD-AND-THRUST BELTS
AND ACCRETIONARY WEDGES

F. A. Dahlen

Wedge-shaped mountain belts on Earth (possibly on Venus and locally on Mars)

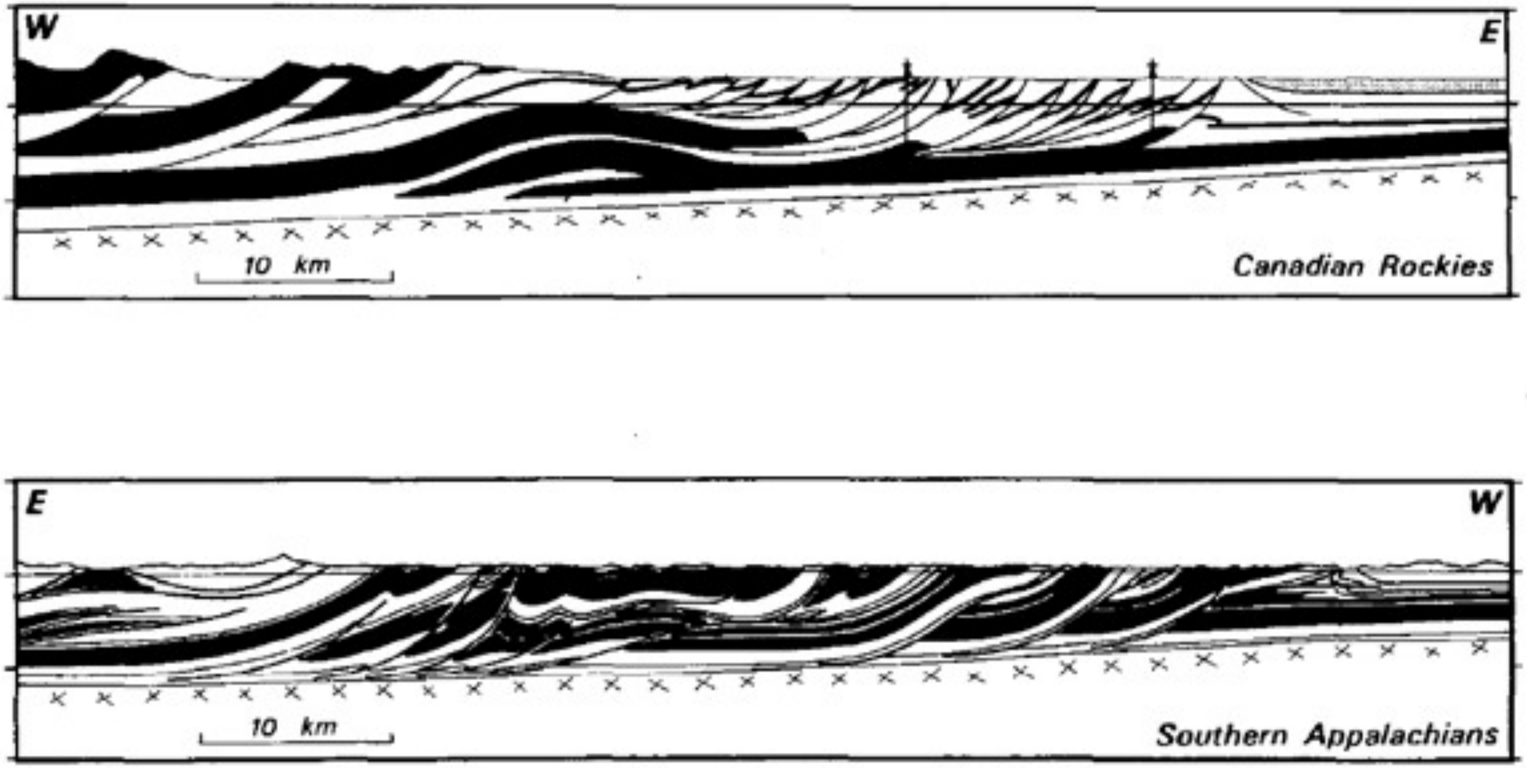


Figure 1 Cross sections of two foreland fold-and-thrust belts. (*Top*) Canadian Rockies (Bally et al 1966). (*Bottom*) Southern Appalachians (Roeder et al 1978). No vertical exaggeration.

