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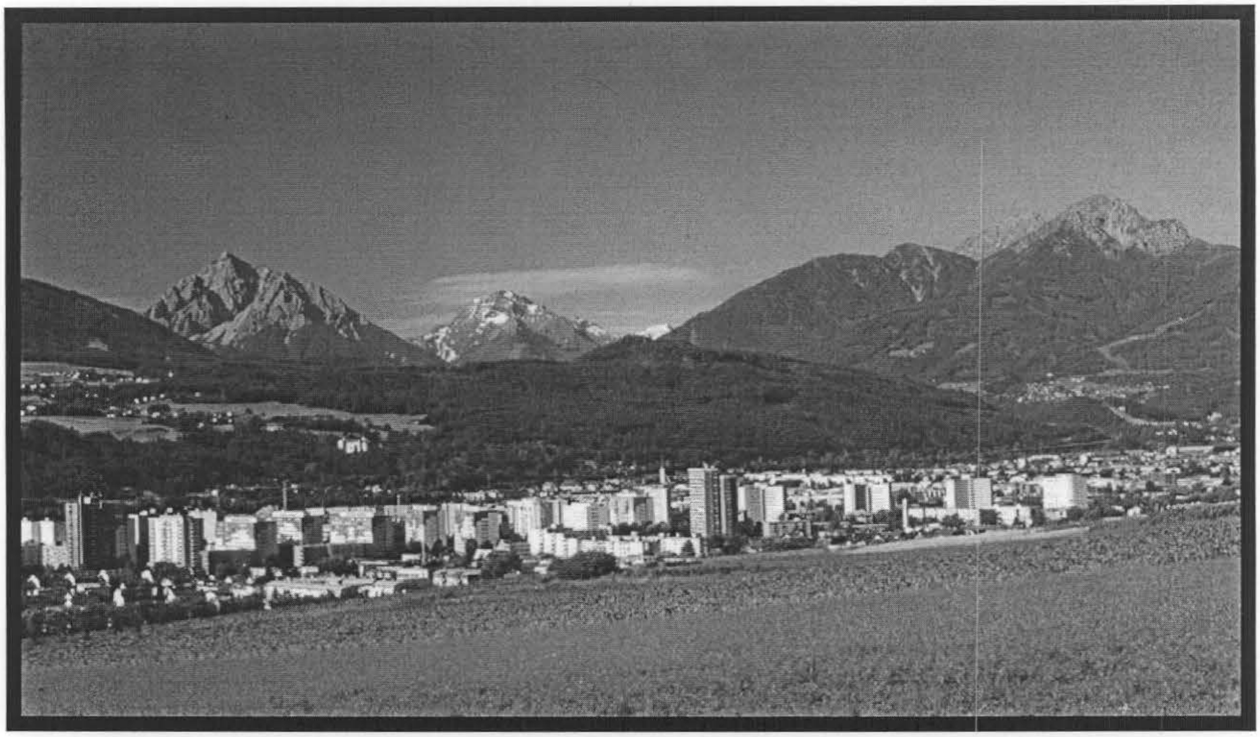

A view of Innsbruck and the metamorphic Alps from Arzl in Summer 2004. The Olympic Village area of eastern Innsbruck is in the foreground. Above and beyond Innsbruck at mid-left is Schloss Ambras, the Habsburg castle. In the background, from left to right, are the rounded flank, but not the crest, of the Patscherkofel, jagged Serles, glacier-bearing Habicht, the peak of the otherwise-hidden Marchreisenspitze, and the Nockspitze with its barren peak but tree-covered flanks.
A color vesion of this image can be seen at www.gly.uga.edu/railsback/AG/TDI286.html.

