

A New Conceptual Model for Understanding Geographical Variations in Weathering

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Weathering is the breakdown and decay of earth materials *in situ*; differences in weathering from place to place thus play a key role in defining the character of a landscape. The prevailing theory of geographical variations in weathering is based on coarse visual observations of correlations between climatic regions and weathering characteristics. We propose an alternative perspective that interprets weathering's spatial variability as a function of microscopic processes. In the new model, variability in weathering involves synergistic interactions of biological, chemical, and physical factors acting at the weathering boundary-layer. In testing these different models, we examine weathering processes in regions that provide the main evidence for the climatic model: cold and warm deserts. Contrary to the climatic theory which holds that mechanical weathering predominates in deserts and that dryland weathering profiles are shallow owing to the lack of water for chemical weathering, the available data suggest that these widely perceived generalizations do not apply at hygroscopic, capillary, pedogenic, landform, and landscape scales. A new perspective on weathering is thus in order.

Perhaps the oldest and most venerable theory of weathering states that climate influences the character of rock disintegration and decay (Middleton 1743 in Yatsu 1988; Ansted 1871). This theory won support in two core domains of physical geography: soil geography and climatic geomorphology. Perhaps the most influential model in soil geography is that of Hans Jenny (1941). Relying on ideas borrowed from the Russian school of soil science (Dokuchaev 1879; Glinka 1914), Jenny mod-

eled soil development as a function of *Climate, Organisms, Relief, Time, Parent material, and human interaction*. In the intervening half century, the "CIORPT" equation of soil development has become a core concept in interpreting spatial variability in pedogenesis (Amundson et al. 1994).

This climatic interpretation of weathering was incorporated likewise into the growing subfield of climatic geomorphology (cf. Tricart and Cailleux 1965; Derbyshire 1976; Büdel 1977). In the graphic models developed by Peltier (1950) (Figure 1a) and Strakhov (1967) (Figure 1b), climate determines weathering intensity (Figure 1a) and the depth of weathering profiles (Figure 1b). Focusing on the independent variables of mean annual temperature and precipitation, Peltier (1950) accentuated linear relationships between temperature/precipitation and climatically based weathering regions (Figure 1a). Extending Peltier's ideas to weathering profiles, Strakhov (1967) noted that these are deep in the wet/warm tropics and wet/cool midlatitude forests and shallow in dry/warm deserts and dry/cold tundra (Figure 1b).

The theory of mesoscale climate as the major control of geographical variability in weathering is widely accepted among scholars. The notion that weathering regions can be derived from mean annual temperature and precipitation offers a case in point. Diagrams that link climate and weathering are reproduced in specialized (for example, Loughnan 1969; Ritter 1978; Price 1981; Chorley et al. 1984; Ollier 1984; Selby 1985) and generalized texts (for example, Scott 1992; de Blij and Muller 1992; Christopherson 1992; McKnight 1993). In the more technical literature on weathering, these

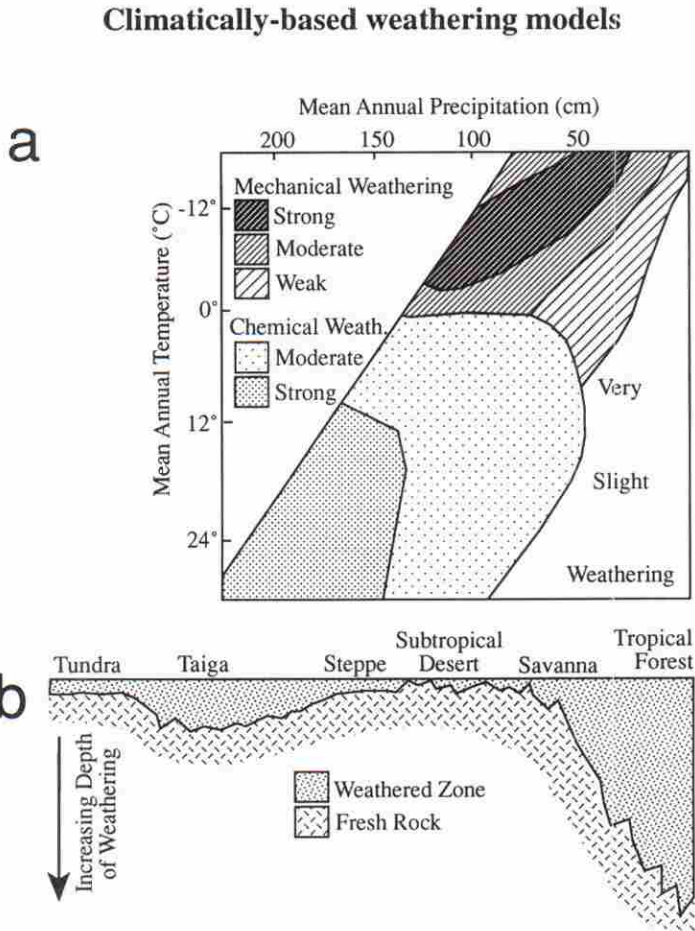


Figure 1. Diagrams illustrating current models of dryland weathering and climate. a) Relative importance of different types of weathering in relation to mean annual rainfall and mean annual temperature. According to Peltier (1950), mechanical weathering predominates over chemical weathering in drylands. b) Depth and structure of the weathering mantle along a Northern Hemisphere latitudinal transect, with corresponding major biomes. According to Strakhov (1967), the weathering mantle is thinnest over Tundra, Polar Desert, and Subtropical Warm Desert.

climatically defined boundaries are transformed into acidity or alkalinity weathering regions (for example, Pedro 1968; Pedro and Sieffermann 1979; Gibson et al. 1982:259; Ugolini 1986:224; Evans 1992:119; Chesworth 1992:34; Nesbitt and Wilson 1992:771). But perhaps the best example of the entrenchment of mesoscale climate theory is the analytic expectation that tropospheric climate exerts the controlling influence over weathering (cf. Gib-

son et al. 1982:259; Ollier 1984:172; Ugolini 1986:224; Young and Saunders 1986:10; Watson 1992:234; Velbel 1993:1061).

One reason for the persistence of the climatic theory has to do with our perceptions of the spatial variability in weathering. Notwithstanding that weathering processes operate at the *microscopic* scale, via hygroscopic, capillary, and gravity pore water (Walder and Hallet 1986; Chesworth 1992), the *visual* differences

in weathering features seem to fit so closely with readily observed differences in climate (Figure 1). This apparent correlation between climate and weathering seems especially evident in cold and warm drylands. In warm and cold deserts, mechanical weathering is thought to predominate over chemical weathering owing to the paucity of moisture or lack of heat energy for driving chemical reactions (Peltier 1950). Similarly, the shallow depth of weathering in drylands is thought to be a function of limited heat in periglacial landscapes and limited water in warm deserts (Strakhov 1967). Textbook images of angularly fractured bedrock from "mechanical weathering" in cold and warm drylands reinforce the perceived links with climate.

Contrary to these arguments, we contend that the spatial variability in weathering reflects processes at the microscopic scale. Although climate plays an important role, the intuitively grounded theories linking tropospheric meso-scale climates and weathering are oversimplified. This paper presents a new conceptual model of geographical variability in chemical weathering and compares this model with the climatic theory of weathering. Our evaluation progresses systematically from micro- to meso-scales, from the hygroscopic mineral boundary-layer to landscapes. Most of our examples derive from the warm and cold deserts which have provided the climatic theory with its best evidence.

A Weathering Boundary-Layer Model

The model of weathering presented here assumes that the processes responsible for the spatial variability in weathering operate principally at the weathering boundary-layer, i.e., the rock-atmosphere-hydrosphere-biosphere interface. This process-based model, which follows the works of Jenny (1941), Reiche (1945), Pedro (1968), Chesworth (1992), and Lasaga et al. (1994), is a multivariable, open-system approach in which no single factor dominates in a multiscalar geomorphic system. Accordingly, the weathering rate (W_r) is a function of several synergistic factors:

- (R_a) availability and proximity to abiotic weathering agents (abiotic reactant chemistry);
- (R_b) availability and proximity to biotic weathering agents (biotic reactant chemistry);
- (C_p) mineralogy and petrology (chemistry) of the parent material;
- (C_e) mineralogy (chemistry) of evolved products;
- (L) lithology of the parent material;
- (S_t) structure (jointing, foliation, bedding) of the parent material;
- (T) temperature at the reaction site;
- (H) hydraulics of water movement;
- (-M) removal of weathered material via gravity water or capillary water;
- (e) eolian input or deflation of organic and inorganic fines;
- (μτ) microtopography of the land surface;
- (A_s) exposed surface area;
- (A_c) accreted coating area (armoring);
- (s) scale;
- (t) time;
- (ΔE) environmental change over time, including biota, climate, landscape, erosion, and weathering products; and
- (. . .) variables still undefined, including anthropomorphic influences;

such that:

$$W_r = f(R_a, R_b, C_p, C_e, L, S_t, T, H, -M, e, \mu\tau, A_s, A_c, s, t, \Delta E, \dots) \quad (1)$$

In sum, we advocate a microscopic, multivariable, open-system perspective on weathering within a multiscalar geomorphic system.

Evaluating "Climate" versus "Weathering Boundary-Layer" Models

Our evaluation proceeds along two lines: the systematics of weathering processes and the role of different scales. In this section, we show how a weathering boundary-layer model can link these processes and their spatial variability.