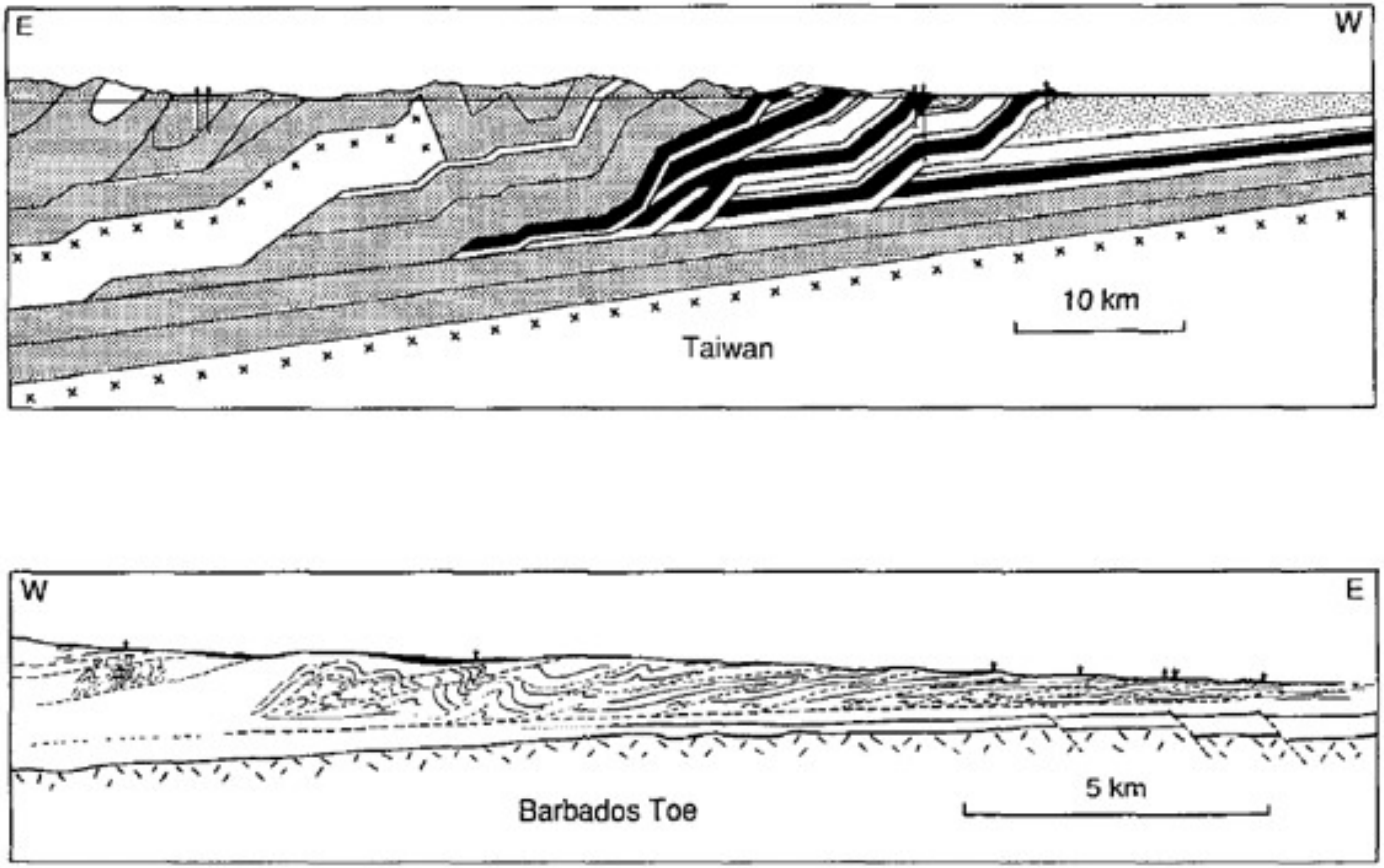


Figure 2 (*Top*) Cross section of the active fold-and-thrust belt in northern Taiwan (Suppe 1980). (*Bottom*) Cross section of the frontal region of the Barbados accretionary wedge near 15°30′N latitude (Behrmann et al 1988). Locations of Deep Sea Drilling Project and Ocean Drilling Program drill sites are indicated. No vertical exaggeration.