Alpine and Glacial Geology

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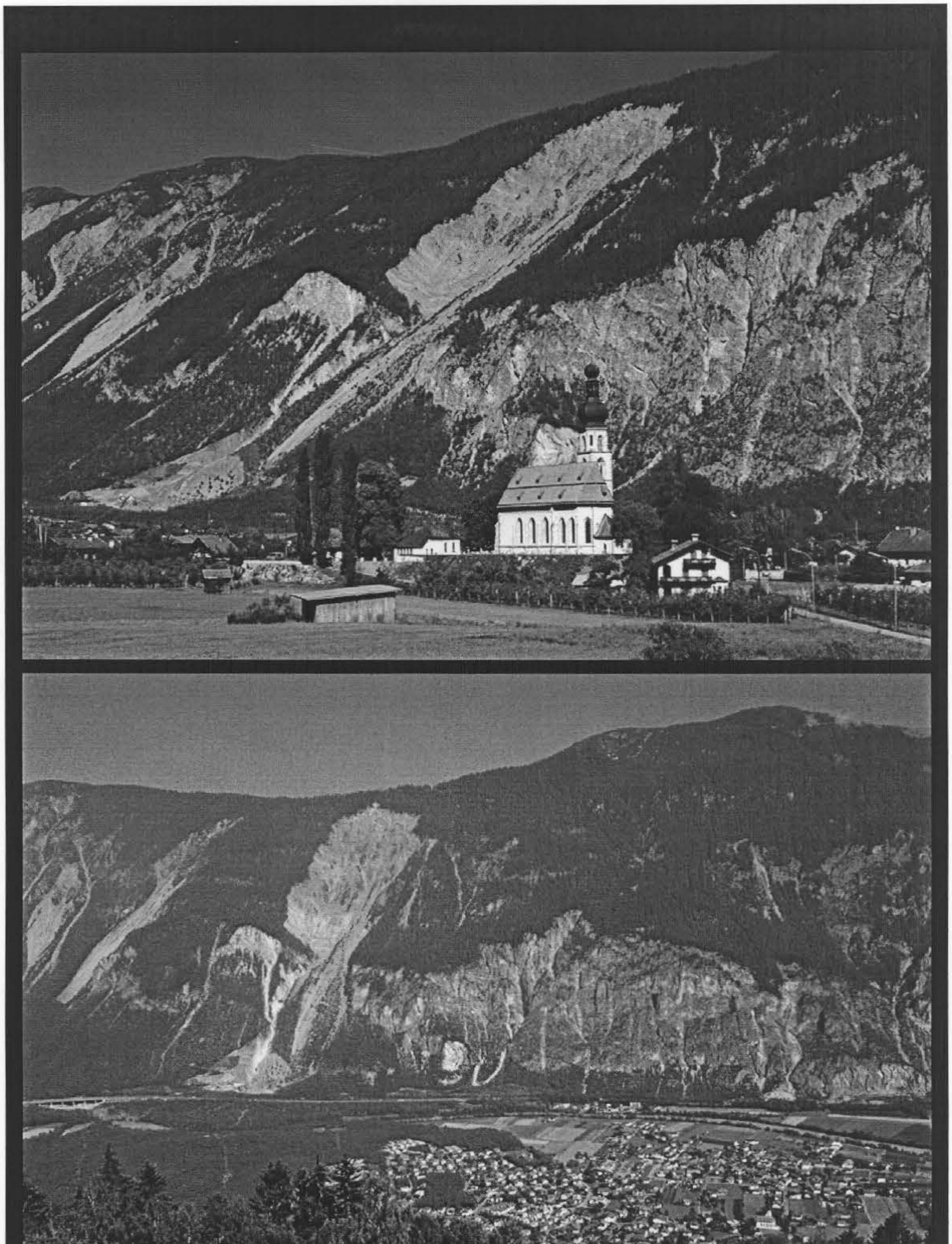
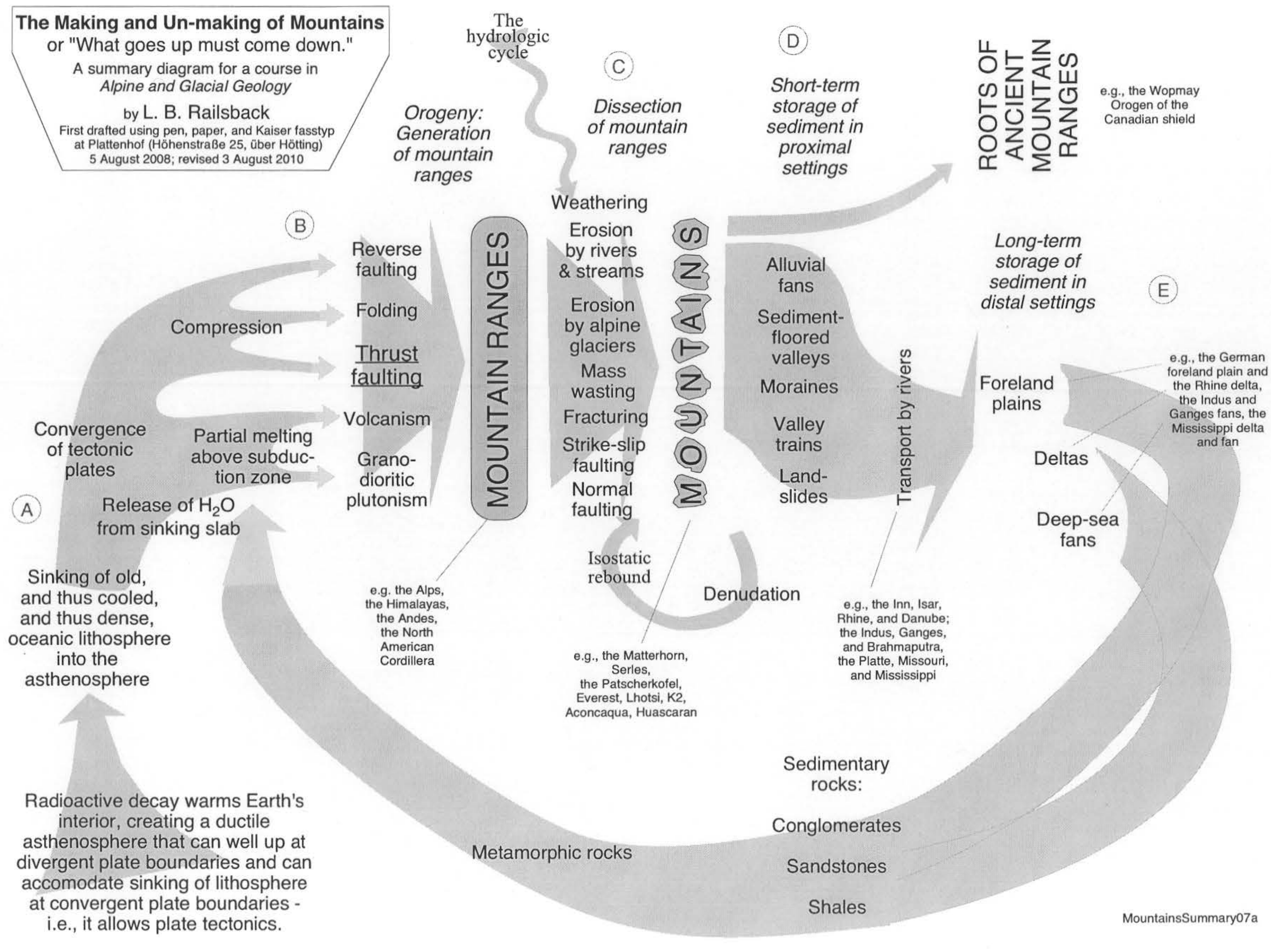


Two views of the village of Haiming in the Inn Valley west of Innsbruck, and beyond it the scarp and talus cone of a landslide on the south side of the Calcareous Alps.