The weathering rate (W_r) function offers considerable flexibility in sorting out the effects of climate on weathering. For example, bio-

chemical acids (Avakyan et al. 1981; Cooks and Otto 1990; Drever 1994; Wasklewicz 1994) are related more to microclimate than to mesoscale temperature or precipitation. The effects of microclimate may vary, however, from place to place, depending on such variables as slope and aspect, surface microtopography, surface thermal properties, vegetation, and local moisture sources. These variables are significant for microscale energy balances (Paradise and Yin 1993) and weathering activation energies (Hodder 1984).

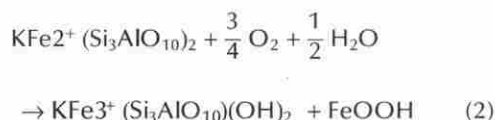
Water serves multiple microscale functions in weathering (Chesworth 1992): 1) as an agent of transport where capillary forces drive microscale movement and gravitational and evapotranspiration potentials drive macroscale movement; 2) as a medium through which reactants diffuse at reaction sites; 3) by exerting a partial pressure; 4) by aiding in mechanical weathering (Walder and Hallet 1986) in increasing surface area; 5) as a chemical solvent; 6) as a constituent of the products of weathering; 7) as a chemical buffer; and 8) as a needed part of chemical reactions (e.g., hydration, hydrolysis, ion-exchange). The magnitude of these weathering functions are not especially well defined by the level of precipitation in a rain gauge.

The flexibility of our model reflects the incorporation of Reiche's (1945:49–50) "weathering potential index" which maintains that primary rocks progress along different paths of evolved products. For example, the combination of C_p , C_e , R_a , R_b , $-M$, e , and T in equation (1) might be thought of as reactivity coefficients in an equation of thermodynamic weathering potential. These variables may increase weathering potential such as when increasing temperature (T) increases the efficiency of chemical reactions; or when the mobility of weathered material ($-M$) is influenced in turn by soil water hydraulic processes (H) (Swoboda-Colberg and Drever 1993); or when agents conducive to chemical reaction are present, such as potent reactants (R_a , R_b) and unstable parent materials (C_p) in the surface environment (e.g., biotite) which have a high potential for chemical weathering. Similarly, environments with high biomass promote chemical weathering through the addition of organic compounds (R_b) and the increase of weatherable surface area (A_s) through physical root action (Drever 1994).

Conversely, other variables may slow the process of weathering. Evolved materials (C_e) such as kaolinite which approach chemical equilibrium with the surface environment tend to be resistant to chemical weathering. Minimal surface area (A_s) and accreted coatings (A_c) may retard weathering potential. Lithology (L) and structure (S_t) also affect water penetration via porosity and jointing. The input and output of eolian material (e) may also modify reactivity by adding airborne salts, carbonates, clays, and silts to the soil, or by winnowing away the fine materials produced by weathering.

Shifting our perspective on weathering variability toward the microscopic scale has the added advantage of resolving confusions between process and form, and between weathering and erosion. Consider "etching," a commonly observed weathering phenomenon (Locke 1979; Brantley et al. 1986; Hall and Horn 1993). While etching is usually considered as a process, it is actually a weathering form (Figures 2b, 3b, 3e, 3f). The difference is not merely semantic. Etching is a form defined by the net loss of ions at a mineral surface that is the result of several different processes, for example hydrolysis or iron oxidation of different minerals.

Another case in point is the confusion over biotite weathering as a product of chemical or physical weathering. Geomorphology and soil texts typically present biotite weathering as an important chemical weathering reaction that breaks down granitic rocks (cf. Farmer et al. 1971):



The argument is as follows: chemical weathering of biotite leads to the production of grus which leads to spheroidal weathering of granitic rocks. But in fact, the decay of biotite is more complex. It may result either from the loss of octahedral iron (Figures 2a and 3a), from mechanical stresses imposed by the reprecipitation of iron hydroxides (Figures 2d and 3g), or from the joint operation of these chemical and mechanical processes. Similarly, the distinction between weathering and erosion often disappears at the microscale perspective. Hydrolysis, for instance, may weaken

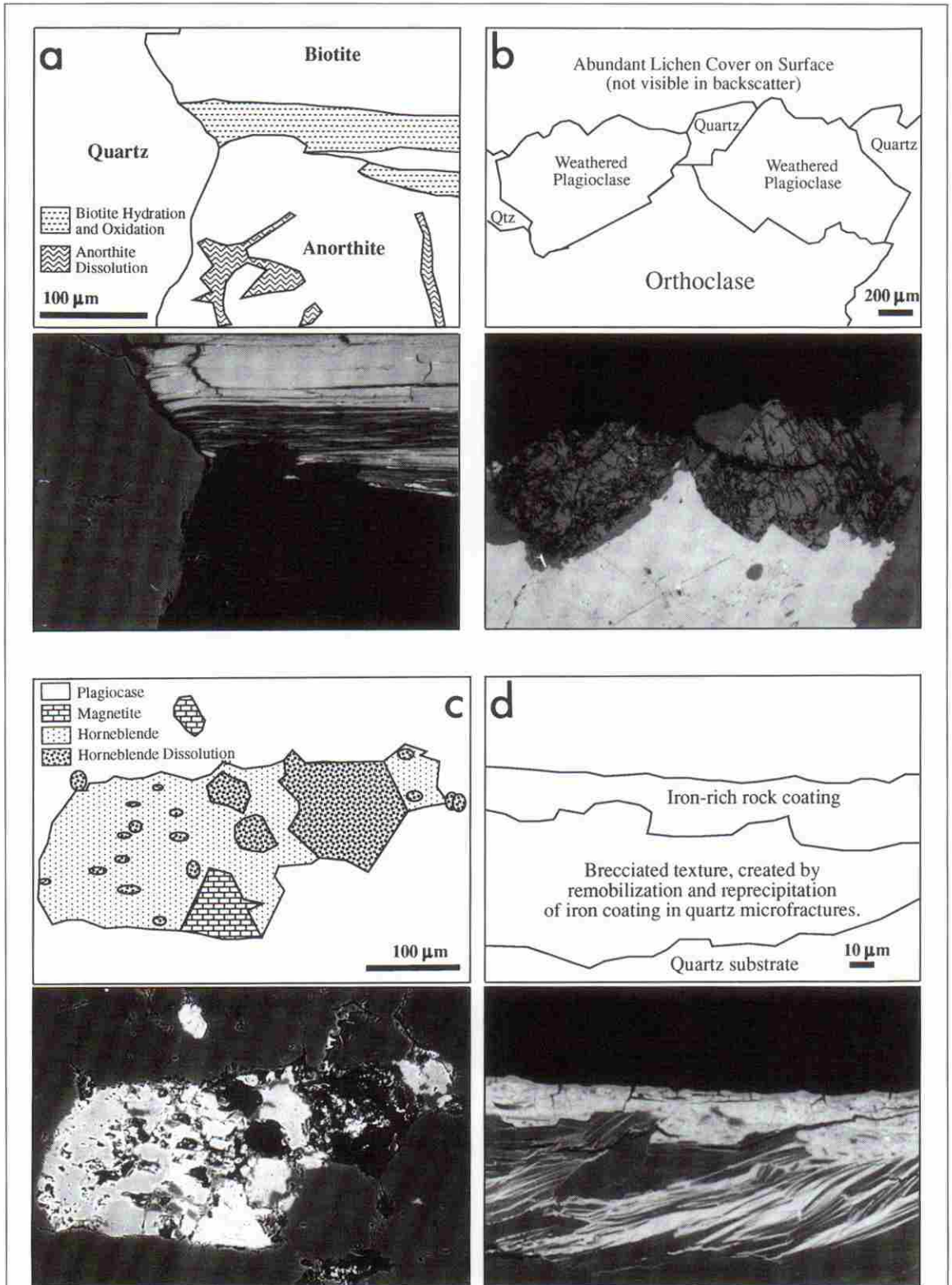


Figure 2. Chemical weathering in cold deserts. Electron micrographs are images with backscatter electron (BSE) (continued on next page)

mineral structures (weathering) but it also may remove ions (erosional transport). In this case, a microscale chemical weathering process is responsible for breaking down minerals and then eroding them.

A microscopic model for spatial variability in chemical weathering accents the synergism among biological, chemical, and physical processes and the interplay between environmental and mineralogical factors in weathering. Consider the case of pyroxene dissolution in basalt (Figure 4a). This dissolution begins as a chemical process which is then enhanced by physical expansion (Wasklewicz 1994). Likewise, gradients of chemical weathering rates, even in tropical basalts, are enhanced by mechanical processes involving microfissures (Schirmer and Störr 1994). These synergies are also evident in biotite weathering which can be initiated by the mechanical process of quartz silt being wedged into mineral weaknesses (Figure 4b). When salt precipitation at the ground-line on basalt hydrates/dehydrates and thermally expands/contracts (Evans 1970), the result is a synergistic combination of chemical and physical weathering (Figure 4c) that enhances erosion around the base of a rock. Or, if one needs another example, consider the weathering of quartz via the mechanical precipitation of salt which enhances dissolution interactions with chloride, as in Figure 4d where barium chloride is precipitated (Young 1987; Xie and Walther 1993).