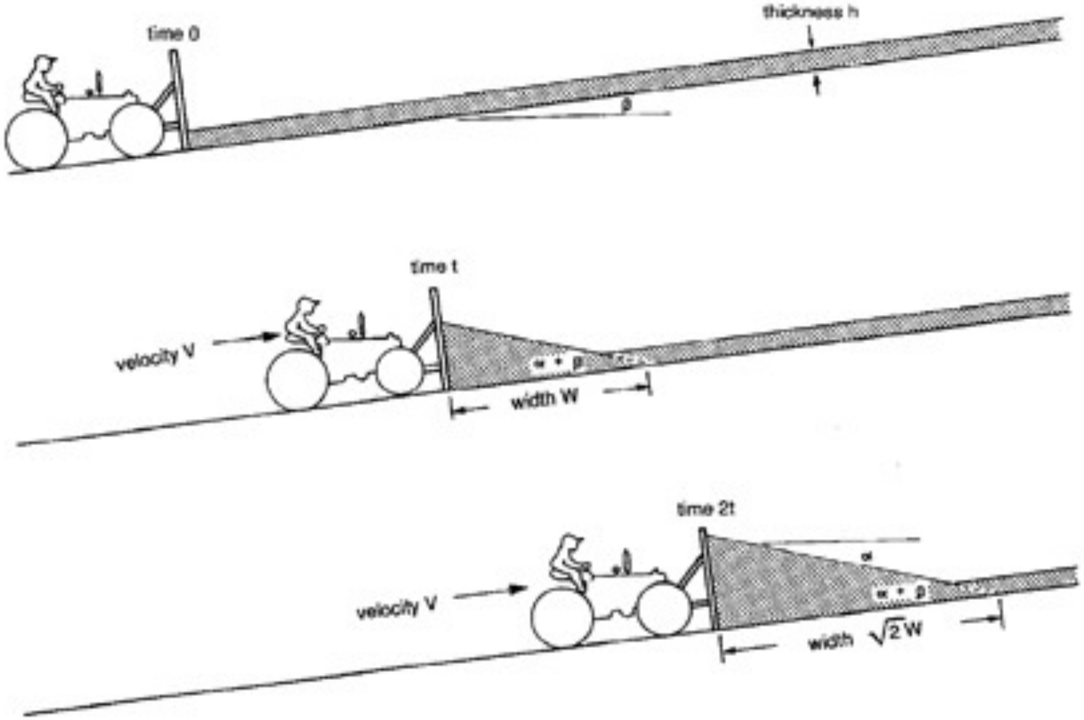


Figure 4 Cartoon depicting the self-similar growth of a bulldozer wedge.

and assume that ρ is a constant. The growth of the wedge with time is described by the mass conservation law

$$\frac{\mathrm{d}}{\mathrm{d}t}\left[\frac{1}{2}\rho W^2 \tan(\alpha+\beta)\right] = \rho h V,$$

As wedge grows self-similarly and the taper is thin, the steady state wedge follows

$$\dot{e} W \sec(\alpha+\beta) \approx \dot{e} W = hV$$

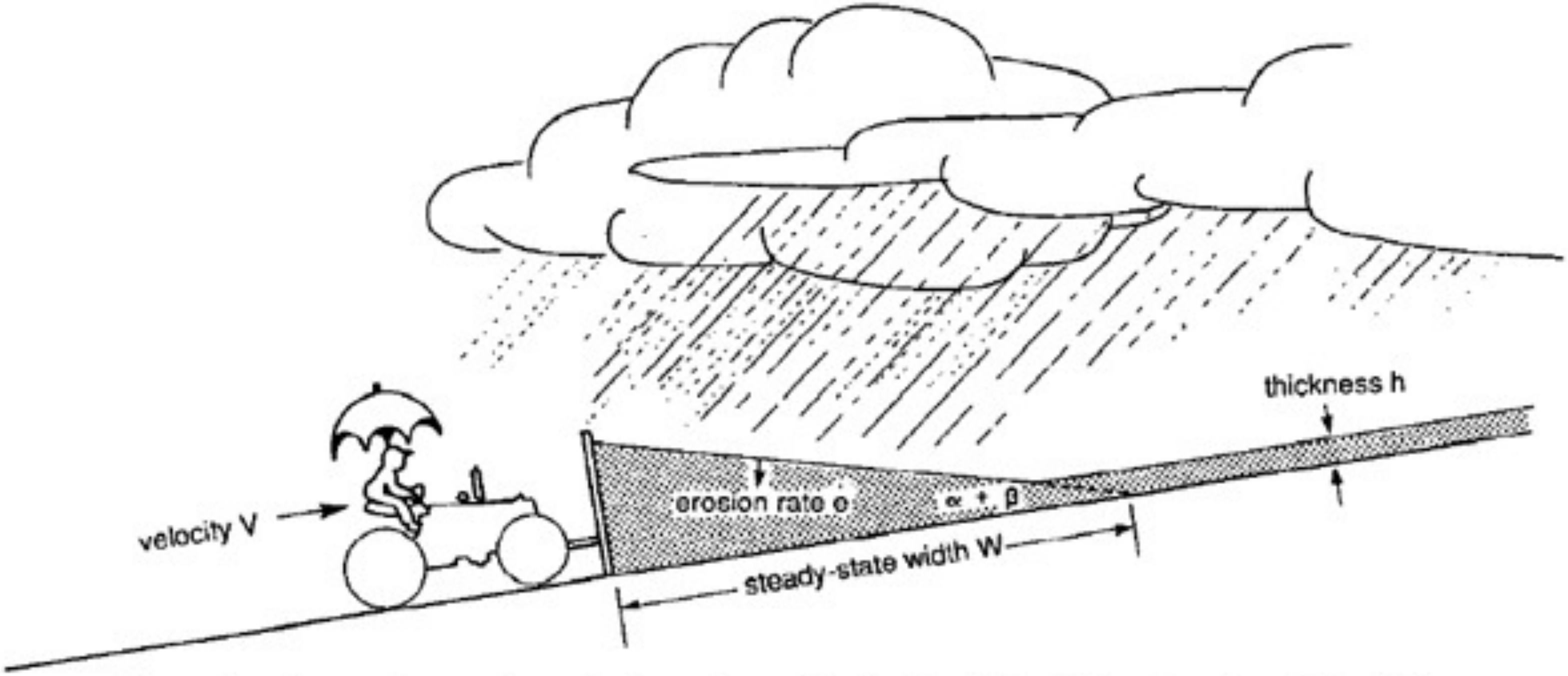


Figure 5 An eroding wedge attains a dynamic steady-state width given by $\dot{e}W = hV$.