A diagram for the third edition of Railsback's *Alpine and Glacial Geology*

PREFACE I: THIS BOOK

This book was written to accompany a course called "Alpine and Glacial Geology" taught in the University of New Orleans's International Summer School in Innsbruck, Austria. The course thus covers alpine geology (the geology of mountains and mountain ranges) and glacial geology (the geology of glaciers and the history of the ice ages). The two topics are combined in the course because Innsbruck provides such a great opportunity to examine mountains and glaciers. However, as the concluding chapter of this book will argue, there are sound scientific reasons to link alpine geology and glacial geology at both regional and global scales.

The UNO-Innsbruck course is open to students with no background in Geology, as well as to those with more experience in Geology. No textbook focuses on the combination of alpine geology and glacial geology, and certainly no other book would attempt treatments of two such specialized topics while also presenting introductory concepts for beginning students. This book therefore attempts to meet the unique needs of the course taught in Innsbruck by covering the most essential geological concepts needed (Part I) and then addressing the two major themes, alpine geology and glaciology (Parts II and III). Hopefully the book will also meet the needs of others interested in mountains and in the glaciers that have sculpted them. As a photocopy-and-bind document, it certainly is a less expensive option for students otherwise faced with buying both a tectonics book and a text on glacial processes.

Part I was written with a minimalist mindset and is intended to provide just those concepts needed to understand the later parts of the book and related materials. It thus contains only part of the material contained in most introductory texts. On the other hand, it also contains some material typically found only in more advanced texts, such as the IUGS classification of igneous rocks, because students at this level inevitably encounter some more advanced terms and concepts.

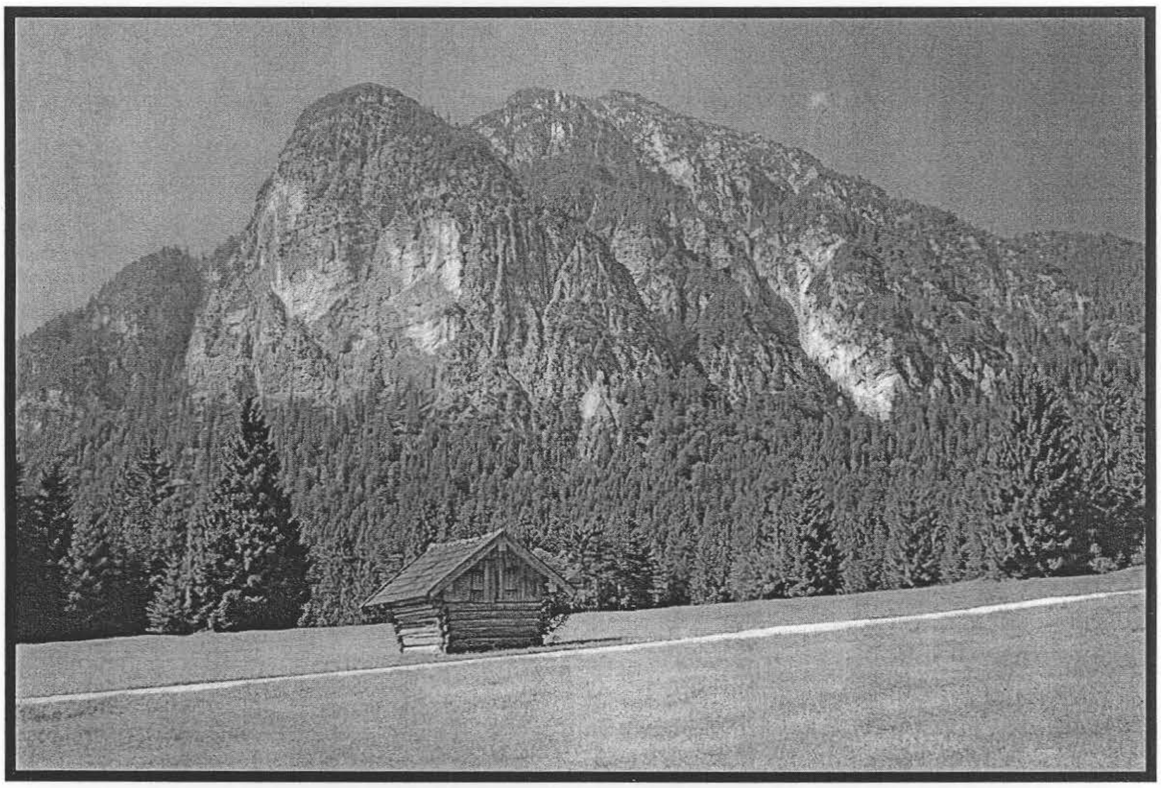
Readers of this book may be amused or aghast to find that it has many footnotes, some of which are lengthy. The logic of this arrangement is to include material that students should definitely know in the text, and to put less critical but interesting or explanatory material in footnotes that students may not be required to read. This allows instructors to assign text alone, or text and specific footnotes, or text and all footnotes.

The three foremost differences between the second edition and the first are that (1) the second edition has three new figures (Figs. 8-6, 9-7, and 9-8), (2) the second edition has new text dealing with the age of the Earth in Chapter 1, with dolostones in Chapter 4, with plate tectonics in Chapter 9, and with mountains and mountain ranges in Chapter 16, and (3) many typographical and editorial errors present in the first edition have been corrected in the second. Each of the appendices was also expanded with new material. The number of pages increased from 185 to 194, and the number of footnotes increased from 109 to 115.