We are not advocating that theories emphasizing the importance of climate in weathering be discarded but rather that the role of climate in the complex process of weathering needs to be reconceptualized and, more precisely, rescaled. An example of these complex relations comes from the case of climate's impact on weathering and pH. Dissolution rates for two of the most common mineral groups

found on the earth's surface—silicates and oxides—vary greatly with pH (Casey and Sposito 1992). Dissolution rates, as measured by experimental activation energies (E_{exp} , measured in kJ/mol), are increasingly dependent upon pH at higher temperatures. According to the Arrhenius relation in Brady and Walther (1990):

$$R_{\text{dis}} = Ae^{-E_{\text{exp}}/RT} \quad (3)$$

where R_{dis} is the experimental dissolution rate ($\text{mol}/\text{m}^2/\text{s}$), A is a frequency of collision factor ($\text{mol}/\text{m}^2/\text{s}$), T is absolute temperature (K), and R is the molar gas constant ($\text{kJ}/\text{mol}/\text{K}$). This equation, albeit a generalization, reveals the functional relations between higher pH values (reflecting acid concentration and therefore higher A), higher temperatures, and enhanced dissolution rates.

In our opinion, the crux of weathering-climate relationships rests not in tropospheric measures of climate, but rather with the microclimate that prevails at the sub-millimeter hydraulic boundary layer. Weathering's thermal environment (T) is determined by energy balances that prevail over meters to millimeters, and with geothermal gradients, exothermic and endothermic chemical reactions, insulation and thermal inertia provided by organic or unconsolidated surface materials and soil moisture, and microtopographical variations ($\mu\tau$) affecting radiation balance. Similarly, the availability of hygroscopic and capillary water (contributing to $-M$)—a function of soil hydraulics (H) and erosion potential (Swoboda-Colberg and Drever 1993)—depends indirectly on controls of precipitation and climate on vegetation. Thus, the role of climate is an indirect one in mediating boundary-layer temperatures, erosion, and the movements of weathered material, and in providing weathering agents.

Weathering occurs over long time intervals

Figure 2. (continued) microscopy in which brightness is proportional to average atomic number. a) While quartz is relatively resistant, biotite hydrates and the feldspar dissolves; from weathering rind of granodiorite glacial erratic, Cottonwood Plateau, White Mountains of California-Nevada (for context see Elliott-Fisk 1987). b) Underneath a crustose lichen, plagioclase dissolves along structural weaknesses, while orthoclase and quartz are less altered; from glacial polish at 5800 m above Everest basecamp, Nepal. c) Hornblende is dissolving more rapidly than the adjacent plagioclase in this sample from the Clark Mountains, Marie Byrd Land, Antarctica (collected by P. Yeats, Texas Tech University collection). d) An iron-rich rock varnish has dissolved and reprecipitated in the quartz substrate; from Karkevagge, Swedish Lapland, at an elevation of -1500 m.

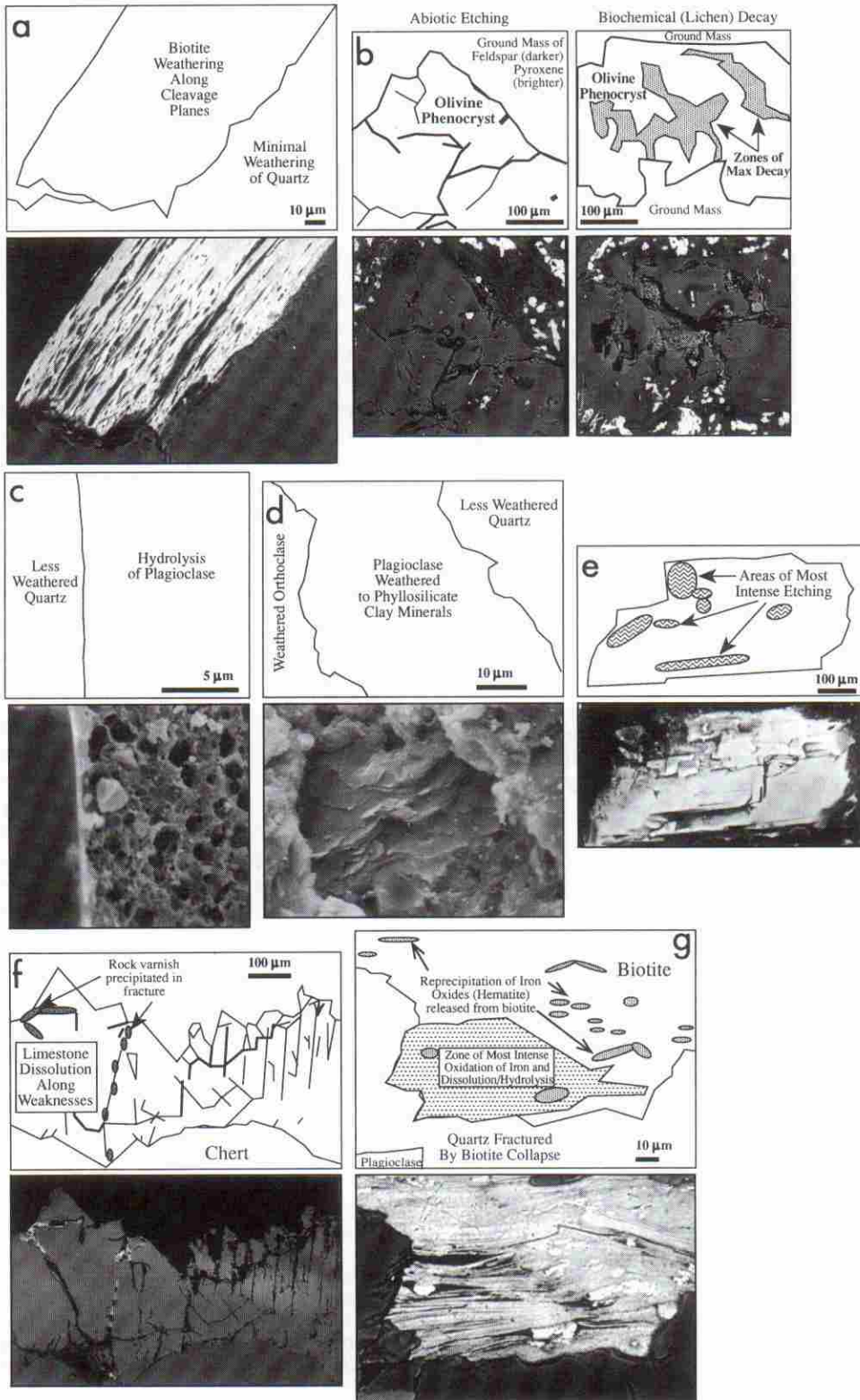


Figure 3. Chemical weathering in warm deserts. Images c and d based on secondary electron microscopy (SE) (continued on next page)

(Vasconcelos et al. 1992; Ollier 1992). Therefore, theories of weathering must also allow for environmental change (ΔE): climates fluctuate; parent materials evolve; ecosystems change; the entire physical geography system is in flux. Although ΔE and t add complexity to the weathering equation, their addition enables other variables to be treated in a dynamic fashion, contrary to static weathering models based on the correlation between current climate and current weathering states. The latter clearly cannot account for the kinds of extreme climate changes that occurred globally during, for example, the Quaternary.

Chemical Weathering at Multiple Scales in Dryland Environments

This section compares the "climatic" and "weathering boundary-layer" conceptual models using examples taken primarily from drylands. These examples suggest that the climatic explanation of spatial variability in dryland weathering (in both cold and warm deserts) is less persuasive than the weathering boundary-layer hypothesis. In doing so, we begin by reviewing the claims for climatic weathering and some of the reservations expressed about overly simplified climatic generalizations.

The climatic theory of weathering rests on intuitively appealing observations. Consider the following:

Mechanical processes of weathering dominate under conditions of low rainfall. (Brady 1984: 395)

In arid regions, whether hot or cold, there is likely to be less chemical weathering. (Ollier 1984: 172)

Adverse conditions in cold and polar deserts are unfavorable for intense weathering, especially chemical weathering. (Ugolini 1986: 224)

... physical weathering dominates in drier, cooler climates. (Christopherson 1992: 371)

Weathering occurs extremely slowly in dry air. Water increases the rate of weathering many orders of magnitude for several reasons ... (Manahan 1993: 502)

Observations of this sort are commonplace in teaching and research (cf. also Loughnan 1969:70-71; Butzer 1970:379; Price 1981:172; Chorley et al. 1984:469; Watson 1992:234). Geographic interpretations of weathering continue to depend on these general climatic models of weathering (Figure 1).

But the evidence does not support these models nor their assumptions that mechanical weathering predominates over chemical weathering in warm and cold deserts (Figure 1). One of the first weathering specialists, Parry Reiche (1945:15), concluded that "physical weathering processes are of secondary importance" even in areas subject to extremes of frost or heat. Chemical reactions can, in fact, proceed at temperatures well below freezing (Tamm 1924), and solution processes are significant in cold alpine environments such as the European Alps (Jackli 1956) and Sweden (Rapp 1960). As the dean of arid geomorphology, Elliot Blackwelder (1954:12), has observed: "Contrary to beliefs prevailing a few decades ago, it now seems well established that chemical processes are vastly more important than purely physical changes, such as temperature variations, in the weathering of sound rocks in a desert climate." Many other scholars affirm the efficacy of dryland chemical weathering, and their studies are listed in Table 1.

Figure 3. (continued) depict topography; all others are produced by backscatter electron microscopy (BSE). a) Biotite weathers by dissolution along parallel crystal weaknesses, while quartz is relatively stable; from Cabo San Lucas, Baja California. b) Contrasts in biotic and abiotic weathering in olivine; weathering is much more rapid under lichens. Both samples from a ca. 2 ka lava flow, Pinacate Volcanic Field, Sonoran Desert of northwestern Mexico. c) Plagioclase weathering creates micropores and clays, whereas quartz is relatively unweathered; central Sinai Peninsula, Egypt. d) Plagioclase weathers to clay minerals; Dunhuang region in western China. e) Hornblende "etchpits" are common in the B-horizon of soil from the Tioga (~19 ka) moraine, Pine Creek, eastern California. f) Limestone preferentially dissolves along structural weaknesses; from desert pavement near Sde Boquer, Negev Desert, Israel. g) Biotite weathers by the mobilization and reprecipitation of iron oxides (brightest areas); from Spook Hill pediment in the Phoenix, Arizona, metropolitan area.

Table 1. Selected Studies Emphasizing Different Chemical Weathering Processes in Warm and Cold Deserts.

Weathering Process	Region	Study
Iron oxidation	Canadian Arctic	Isherwood 1975; Locke 1979; Watts 1981; Campbell and Claridge 1992
	Antarctica	Kelly and Zumberge 1961; Ugolini 1986; Campbell and Claridge 1992
	Pacific N.W. ranges	Reynolds and Johnson 1972; Dixon et al. 1984
	Rocky Mtns., U.S. Basin and Range	Dixon 1983; Hall and Horn 1993
	Hawai'i warm deserts	Marchand 1971; Conca and Rossman 1982; Nesbitt and Wilson 1992; Wasklewicz 1994; Blackwelder 1954; Conca and Rossman 1985
Hydration	Canadian Arctic	Isherwood 1975; Watts 1981
	Pacific N.W. ranges	Dixon et al. 1984
	Rocky Mtns., U.S. Basin and Range	Pierce et al. 1976; Dixon 1983
	Middle East	Conca and Rossman 1982; Pope 1992
	African Rift Valleys	Blackwelder 1954; Amit et al. 1993
	Atacama Desert	Ericson et al. 1986
	warm and cold deserts simulated hot desert	Lynch and Stevenson 1992; Evans 1970; Doornkamp and Ibrahim 1990; Smith and McGreevy 1988; Goudie 1993
Solution	Arctic	Smith 1976; Pecher 1994
	Sweden	Rapp 1960
	Alps	Jackli 1956
	Rocky Mtns., U.S.	Caine 1976; Caine and Swanson 1989; Stednick 1989
	Atlas Mtns. various cold regions warm deserts	Smith 1986; 1988; Corbel 1959; Dredge 1992; Blackwelder 1954
Silicate dissolution (hydrolysis)	Antarctica	Boyer 1975; Ugolini and Jackson 1982; Ugolini 1986
	Canadian Arctic	Bockheim 1979; Locke 1979
	Norway	Ellis 1980; 1983; Darmody et al. 1987
	Pacific N.W. ranges	Reynolds and Johnson 1972; Dixon et al. 1984
	Rocky Mtns., U.S.	James et al. 1982; Dixon 1983; 1986; Thorn et al. 1989
	Baja California	Conca and Rossman 1985; Graham and Franco-Vizcaino 1992
	Basin and Range lab simulation	Marchand 1971; Butler and Mount 1986; Goudie 1993; Xie and Walther 1993
Biochemical processes	South Africa	Cooks and Otto 1990
	S.W. North America	Whitford and Freckman 1988
	Middle East	Danin et al. 1983; Shachak et al. 1987; Paradise 1993
	warm and cold deserts	Avakyan et al. 1981; Friedmann and Ocampo-Friedmann 1984
	paleoenvironments	Schwartzman and Volk 1991; Danin 1985

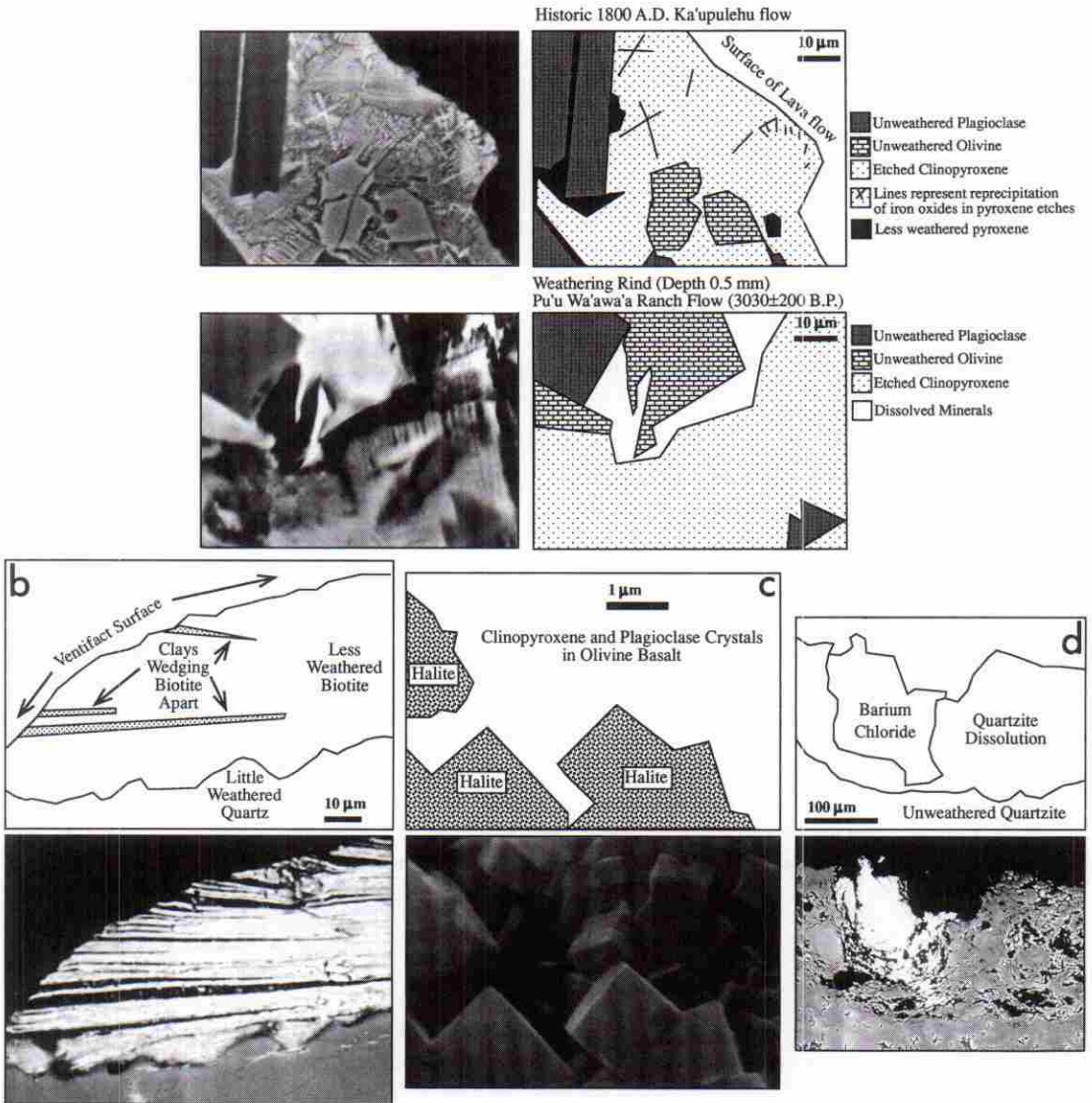


Figure 4. Synergism between chemical and physical weathering. Images a, b, and d are produced by BSE, image c by SE. a) A time progression of clinopyroxene weathering from ca. 190 years to 3000 years; from weathering rinds on the arid rainshadow of Hualalai Volcano, Hawaii. The olivine and plagioclase remain relatively stable, while clinopyroxene starts to weather after only 190 years by etching and reprecipitation of bright beads of iron oxides. After 3000 years, the etching has progressed to the point where pore spaces (from dissolution) have opened up. b) Silt- and clay-size fragments have been mechanically wedged into a biotite polished by eolian abrasion; from a ventifact on Bishop Creek moraines, Owens Valley in eastern California. c) NaCl precipitation on basalt fragments; from the ground-line band on "mushroom rock" in Death Valley, California. d) Barium chloride enhances the dissolution of quartz increasing the microporosity; from the 300 m marine terrace adjacent to the Rio Grande, southern Peru.

These studies and others suggest that the apparent paucity of liquid precipitation in cold and warm dryland environments is not an impediment to chemical weathering. That is because the water necessary for chemical weathering consists of hygroscopic films on particle surfaces, and these are ubiquitous even in warm deserts (Roth 1965). Consider the fate of a granite statue in Egypt. The statue, "... fallen on its side, has disintegrated and exfoliated where the rock has remained in contact with the ground in which moisture is more prevalent, while the upper side, exposed to almost continual sunshine, shows no signs of alteration even after 3,000 years. The breakage of rocks by changes of temperature is of negligible importance even in the hottest deserts" (Blackwelder 1954: 13). In this case, hygroscopic water weathers the mineral, while capillary water and hydraulics help to remove these weathering products away from reaction sites (Swoboda-Colberg and Drever 1993). In warm deserts, for example, weathering potential is greater under high temperatures, particularly in the presence of seasonably available capillary water and higher pH values (Casey and Sposito 1992). Similarly, weathering in periglacial climates depends on seasonally available summer precipitation for the transport of abundant solutes (Pecher 1994).