This book would not exist without the support of many people. The administrators of the UNO-Innsbruck program, including Dr. Margaret Davidson, Dr. Carl Malmgren, Dr. Milton Pressley, Dr. John Altazan, Dr. Alea Cot, and Mr. Peter Alongia, kindly invited the author to teach in the program, despite the author's sometimes dour demeanor. Mr. Peter Alongia and Mr. Andre Martinez facilitated inspiring field trips to the Gaisbergferner Glacier above Obergurgl. Herr Walter Freytag of Innsbruck generously provided insights about the landscape of Tirol. Herr Prof. Dr. Karl Krainer of the University of Innsbruck kindly guided the author into the field to provide the latter's first glimpses of the geology of the region around Innsbruck. Cornelia Markut and Christoph Bjerler of Innsbruck kindly allowed the author to use their computer for writing in their apartment during his summer stays in Fallbachgasse by the St. Nikolaus Kirche near the Inn. Drs. James E. Wright and Sandra J. Wyld of the University of Georgia shared their insights about tectonics. Drs. Anna Williams and Amy Reeder proofread parts of an early draft the first edition, and Margot Augustin did a spectacular job of carefully editing and proofing the first edition, to the betterment of the second.

Perhaps most importantly, the enthusiasm of many Alpine and Glacial students over the years also motivated the author to write this book for their successors. Those students include, but are not limited to, Hans Knopfel, Laura Johnson, Eric Jovanovich, Ashley Damhorst, Robert Schwing, Lyndsey Berryman, Kacey Marshall, Sara Stanford, Bill Blount, Amanda Ensminger, Currie Moore, Lillian Luffey, Callie Manget, Sara McIntyre, Nicole Pelayo, Margot Augustin, and Jimmy Wylie. Nothing promotes the enthusiasm of professors like enthusiastic students.

Most of this book has not undergone peer review. Professional geologists who are appalled by this situation are encouraged to read the chapters in their area of expertise and send their constructive comments to the author, who will very much appreciate such help. Students are likewise encouraged to record errors or problems on the "Notes" page at the back of the book, and to tear out that page at the end of the course and give it, or send it, to the author. Any such help will be appreciated.



A scene in the Calcareous Alps near Garmisch-Partenkirchen

PREFACE II: TIROL

This book was written for use in a class taught in Innsbruck, in the Austrian state of Tirol, and the book in part focuses on the geology of Tirol. It therefore seems fitting to devote some attention to Tirol from a historical and cultural perspective, before turning to the geology of Tirol.

In a discussion of Tirol, the first issue is the name itself. Speakers of English commonly use the old spelling "Tyrol", and they commonly use the expression "the Tyrol". However, local usage in modern German almost always employs the spelling "Tirol". Some names of nations in German use a definite article, such as "die Schweiz" for Switzerland, but English speakers don't feel obligated to say "the Switzerland" when speaking of the nation to the west of Austria. The English usage most consonant with modern local usage in the German language thus seems to be "Tirol", without a "the" before it.

What is Tirol?

"Tirol" has multiple meanings. Tirol in the *historical* sense is a region extending from the northern Alps, and thus from the Karwendel mountains north of Innsbruck in today's western Austria, southward beyond Bozen (Bolzano) in northern Italy. In 1804 its borders were extended even farther south to include Trento and northern-most Lake Garda. One powerful reminder that historic Tirol extended into, and was largely in, modern Italy is the origin of the name "Tirol" itself. The name comes from Schloss Tirol, the castle above Meran (Merano) that was the seat of Tirol's rulers, the Counts of Tirol, in the Middle Ages. Dorf Tirol is the village by that now-ruined castle. Today Innsbruck is the capital of Tirol, but Tirol's first capital would have been Meran.

In the *cultural* sense, Tirol is an Alpine region inhabited by speakers of German (and Ladin, a Romance language spoken in South Tirol). Thus it is much the same region as the historical Tirol, but with its southern border defined by the transition from speakers of German to speakers of Italian. In today's northern Italy, one still commonly hears Italian spoken in the valley towns and cities, and German spoken in the uplands. The southern border of the historic Tirol was also Europe's "bread line", an important cultural division between peoples to the north who ate dark rye breads and peoples to the south who ate white wheat breads.[1]

In the modern *political* sense, Tirol is one of the nine federal states (Bundesländer) of the modern Austria. This modern political Tirol is only the northern part of the historical and cultural Tirols, and it consists of two non-contiguous regions most readily linked by travel through today's northern Italy. Innsbruck is the capital of this modern Austrian state, and Lienz (not Linz!) is the administrative center or "sub-capital" of Ost-Tirol (east Tirol).

Tirol to 1918

Tirol's most defining geographic characteristic is its spectacular mountains, and its rugged landscape means that many of its people have lived in valleys isolated from each other, and with their own dialects of Tirolean German. However, the linkages binding the historical Tirol together have been its river valleys and passes. They most notably include the Inn River, running west-to-east in a major valley in northern Tirol. If the Inn makes the top of a "T", the stem of the "T" then consists of the valley of the Sill River southward up to the Brenner Pass, and then the valley of the Eisack southward to Bozen (Bolzano). Innsbruck, Tirol's modern capital, is at the meeting of the two parts of this Tirolean "T".

Tirol's Germanic heritage dates from at least the time of the Roman Empire and its fall, when Germanic peoples moved southward. In the Middle Ages, Tirol was part of the Holy Roman Empire (das heilige Römische Reich Deutscher Nation), the loosely-defined realm of Germanic lands whose emperor was often referred to as "the German king". In 1363, Margaret of Carinthia, Tirol's last ruler descended from the lords of Schloss Tirol, was left heir-less by the death of her only son, and so she ceded Tirol to the Habsburg Duke of Austria, Rudolph IV[2]. Tirol was thereafter under the rule of the Habsburgs, and thus part of Austria, until 1918. The Habsburgs were commonly Holy Roman Emperors until Napoleon's dissolution of that institution in 1806,

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[1] For more, see Chapter 3 of Ward (1993).

[2] Margaret was also known as Margaret Maultasch. Her other lasting legacy to Tirol was to introduce a symbol still commonly seen today, the Tirolean eagle. It's a red eagle facing left, with a green laurel crown. It was actually the heraldic eagle of the family of her second husband Ludwig, the margrave of Brandenburg.

and thereafter they were emperors of Austria-Hungary until their abdication in 1918 and the dissolution of the Austro-Hungarian empire after World War I. Thus for most of Tirol's history, it has been part of German-speaking political realms.

One noteworthy if brief interlude in Habsburg control of Tirol took place from 1805 to 1814, when the French ruler Napoleon Bonaparte placed Tirol under Bavarian rule. The Tirolean peasantry under the leadership of Andreas Hofer, an innkeeper, revolted against the Bavarians in 1809, and Hofer's small army defeated the Bavarians multiple times. In 1810, Hofer was captured and shot on Napoleon's orders, but he remains the Tirolean national hero, with a monument in Innsbruck and an annual celebration in Meran. In 1814, during Napoleon's removal from power, Tirol from the Karwendel to Trento was reunited with Habsburg Austria.

South Tirol

The historic Tirol was split asunder in 1919 in the redefinition of European geography following World War I, and in the dismemberment of the German and Austro-Hungarian empires. Early in World War I, Italy had remained neutral, but the secret 1915 Treaty of London promised Italy Tirol up to the Brenner Pass, if Italy would enter the war on the Allied side. Italy did so, and southern Tirol was the site of harsh fighting between Italian and Austro-Hungarian troops. In 1919, after the war, international diplomacy gave Italy its promised reward for supporting the victorious Allies.

With this partitioning of Tirol, the southern part became the "South Tirol" of today, although to Italians it is known as "Alto Adige", the high (alto) watershed of the Adige (or Etsch), the river that flows through Meran, Bozen, Trient, and Verona. Teaching of the German language to children was forbidden, except for private religious schooling, and expressions of Tirolean nationalism, such as displays of red and white in clothing, were commonly suppressed by the Italian government. Migration of Italians northward into the towns of South Tirol was encouraged. Unhappiness among Germanic people in South Tirol was sufficiently great that, when fascist Italy and Nazi Germany held a plebiscite in 1939, a majority of the Germanic population voted to leave their ancestral farms and move northward across the Brenner Pass. Negotiations between Italy and Austria over conditions in South Tirol went on until the 1960s but stalled in response to terrorism by Tirolean separatists, who carried out more than 300 attacks between 1956 and 1988. In Innsbruck, squares and streets named Bozner Platz, Sud Tiroler Platz, Meraner Straße, Salurner Straße, Brixner Straße, Brunecker Staße, and Sterzinger Straße are public indications of the memory of South Tirol[3], and in the 1990s graffiti on the Inn floodwalls still demanded "Free South Tirol". A 2006 survey found that 54% of German speakers in South Tirol still favored reunification with Austria.

(North) Tirol

The partitioning of Tirol in 1919 left North Tirol, previously part of the large and powerful Austro-Hungarian empire, as part of the much reduced and less powerful new nation of Austria. Unimpressed with the new "rump" Austria and instead seeing the power and cultural similarity of Germany, Tiroleans voted overwhelmingly (>98%) in 1921 that their land should be united with post-war democratic Germany (much as Austria's westernmost state, Vorarlberg, had voted by more than 80% in 1919 to join Switzerland). The border gates between Tirol and Germany at Kufstein and Scharnitz were almost immediately torn down. France and her allies, however, were intent on breaking up Germany rather than seeing it grow, and so northern Tirol remained part of the new Austria[4].

Tirol, along with the rest of Austria, was incorporated into Hitler's Third Reich with the Anschluss, or annexation of Austria into Germany, in 1938.[5] Tirol was thus a part of the Third Reich

[3] The places commemorated are Bolzano, South Tirol, Merano, Salorno, Bressanone, Brunico, and Vipiteno. Note that Innsbruck has no Landecker, Imster, Telfser, Seefelder, Jenbacher, Wörgler, Kufsteiner, or Lienzer Straße – all the streets and squares named after Tirolean towns are named after South Tirolean towns.

[4] In fact, the entire new Austrian nation had sought inclusion in the German republic, declaring in its constitution of November 1918 that "German-Austria [the new post-war Austria] is a component part of the German Republic". However, France made clear that Germany's reparations for World War I would be much greater if Anschluss with Austria took place, and Germany did not pursue the matter further.

[5] The idea of Anschluss ("connection" or "joining") of Germany and Austria began in the 1870s with the formation of the German Empire and was revived in

in World War II, as its cemeteries and soldiers' monuments attest. Because the railroad yards in Innsbruck are the take-off point of the tracks up the Brenner Pass and Italy, Allied bombing focused on Innsbruck, and bombs fell not only on the railroad yards but across much of the city. Brixlegg, a town down the Inn, was the site of copper mines and thus a had a factory making shell casings, and it was bombed just a few days before the war ended.

With the end of World War II in 1945, Tirol and the rest of Austria were extracted from Germany and re-established as a nation, necessarily and deliberately neutral between NATO and the Soviet-dominated Eastern Bloc. In 1995, after great debate, Austria voted to enter the European Union, despite grave concerns about unification's implications for Austria's economy and especially for its farmers. In 2002 Austria adopted the euro and gave up its national currency, the schilling, which had been a source of national pride because the schilling had not been allowed to inflate nearly so much as some other European national currencies. In Tirol, signs on barns still (as of 2007) condemned entry into the European Union, and in 2006 much dismay arose as new health regulations imposed by the EU forced the closing of many alpine hüttes, the tiny inns feeding and sheltering hikers high in the mountains.