As the evidence against the climatic model of dryland weathering has mounted over the last century (Table 1), scholars have identified a variety of chemical reactions that operate in warm desert and periglacial environments. We now know that hydration, which is common with clays, micas, and salts (cf. Evans 1970; Isherwood 1975), also occurs in silicates (cf. Ericson et al. 1986; Lynch and Stevenson 1992; Pope 1992); that hydrolysis is prevalent from the polar deserts of Antarctica (cf. Ugolini 1986) to the subtropical deserts of Mexico (cf. Graham and Franco-Vizcaino 1992); that carbonate solution occurs from arctic (Smith 1976) and alpine (Rapp 1960) regions to warm deserts (Blackwelder 1954); that endolithic organisms contribute to biogeochemical weathering in even the harshest environments (Friedmann and Ocampo-Friedmann 1984); and that iron oxidation commonly occurs in a variety of dryland environments (Conca and Rossman 1985; Wasklewicz 1994).

The point of our argument is that the chemical weathering processes that operate in all

other environments also operate in cold (Peltier 1950) and warm (Roth 1965) drylands. It is now well-known that low average annual precipitation, desiccating heat, intense cold, and permafrost do not prevent chemical reactions. The ubiquity and efficacy of these processes in drylands are evident at various scales.

The Hygroscopic Scale

The water that enters the weathering system as atmospheric precipitation subsequently reacts with earth materials in the form of hygroscopic, capillary, and gravity water. Hygroscopic water binds tightly to mineral surfaces (Martini 1981), and it is *ubiquitous* on minerals in all terrestrial weathering surfaces. In drylands, Casey et al. (1993:8) report that natural weathering proceeds in "very small channels ... on the order of ~ 30 angstroms in diameter." The best way to view effects of hygroscopic water is by high resolution transmission electron microscopy, a method whereby crystallinity is imaged at the nanometer-scale. The microscopic image in Figure 5a illustrates that quartz from semiarid Owens Valley in California becomes amorphous when hygroscopic water is incorporated (Pope 1992), while Figure 5b illustrates that clay mineral weathering can occur even in one of the earth's most desiccating environments—in rock varnish on the barren side of a nunatak in Antarctica. The varnish matrix in Figure 5b is composed of illite/montmorillonite clays, which start out with even parallel lattices, but progressively split and weather into monolayers (cf. Robert and Tessier 1992).

The key variables that determine the rate of weathering from hygroscopic water are mineral surface area and the hydrodynamic environment that physically transports reactants (Chesworth 1992; Friedman et al. 1994). The hydrodynamic environment (analyzed in the next section) involves capillary and gravitational water. In the cases of the surface areas of minerals exposed to hygroscopic water, these vary considerably from place to place. This is best illustrated by way of example. The surface area exposed to hygroscopic water is the same for a classic landform such as a bornhardt dome with a surface area of ~ 1 km², a 60 m by 60 m by 120 m talus ramp composed of 1 m³ blocks, and 0.67 m³ of a debris flow

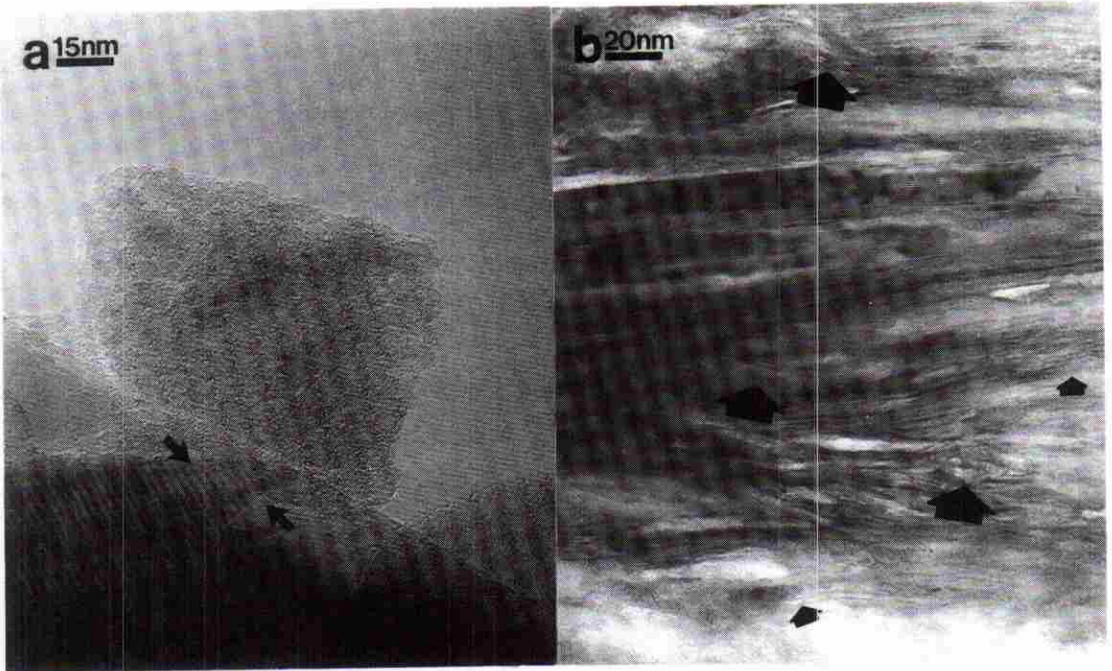


Figure 5. High Resolution Transmission Electron Microscope imagery of dryland weathering. Scale bar in nanometers. a) Quartz has become amorphous along crystalline defects. The unweathered, crystalline quartz is characterized by the regular lattice arrangement (arrows). In contrast, the amorphized zones are disorganized along zones of structural weakness, probably from the addition of water. Sample from Sherwin till in the semiarid Owens Valley of eastern California. b) Rock varnish is composed of over half clay minerals; this imposes a layered texture even at the hygroscopic scale. As manganese and iron oxides precipitate and cement the clays together, clay minerals weather by splitting into monolayers (arrows); from Mt. Van Valkenburg, Clark Mountains, Marie Byrd Land, Antarctica.

with a loamy silt matrix. As these examples suggest, weathering of exposed bedrock surfaces in warm and cold deserts proceeds slowly because less surface area is exposed to hygroscopic water, and not because of the impotence of chemical weathering in these drylands. We will return to this theme of slow weathering of exposed bedrock in the section on weathering at the pedogenic scale.

The Capillary Scale

The flow of water through soil and rock pores involves capillary and gravitational forces (Chesworth 1992). Capillary water is critical to mineral weathering because it serves as the medium through which reactants diffuse at reaction sites. As a mineral dissolves and surface

area increases, new pathways are opened up for capillary water. The role of capillary water in weathering depends on hydrodynamic flow, which depends in turn on the volume of water, the texture of the weathering materials, position in a catena, and location with respect to the water table. It also depends on the size and interconnectivity of the capillary pores which can control the length of time that capillary water remains in contact with mineral surfaces.

Processes involving capillary water are best viewed by scanning electron microscopy (SEM), a method which images weathering at the micrometer scale. SEM records images of secondary electrons (SE) and backscattered electrons (BSE). The SEM in secondary electron mode reveals the microtopography of the mineral surface and has been used extensively by geographers for studying weathering, soils, and

environmental change (see Krinsley and Doornkamp 1973; Mahaney 1990; Tchakerian 1991). BSE, though used less frequently for studies of weathering, is an ideal tool for doing so. Images formed by BSE reveal variations in mineral composition as measured by the backscattered-electron yield—a function of the average atomic number (\bar{Z}) of the sample. The various shades of gray in the BSE image represent varying elemental compositions within the sample; regions with lower \bar{Z} appear darker and regions with higher \bar{Z} appear brighter. In addition, microporosity from dissolution is imaged as black areas which BSE can distinguish to a micropore resolution of $\sim 0.1 \mu\text{m}$ (Krinsley et al. 1993). Virtually every dryland sample (out of thousands of observations over the last five years of research) that we have analyzed with BSE reveals the ubiquity of chemical weathering.

Figure 2 presents four BSE images from cold environments. These images depict the different ways whereby chemical weathering processes open up capillary pores and expose surface area to hygroscopic water. Figure 2a depicts the dissolution of anorthite and the opening up of cavities; these contrast with the long, linear pores opened by the hydration/oxidation of biotite. Figures 2b and 2c illustrate differential rates of mineral weathering in different microenvironments (cf. Wasklewicz 1994). In Figure 2b, lichen-covered plagioclase weathers along structural weaknesses while quartz and orthoclase are relatively unweathered. Figure 2c shows the result of plagioclase weathering when lichens are not present; plagioclase is in this case relatively stable while hornblende dissolves rapidly. Figure 2d illustrates the fracturing of quartz by chemical and mechanical weathering processes. The process begins with the precipitation of an iron-rich coating (microprobe analyses: 76–79 percent FeO, 0–5 percent MnO, SO_3 , P_2O_5 , Al_2O_3 , SiO_2); this is followed by remobilization of iron oxides, and subsequent reprecipitation in quartz microfractures. This synergy of physical and chemical weathering produces a brecciated texture of silt-sized quartz wedges.

The series of illustrations in Figure 3 illustrate various chemical weathering processes in warm deserts. These illustrations reveal, among other things, that the same mineral can weather in very different ways. Biotite, for example, can dissolve (Figure 3a) or undergo hy-

drolysis, hydration, and reprecipitation of iron oxides (Figure 3g). Weathering rates may be enhanced by the secretion of organic acids, as when lichens hasten the weathering of olivine (Figure 3b) (cf. Wasklewicz 1994). Chemical weathering may also occur along lines of structural weakness as in the cases of limestone dissolution (Figure 3f) and the etching of hornblende (Figure 3e). Chemical weathering is also evident in the hydrolysis of plagioclase which produces clay minerals (Figures 3c and 3d).