The Tirol(s) of today

In the early 21st century, Tirol tries to balance its past and its future, its extent as the larger cultural region and the smaller Austrian state, its Tirolean sense of independence and its EU membership. Tirol's defining characteristics, the Alps and Tirolean culture, provide a magnet drawing tourist dollars and euros that fuel the Tirolean economy. Tirol's T intersection of the Inn Valley and the route through the Brenner Pass assure that trains, tourists, and trade will pass through Tirol. However, the Autobahn superhighways and the many ski lifts mar the alpine landscape, and the influx of tourists inevitably intrudes upon and changes the native Tirolean culture. It's hard to teach your children the Tirolean ways and the dialect of the valley in which your family has lived for centuries, when you spend all day speaking English to the carloads and busloads of tourists who come up the new highway to your once-remote village. For visitors who would never think of leaving trash on the splendid Alpine landscape, it may also be important to not bring foreign cultural expectations when coming to see Tirolean culture. Saying "Gruss Gott" in greeting rather than "Hello" or "Guten Tag"[6], obeying a "don't walk" signal, keeping quiet in a church, and other ways of blending in, rather than standing out, are small but significant ways of helping preserve the cultural landscape, as well as the physical landscape, that we've come to enjoy in Tirol.

1918 when German Austria, the westerm remant of the Austro-Hungarian Empire, sought to join Germany but was barred by the Treaty of Versailles. That desire was very different from the annexation of Austria by Nazi Germany in 1938. Innsbruck's memorial in Eduard Wallnöfer Platz on Salurner Straße is a reminder of the resulting persecution of Jews that took place in the Tirolean capital.

[6] Some bits of Tirolerisch (for each, the first is the most common and standard):

Saying "Hello"
Grüss Gott ("let God greet you" - probably used more among older or more traditional people, but common throughout)
Grüss Di ("greetings to you" - perhaps used by younger or more secular people)
Servus (used more by men or to men, especially younger ones)

Saying "Thanks"
Danke Schön ("Thanks nicely" - the universal Bavarian-Tirolean thank you)
Danke Dir ("Thanks (to) you" with the informal you - used more by and/or to women)
Gelts Gott ("May God repay you" - a peculiarly Tirolean expression rarely heard on the street)

Saying "Good-bye"
Wiedersehn (just a shorter form of the standard German expression "Auf Wiedersehen")
Wiederschauen (the short version of the southern German "Auf Wiederschauen" for good-bye)
Tschüss (a casual good-bye more common among younger people)

Add two critical words universally used in German:
Bitte ("please")
Entschuldigung ("Excuse me")

Sources:

Encyclopedia Britannica (1968) entries on Tirol and Margaret of Carinthia.

Steininger, R., Bischof, G., and Gehler, M., 2002, *Austria in the Twentieth Century:* Transaction Press, 269 p.

Stodddard, F. Walcott. 1914, *Tramps Through Tirol:* New York, James Pott & Co., 299 p.

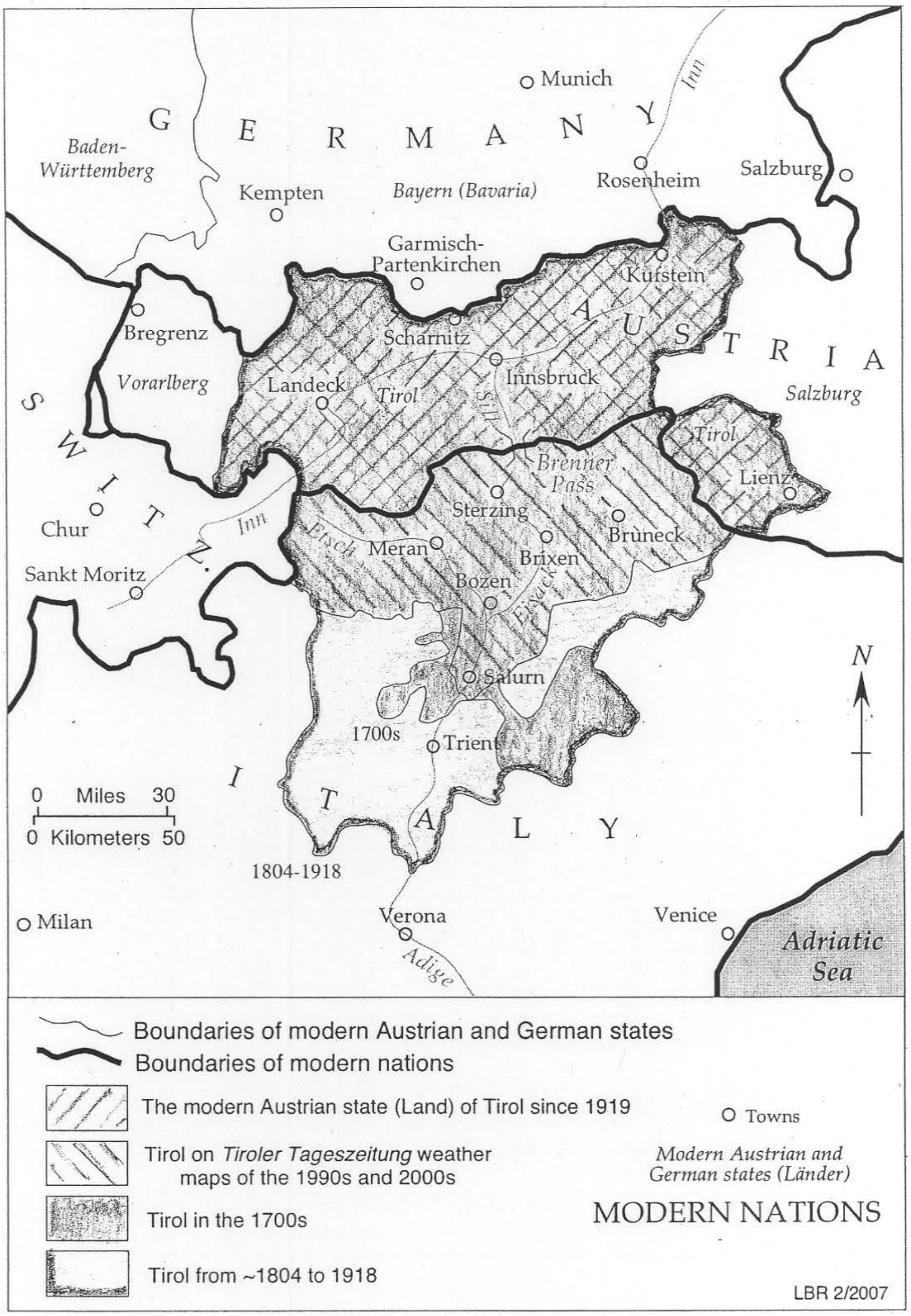
Ward, M.C., 1993, *The Hidden Life of Tirol:* Prospect Heights, Illinois, Waveland Press, 222 p.