Weathering Mantles and Soils at the Pedogenic Scale

The pedogenic scale involves the development of weathering mantles and soils. A key argument for critics of the role of chemical weathering in dryland and cold environments hinges on the shallow depth of weathering (Figure 1b). Their critique ignores, however, the widespread distribution of thick weathering profiles that can be found in warm and cold deserts. Soils in the glaciated regions of Scandinavia (Lundqvist 1988), Scotland (Le Couer 1989), and the Sierra Nevada of California (Schaffer 1993) have been strongly influenced by pre-glacial weathering. As Ollier (1988: 287) observes:

A common assumption about deep weathering is that it is associated with warm wet climates: the literature is full of references to "tropical deep weathering." This assumption is not justified. Some deep weathering may have occurred under tropical conditions, but there is no necessary connection. All that is required for deep weathering is deep groundwater and plenty of time.

Deep (>15 m) granitic *grus* profiles are widespread at elevations in excess of 2,500 m in Colorado's Rampart Range (Varnes and Scott 1967). In the now-arid metropolitan area of Phoenix, weathered granodiorite profiles are as ubiquitous as granitic pediments. In road cuts, some profiles are over 40 m deep. Thick regolith is also found in Australian drylands (Butt 1985; Bird et al. 1990). For weathering mantles that are produced over millions of years (Mabbutt 1988; Vasconcelos et al. 1992; Ollier 1992) by geomorphic processes that transgress climatic periods, the assumption that climate and regolith thickness are neatly correlated (Figure 1b) is fraught with error.

In accounting for the depth of regolith,

mesoscale climatic models too readily mix cause and effect (Figure 1b). From the standpoint of regolith, the efficacy of erosion is more important than the average climate. Instead of a slow rate of weathering, thin regolith may reflect simply an increase in erosion. When Gilbert (1877) identified deserts as "weathering-limited landscapes," he accented the role of erosion in removing weathered debris faster than it is produced. But erosion is largely a function of the lack of vegetative cover, not the impotency of weathering processes. Huntington (1907) long ago recognized that an arid climate that follows a humid climate will decrease vegetation cover on hillslopes and increase the erosion of weathered bedrock. Regolith's cyclic preservation in periods with a more extensive vegetation cover and its ensuing erosion when vegetation cover is less extensive is now widely recognized as an important factor in shaping desert hillslopes (Bull 1991).

For weathering, the accent should fall not on the depth of regolith but rather on the effects of vegetation on regolith. In a reverse feedback, the accumulation of soil (by the combined products of biomass and weathering) may actually retard chemical weathering over time. Rapidly forming soils under dense tropical vegetation act as a shield against further chemical weathering of bedrock, while chemical weathering rates remain high in "weathering-limited" periglacial regions with thin soils (Bluth and Kump 1994). As Bluth and Kump (1994:2355) summarize: "... optimum climatic conditions [high temperature and precipitation] do not necessarily produce the highest rates of chemical denudation."

The preservation of a weathered mantle depends on other non-climatic factors as well. Consider the inverse relation between proximity to base level and tectonic stability and erosion rates. Tectonic stability in Australia reduces erosion rates and fosters thick regolith profiles (Fairbridge 1988; Mabbutt 1988; Gale 1992). Similarly, the deeply weathered granodiorite in the Phoenix region has been preserved by a protective combination of calcrete duricrust and slow rates of tectonic activity during the Quaternary.

Proponents of the mesoscale climatic hypothesis also argue that dryland soils are shallow, because of limited chemical weathering. But shallow soils may also reflect landscape

instability. Indeed, deeper soils in cold regions are prominent provided that they are protected from periglacial erosion processes. Soils with depths of a meter or more (and with an unknown depth to bedrock) have been reported in several polar desert environments including the Canadian Arctic (Bockheim 1979) and Antarctica (Gibson et al. 1982; Campbell and Claridge 1992). Both wet and dry tundra exhibit weathered clay accumulation and clay mineral migration (Dixon 1983; 1986; Darmody et al. 1987; Litaor 1988; Dixon and Thorn 1993). And iron and aluminum accumulation is common, which leads to the widespread incidence of Al-Fe humus soils (Birkeland 1978; Campbell and Claridge 1992). However accurate the belief that much of the clay fraction of polar desert soils is derived unaltered from the soil parent materials (Campbell and Claridge 1992), polar desert soils do contain a variety of weathering-derived clay mineral species such as kaolinite, vermiculite, illite, smectite, and hydrous micas (Claridge 1965; Boyer 1975; Bockheim 1979; Ugolini 1986).

As for hot, dry environments, soils there are commonly thick where landscapes are not eroding (Bull 1991). To be sure, much of this thickness reflects the addition of externally derived salt, silt, and clay (Mabbutt 1979; Holliday 1985; 1990; Amit et al. 1993), but there is also widespread evidence for *in situ* weathering of parent materials (Figure 3) including the oxidation of iron (Conca and Rossman 1985), the formation of secondary clay minerals (Nettleton et al. 1973; Folkoff and Meentemeyer 1987), the working of biochemical processes (Whitford and Freckman 1988), and the dissolution of primary mineral grains.

Weathering at the Scale of Landforms and Landscapes

Where erosion rates are faster than weathering, the weathered mantle is transported and bare rock is exposed in "weathering-limited landscapes." In contrast, where weathering rates are faster than erosion, weathered material covers bare rock in "transport-limited landscapes" (Gilbert 1877). This classic perspective has two corollaries: 1) that weathering forms occur where a weathered mantle occurs; and 2) the type of weathering forms depend upon

whether mechanical or chemical weathering is faster. Where mechanical fracturing of rocks along angular joints is faster than chemical decay, geomorphologists interpret landforms such as *felsenmeer* and frost riven cliffs to imply that mechanical weathering is faster than chemical weathering. To be sure, much of the shattered rock that is commonly found in regs and talus in warm and cold deserts is due in part to mechanical weathering. Thus, textbook photographs of angular shattered rock with captions emphasizing mechanical weathering create a resilient image that resists revision, especially among those who have not worked in warm and cold drylands.

Yet the reverse is also true. Chemical weathering landforms are created by the slow erosion of weathered debris away from differentially weathered rock. Common examples of chemical weathering landforms include: pedestal rocks where erosion has removed weathered material from the base of a boulder (Twidale and Campbell 1992; Cooke et al. 1993); granitic core stones where decomposed rock has been transported away from spheroidally weathered boulders (Chapman and Greenfield 1949); solution features on silicate rocks (Butler and Mount 1986; Kejonen et al. 1988; Twidale 1984); rillenkarren on carbonate rocks (Smith 1986; 1988; Danin et al. 1983; Dredge 1992 (Figure 3f); honeycombed patterns in sandstone (Young 1987; Young and Young 1992); gnamma depressions in granitic rocks (Twidale 1982; 1984; Paradise and Yin 1993); polygonal cracking (Williams and Robinson 1989; Robinson and Williams 1992); desert pavement (Smith 1988; Amit et al. 1993); weathering rinds (Dennen and Anderson 1962; Colman and Pierce 1981); and exfoliation scales (Blackwelder 1954).

Were mechanical processes faster than chemical ones, these landforms would be relatively uncommon. The existence of chemically weathered landforms implies that chemical weathering occurs faster than mechanical weathering. Chemically weathered features can only persist and grow when rates of chemical weathering exceed mechanical breakdown. This is even true in periglacial regions (Lidmar-Bergstrom 1988), despite glacial erosion and active hillslope processes. In fact, recent evidence points toward the importance of chemical weathering as a precursor to the creation of many bedrock glacial landforms

(Hall and Mellor 1988; Lindstrom 1988; Lundqvist 1988; Peulvast 1988; Cloutis 1992; Peuraniemi and Pulkkinen 1993; Schaffer 1993). In these cases, the scoured erosional remnants are vestiges of weathering resistance, where valleys have been cut in paths of least resistance partially created by deep weathering.

Ancient weathering landforms are best preserved in warm deserts that are tectonically stable. Like the cold climate examples noted above, these forms persist either because the rates of weathering exceed the rates of erosion or because they are somehow protected from erosion (Brunsdon 1993; Griffiths 1993). For example, several different theories maintain that granitic pediments in deserts are partially or entirely due to a combination of deep weathering and slope processes (Oberlander 1974; Twidale and Bourne 1978; Twidale 1982; 1983). Bornhardts, a landform typically associated with humid subtropical and tropical regions, are also found in present-day deserts, and these have been interpreted as the product of deep weathering (Twidale and Bourne 1978).

Many chemical weathering landscapes may be "inherited" from past climates or are products of a sequence of climatic changes and ensuing landscape stability; carbonate duricrusts (Hutton and Dixon 1981), laterite duricrusts (Brimhall et al. 1988), and etchplains (Twidale 1983) are cases in point. The prevailing climatic theory of weathering (Figure 1), by contrast, ignores or glosses over climate change. Yet, we know that weathering inherited from past climates can amplify contemporary chemical weathering reactions by opening pathways for water transport. Deep subsurface weathering (Mabbutt 1988; Bird et al. 1990) also negates the activation of physical weathering mechanisms such as pressure-release; indeed in our studies of the ubiquitous granitic rocks of the Mojave and Sonoran Deserts we have never encountered pressure-release shells simply because the *grus* cannot support the stress/strain requirements. Pressure-release shells are only found where the weathered granite has been stripped away by erosion.

Although our discussion has emphasized landforms created by the slow erosion of differentially weathered rock, chemical weathering also helps define the character of fluvial and eolian depositional systems (Johnsson and

Meade 1990). The load of the stream is in large part a function of the material supplied by weathering. If weathering does not produce enough silt, for example, an arroyo side wall cannot maintain a vertical profile, and hence sandy dryland streams assume a very different character. Similarly, the eolian system depends on sediment availability. The breakdown of rock and sediment by weathering produces silt- and sand-sized particles that are suitable for eolian transport (Pye 1989; Lancaster 1990; Tchakerian 1991).

Measuring the Extent of Chemical Weathering

Our entire argument for the boundary-layer perspective on weathering has been qualitative to this point, and for a good reason. Because weathering is a slow process, precise estimates of the rates of rock decomposition are rare (Meierding 1993a; 1993b; Paradise 1993; Swoboda-Colberg and Drever 1993). Moreover, the rates that we do have are specific to particular lithologies and these are not comparable. And in cases where quantitative data on weathering are available, these are often in measurement units that cannot be compared. For example, the measure of the thickness of "weathering rinds" (Colman and Pierce 1981) cannot be compared with a measure of the change in hardness as measured by the Schmidt hammer (Cooke et al. 1993), or of solute loads (Pecher 1994), or of changes in the geometry of weathering forms such as tafoni (Young and Young 1992). But when comparable quantitative data are available, these do not (contrary to the prevailing theory of weathering) reveal a climatic signal.

Perhaps the most useful and certainly the most comparable data on chemical weathering come from transported weathering products, in general, and from solutes produced by chemical weathering in different climates, in particular. Here we assume that solutes are derived from chemical weathering of the parent material (eolian and other unconsolidated sediments and rock in the drainage basin), and hence that these data measure indirectly the chemical denudation of the landscape. Using the estimates of different erosional processes from eight climatic regions compiled by Saun-

ders and Young (1983) and Young and Saunders (1986) (Figure 6) and additional data not included in their review (Corbel 1959; Caine 1976; Monaghan et al. 1992; Meierding 1993b; Paradise 1993) we reanalyze these estimates of chemical denudation and weathering. For purpose of standardization, we analyze only the data derived from siliceous rocks (including metamorphic and igneous rocks and clastic sediments). Rates of chemical denudation, measured in Bubnuffs ($1B=1\text{mm}/1\text{ka}$), range from 4 to 8 B in polar and montane environments and from 3 to 66 B in semiarid and arid environments. By contrast, chemical denudation rates in temperate maritime, temperate continental, subtropical, and tropical humid environments report minima that are consistently lower than the rates in polar/montane and semiarid/arid environments. Saunders and Young observe that their data indicate "no clear relation with temperature or rainfall" (1983:486). Indeed, our analysis of silicate rocks, albeit based on a small sample ($n = 40$), reports a statistically insignificant difference ($p = 0.30$) between the different climate regions.

Our second comparison of mechanical and chemical weathering is based on silt- and clay-sized detrital sediments (Pye 1989). Once again, the solid sediment offers no clear climatic signal. Abundant clay minerals produced by weathering and found in Quaternary lacustrine deposits (Curry 1990) and Quaternary alluvium (Bull 1991) attest to chemical weathering in drylands. Similarly in the dryland Cheyenne River basin of Wyoming, weathered clay constituted from 10 percent to over 75 percent of the sediment loads of several streams (Rolfe and Hadley 1964). Similar quantities of suspended silt- and clay-sized sediments occur in drainage basins with vastly different climates; consider, for example, the sediment yield of the more xeric pre-dam Colorado River basin ($190\text{ t}/\text{km}^2/\text{yr}$) which compares favorably with the sediment yields of the much wetter Amazon ($195\text{ t}/\text{km}^2/\text{yr}$) and Orinoco ($152\text{ t}/\text{km}^2/\text{yr}$) basins (McLennan 1993).

Although chemical weathering in the tropics or humid midlatitudes (Macias and Chesworth 1992; Clemens et al. 1993) proceeds more rapidly than in warm and cold deserts, the data for assessing the quantitative differences are simply unavailable. But even these data are not indispensable for our main point, which is that

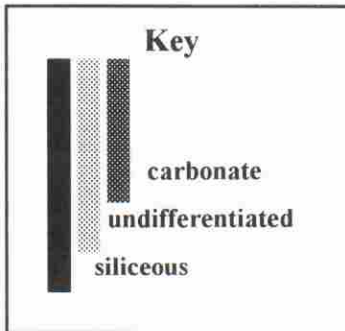
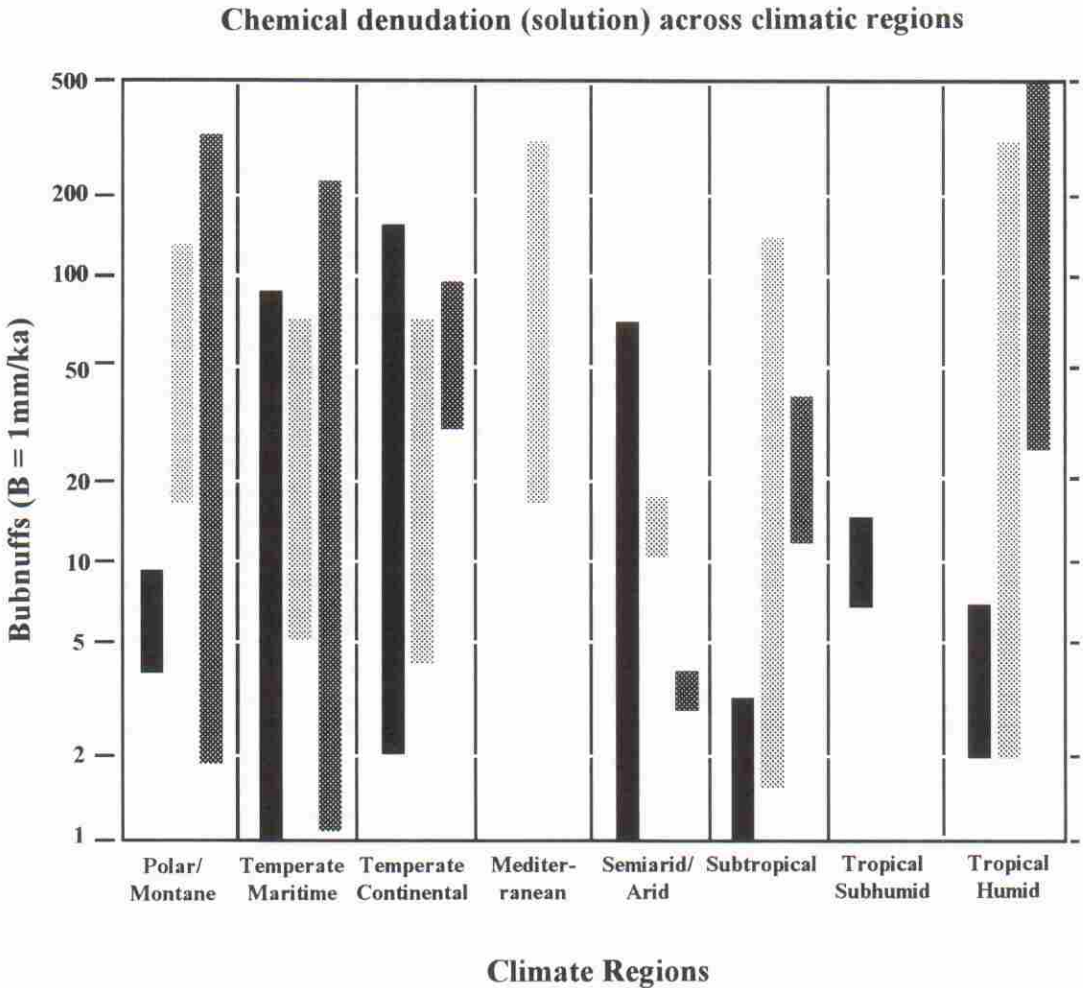


Figure 6. Variations in chemical denudation (solution) across climatic regions. The solution ranges for siliceous rocks, siliceous and carbonate rock undifferentiated, and carbonate rocks are remarkably similar across climate regions. The sample size for the statistical analysis of siliceous rocks by climate region: Polar/Montane, 3; Temperate maritime, 10; Temperate continental, 14; Mediterranean, 3; Semiarid and Arid, 4; Subtropical, 2; Tropical subhumid, 2; Tropical humid, 2. Source: modified from Saunders and Young (1983).

the net production (via weathering) and transportation of solutes and fine sediments is not a simple function of average climate—a host of other factors are involved (see equation 1).

Evaluating the Competing Models of Weathering Processes

We have shown that the climatic theory of weathering is inadequate in accommodating a variety of weathering data derived from dryland environments. In its stead, we have presented an alternative model of weathering focused on multiple factors operating at the boundary layer. As a final evaluation, we assess the two competing explanations of the geographic variability in weathering using widely-accepted criteria for adjudicating among competing hypotheses: 1) the quantity of data explained; 2) the diverse kinds of data explained; 3) the consistency of hypotheses with established theoretical frameworks and accepted theories; 4) predictive capabilities; 5) relevance; 6) plausibility; and 7) simplicity (Hempel 1966:33–46; Popper 1966:268; Newton-Smith 1981:226–230; Copi 1982; Farr 1983:165).