Wikipedia entries on Andreas Hofer, South Tyrol, Trentino-South Tyrol, and Sterzing.

Further browsing:

The Tirol Atlas produced by the Department of Geography of the University of Innsbruck (http://tirolatlas.uibk.ac.at/content.html.en)

The Inn River gorge between Imst and Roppen, west of Innbsrück



A map of Tirol.

Part I. Fundamentals of Geology

Chapter 1: Geologic Concepts

Geologists, and persons with some education in geology, take a somewhat different view of the world than do people who have never considered geology before. This chapter introduces a few ideas that may help newcomers get into the flow of geologic thinking.

The size of the Earth

From a human perspective, the Earth is large, and probably larger than our casual thoughts would suggest. Our deepest mines impress us with the temperatures at which miners work, and the long trips in and out that miners must make each day impress us with regard to depth. However, these mines only penetrate a small part of Earth's crust (Fig. 1-1). Our deepest wells and boreholes go deeper, to about 10 kilometers or 6 miles, but they similarly don't get far into the continental crust. With that said, the crust is only a very thin film compared to the entire volume of the Earth, and to the thickness of the underlying mantle.

Another way to consider the size of the Earth is to look at its most prominent features. Earth's highest mountains and deepest seafloor are barely perceptible as blemishes on the smooth sphere of the entire planet (Fig. 1-1). Also consider the international space station, the laboratory we have boldly constructed in outer space. We envision the space station as something far above the Earth. In fact, compared to the size of the Earth, our far-away space station is barely above our planet's surface (Fig. 1-1).

The age of the Earth

According to modern widespread consensus among scientists, the Earth is 4.6 billion years old. We know that age because we have measured the age of Earth's oldest rocks and minerals to be more than 4.0 billion years old, and we recognize that those rocks and minerals formed *after* the origin of the Earth, so that they only provide a minimum age. Furthermore, we know the age of the Earth because we can use the composition of Earth's lead, the product of radioactive decay, to determine Earth's age. Those calculations yield an age of 4.6 billion years. In addition, we have determined the age of the oldest rocks on Earth's moon, a body that we think formed at the same time as the Earth. We find an age of 4.6 billion years there too. We have also determined the age of meteorites that we think formed when the solar system condensed to form the planets, including Earth. Again we find an age of 4.6 billion years. In addition, we can deduce from the time required for many processes on Earth (cooling and exposure of rocks formed from molten mineral material, thicknesses and rates of deposition of sediments, accumulation of salts in the sea, etc.) that the Earth must be very old, and not merely thousands or even a few million years old.

The age of the Earth is important in geologic thinking for at least three reasons. First, it explains how slow processes can nonetheless cause major changes. Erosion of mountains, deposition of deep-sea sediments, and movements of continents seem like incredibly slow processes at human time scales. However, if a process acts for millions of years, the results are impressive. If something accumulates at the rate of an inch per year, 1026 miles of it will have accumulated since the last dinosaur died 65 million years ago.

Secondly, the duration of geologic time makes rare events common. If we estimate, for example, that a meteorite of a certain size hits the Earth once every 100,000 years, we can dismiss meteorite impacts of that size as events that are unlikely to happen in our lifetimes. However, such impacts will have happened 46,000 times in the time allowed by the age of the Earth. To summarize, extension of processes to geologic time scales makes the slow very effective and makes the rare common.

A third implication of the immensity of geologic time is that we see the Earth at just a moment in its history. This means that geologists look at the Earth, and especially its landscapes, and envision what those mountains were previously and what they may become. For example, a casual visitor to Stone Mountain, a large hill of granite east of Atlanta, sees a sizeable bump on the landscape. Someone acquainted with geology, however, will realize that the granite formed deep beneath Earth's surface and is exposed only because kilometers of overlying rock have been removed by erosion. Thus the geologically aware will think a bit about the landscape of the past. Someone acquainted with geology will also realize that someday erosion will have removed Stone Mountain and perhaps leave only its underlying root. Thinking like this makes one realize that other analogs of Stone Mountain may have existed in the past but have been lost to erosion by now, and other bodies of granite deeper in the Piedmont may be waiting to be exhumed and then to stand as high above the landscape as Stone Mountain does today. Thus an appreciation of geologic time makes on realize that present landscapes are just one snapshot in a very long series, and likewise that the present configuration of continents and ocean basins is just the one step that we happen to see in the progression from the various geographies of the past to those of the future.

Describing this expanse of geologic time requires a vocabulary that is collectively the geologic time scale. Just as one might divide 19th-century American history into an Early Republic, Ante-Bellum, Civil War, and Reconstruction period, geologists use terms for periods with distinctive characteristics, usually characteristics of the fossils of those times. Just as a scholar of American History might subdivide one of the broad periods above - the Civil War might be divided Bull Run to Antietam, Antietam to Gettysburg, etc. - geologists break down their broadest time intervals into shorter subdivisions. The result is the hierarchical scheme shown in Figure 1-2.