Quantity of Data Explained

The climatic model is based on assertions, rather than evidence; it claims that mechanical weathering predominates over chemical weathering in warm and cold deserts and that the intensity of chemical weathering varies in accordance with mesoscale climatic regions. The site-specific data that we do have demonstrate that chemical weathering processes are critically important in dryland weathering (Table 1; Figures 2–5). Similarly, cross-climate comparisons of solute loads reveal little if any statistical difference between different climates (Figure 6).

These site-specific and cross-climate data are consistent with the boundary-layer model of weathering at a variety of spatial scales. As high-resolution transmission electron microscopy reveals, hygroscopic water weathers minerals at the nanometer scale in warm and cold deserts. Similarly, at the capillary scale, backscatter electron microscopy demonstrates the ubiquity of chemical weathering in warm

and cold deserts. Turning to the pedogenic scale, we note that while advocates of the climatic model point to the shallow depth of weathering at this scale (Figure 1b), their argument ignores the widespread distribution of thick weathering profiles that are found in warm and cold deserts. To be sure, hillslope weathered debris in desert or periglacial environments exhibits shallow profiles, but these reflect the increase of erosion in dry periods and not the lack of chemical weathering—a point that is underlined by the thick weathering profiles in dry regions where tectonic and base-level stability slow erosion rates. Lastly, at the scale of landforms and landscapes, the high incidence of chemically weathered dryland landforms (e.g. pedestal rocks, core stones, bornhardts, etchplains) can only occur when rates of chemical weathering exceed rates of mechanical weathering.

Diverse Kinds of Data Explained

Switching to a microscopic perspective of weathering variability has the advantage of addressing certain false dichotomies between process and form, weathering and erosion, and chemical and physical weathering. A boundary-layer viewpoint also accents the synergism among biological, chemical, and physical processes and the interactions between environmental and mineralogical factors in weathering. Finally, the boundary-layer model forthrightly addresses the environmental changes that are inherent in geomorphic systems. Whereas the climatic theory visually matches contemporary climates with contemporary weathering patterns, the boundary-layer model can accommodate the fact that weathering occurs over millions of years and thus involves vastly different physical geographic systems at any given place and time.

Consistency of Hypotheses with Established Theory

The boundary-layer model, based as it is on the well-known works of Jenny (1941) and Pedro (1968), employs factors that are already familiar to students of weathering. The W_r function likewise incorporates Reiche's (1945)

notion of "weathering potential." Moreover, the boundary-layer model unites studies of the spatial variability of weathering with recent advances in low-temperature geochemistry (Brantley and Velbel 1993; Sverdrup and Warfvinge 1993; Swoboda-Colberg and Drever 1993; Lasaga et al. 1994), microenvironmental variability (Thorn et al. 1987; Caine 1992; Casey and Sposito 1992; Hall and Horn 1993; Brantley and Velbel 1993; Meierding 1993a; 1993b; Sverdrup and Warfvinge 1993; Paradise 1993; Wasklewicz 1994) and microstructural variability (Petit et al. 1989; Casey et al. 1993).

Predictive Capabilities

Doubtless mesoscale climates serve as *convenient* predictors of weathering regions for the simple reason that climate maps are readily available. But these predictions are, as we have seen, always crude and often wrong. What is needed is a theoretical link between processes at the molecular scale and the spatial variability in weathering forms and products, and that linkage is provided by the W_r function of the boundary-layer model. In sum, the climate theory of weathering does not provide accurate predictions of weathering and the boundary-layer model does via the spatial interpretation of weathering potential that may be predicted based on the W_r factors.

Relevance

The boundary-layer model has significance well beyond weathering research. An appreciation of weathering leads to a better understanding of our surface environment. Every individual who has scrambled over rocks or viewed spectacular scenery appreciates in some fashion spatial variability in rock weathering. The growing recognition of the importance of chemical weathering processes in geomorphology (for example, Caine 1992; Young and Young 1992; McLennan 1993) has been mirrored in such fields as biogeochemical cycling (Schwartzman and Volk 1991), geochronology (Danin 1985), paleoenvironments (Clemens et al. 1993), soils (Holliday 1990; Folkoff and Meentemeyer 1987), and geoaerchaeology (Paradise 1993). Advances in weathering have also been applied to studies of acid

rain (Meierding 1993a), CO_2 feedbacks in global climate regulation (Brady 1991; Velbel 1993), planetary geology (Gibson et al. 1982), and water quality. For example, the natural geographical variability of the toxic element mercury in the streams of the Colorado Plateau (Graf 1985) may be viewed as a function of the variability of W_r in our model. The boundary-layer model assessment of the geographical variability in rates of weathering may afford an "estimation of the rates of mineral-water reactions at watershed, continental and global scales [which] is of paramount importance in understanding the short- and long-term effects of natural and anthropogenic influences on the Earth system" (Brantley and Velbel 1993:vii).

Plausibility

Climate plays an important role in weathering processes, and the boundary-layer model accommodates it by translating mesoscale climatic variables to the microscale (sub-millimeter) variables that operate at the hydraulic boundary layer, i.e., at the reaction sites where weathering processes operate. The translation involves factors such as thermal properties, the input of weathering agents, and removal of solutes and particles. We suggest that these microscopic-scale factors offer a more plausible explanation of the geographical variability in weathering than do the ambiguous correlations of mesoscale climatic factors observed at the visual scale.

Simplicity

While the boundary-layer model may appear complex to those who are used to working with the climatic model, the former's simplicity is manifest in the way that it combines the diverse factors pertinent to the regulation of weathering rates. Stated in the most basic of terms, the factors that regulate W_r are simply weathering inhibitors and weathering promoters that together define weathering potential (Reiche 1945) over scale and time. While comprehensive in nature, the boundary-layer model can be viewed as a process of inputs and outputs within a systems perspective that is familiar throughout geomorphology (Chorley and Kennedy 1971).

The Need for a Strategic Shift in the Geographical Study of Weathering

A longstanding theory of geographic weathering variability assumes that climate controls the rates and processes of weathering. This theory rests on two widely held beliefs: 1) that mechanical weathering processes predominate over chemical processes in warm and cold deserts; and 2) that weathering depth is shallow in drylands (Figure 1). Proponents of the climatic thesis maintain that because warm and cold deserts lack the water that is necessary for chemical weathering processes, these processes proceed "slowly" and/or "insignificantly" as compared to mechanical weathering. This theory reflects the deep-seated perception, derived from *visual* observations, that the processes of weathering are largely a function of mesoscale tropospheric climate as measured at meteorological stations.

In contrast to this "top-down" visual perspective, proponents of weathering processes at the boundary layer emphasize the role of *microscopic* processes and observations, i.e., a "bottom-up" perspective. In subscribing to the latter view, we hold that geographic variability in weathering is controlled by variability in weathering processes that operate at the microscopic scale. We also hold that climatic models of weathering (Figure 1) introduce a series of erroneous assumptions on weathering processes: 1) that average tropospheric climate is identical to boundary-layer conditions where weathering processes operate; 2) that weathering processes are controlled by climate rather than by other controls on chemical reactions; and 3) that vegetation (a source of weathering agents as well as a factor in erosion, for example, Drever 1994) responds to climate in a strictly linear fashion. In place of these conceptual climatic models and their assumptions, we have proposed a multivariable model of the geographical variability of chemical weathering in which the rate of weathering is a function of a number of different reactive factors operating at multiple scales (e.g. temperature at reaction sites, scale, mineralogy of parent material and weathered products, soil hydraulics, microtopography, the nature of biotic and abiotic solutions). In this "open" framework, the relevant climatic variables are

the ones that come into play at the weathering boundary layer, i.e., the reaction sites.

Our assessment of these two competing theories is based on evaluations of the systematics of weathering processes and of empirical data at different scales (ranging from the hygroscopic mineral boundary layer to entire landscapes), in cold and warm drylands—the regions which have provided the visual observations on which the climatic thesis of weathering is predicated. Our evaluation of these competing theories of geographical variability in weathering argues in behalf of abandoning the mesoscale, bivariate, and static climatic theory (based on mesoscale tropospheric means) because the latter fails to account for the quantity and diversity of different kinds of data, is incompatible with established theoretical frameworks, and does not encompass the scope of weathering research in physical geography's cognate disciplines. We advocate instead a microscopic perspective on the boundary layer where weathering actually occurs. We believe that the boundary-layer model has considerable potential as an analytical and predictive tool and that it invites testing and refinement by geographers as well as weathering researchers in related fields. For it is, after all, in the microscopic details of the boundary layer that the processes of weathering are most clearly revealed.

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The prevailing theory used to explain geographical variability in weathering is based on visual correlations with climatic regions. For instance, mechanical weathering is assumed to predominate in warm and cold deserts. Yet this visual perspective fails to account for a diversity and quantity of data at the mineral-atmosphere-hydrosphere-biosphere interface where weathering processes actually occur. To address these discrepancies, a new model is proposed which views geographical variability in weathering as a function of synergistic biological, chemical, and physi-

cal processes that are controlled by factors that vary at the microscopic weathering boundary-layer. The new multivariate model better explains weathering observations at hygroscopic, capillary, pedogenic, landform, and landscape scales. **Key Words:** climate, desert, geomorphology, microscopy, periglacial, soils, weathering.

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