Rock, bedrock, and the water within

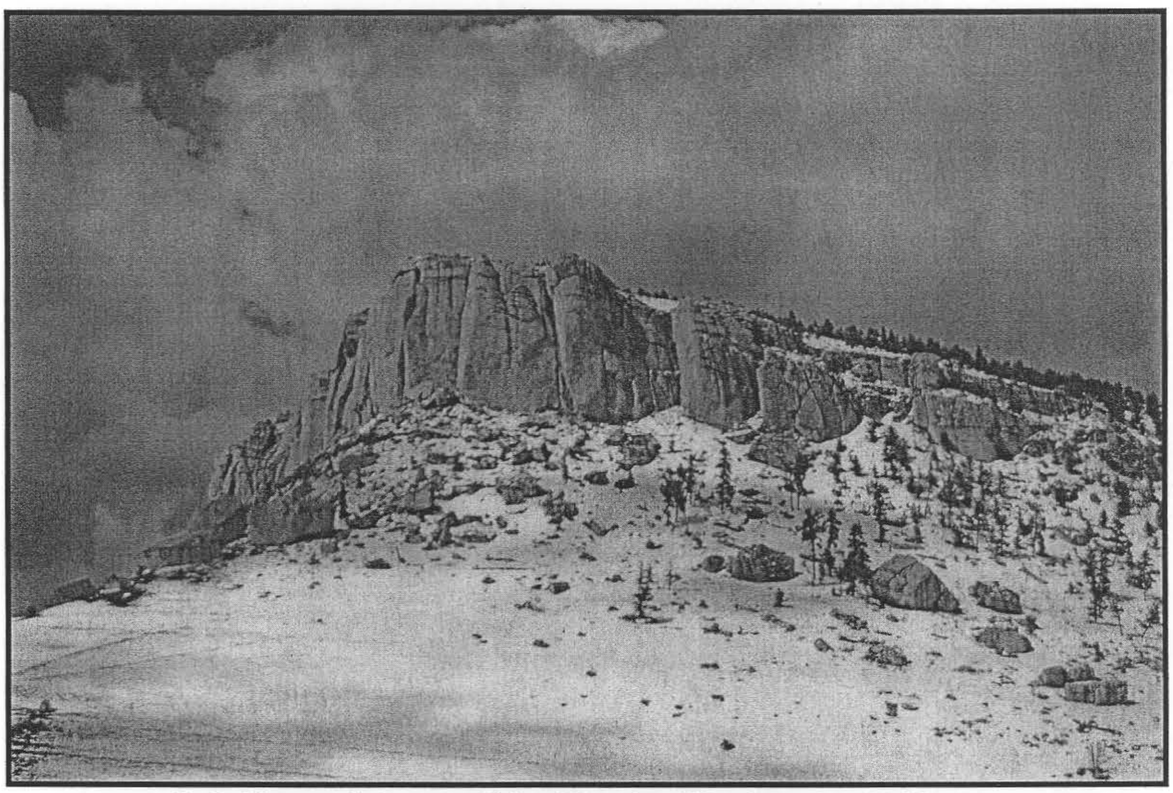
The Earth surface on which we walk does not typically consist of solid rock. We commonly walk, garden, and farm on soil, a loose material resulting from action of organisms and weather on the land surface. We may also walk on sediments, especially if we walk or wade on sands at the shore or in a river, or if we track through muds of floodplains and marshes. We may also come across larger chunks of mineral material that we call rocks, but they are commonly just large chunks that sit on or amidst the soil or sediment.

Beneath the soil or sediment, and sometimes at the land surface, there is bedrock (Fig. 1-1). Bedrock is solid rock that has not moved, at least relative to its tectonic plate, in thousands if not millions of years, and below which there is only more rock. This bedrock may have cracks in it, but they decrease in abundance and width downward. Bedrock is commonly visible at the land surface in mountains, in canyons, and in soil-free patches called barrens or balds. It is commonly a few feet to a few tens of feet below the surface of soils. In river valleys it may be tens to a few hundred feet below the surfaces of floodplains, and in glaciated landscapes it may be similar distances below the surfaces of glacial tills. On coastal plains, it may be hundreds to thousands of feet below the surface of the loose sediments making those landscapes. However, whether it begins a few inches or a few thousand feet below the land surface, bedrock is the solid thing that makes up the rest of the 1800 miles (2900 kilometers) of crust and mantle beneath our feet.

The other thing beneath our feet is water (Fig. 1-1). Dig a well into any soil or sediment, or drill a well into any bedrock, and you will encounter water, commonly within a few tens of feet and certainly within a thousand feet or so. This water fills the pore space, or inter-particle space, in soils and sediments, and it fills the cracks or fractures in bedrock. The water table is surface below which water, rather than air, fills those pores and fractures. The significance of this water is at least twofold. First, it means that we should not think of air as filling the deep spaces within the Earth - in fact, groundwater commonly precludes the resupply of atmospheric oxygen into the subsurface. Secondly, it means that movement of solutes through or with water will explain many subsurface processes, such as chemical precipitation of minerals like ore deposits. It also means that we all walk on water.

Perspectives on the Earth

Geologists try to think in terms of three-dimensional images of Earth's interior. Sometimes they draw those, as in block diagrams (Fig. 1-3). However, they more commonly draw in terms of two-dimensional vertical or horizontal slices or surfaces. Images of vertical slices are cross-sections. Images of horizontal slices or semi-planar surfaces are maps. Whenever you're completely at a loss about a diagram, the first question to ask may be "Is this a cross-section or a map?", and the second question is either "From where to where does this cross-section extend?" or "Of what surface, or at what depth, is this map drawn?". Another good question will be "What is the scale?". A good diagram answers these questions on its own, but a student should never be afraid to ask these questions if the diagram doesn't provide the answers.



Newly fallen snow in the eastern Bighorn Mountains of Wyoming on June 7, 2007

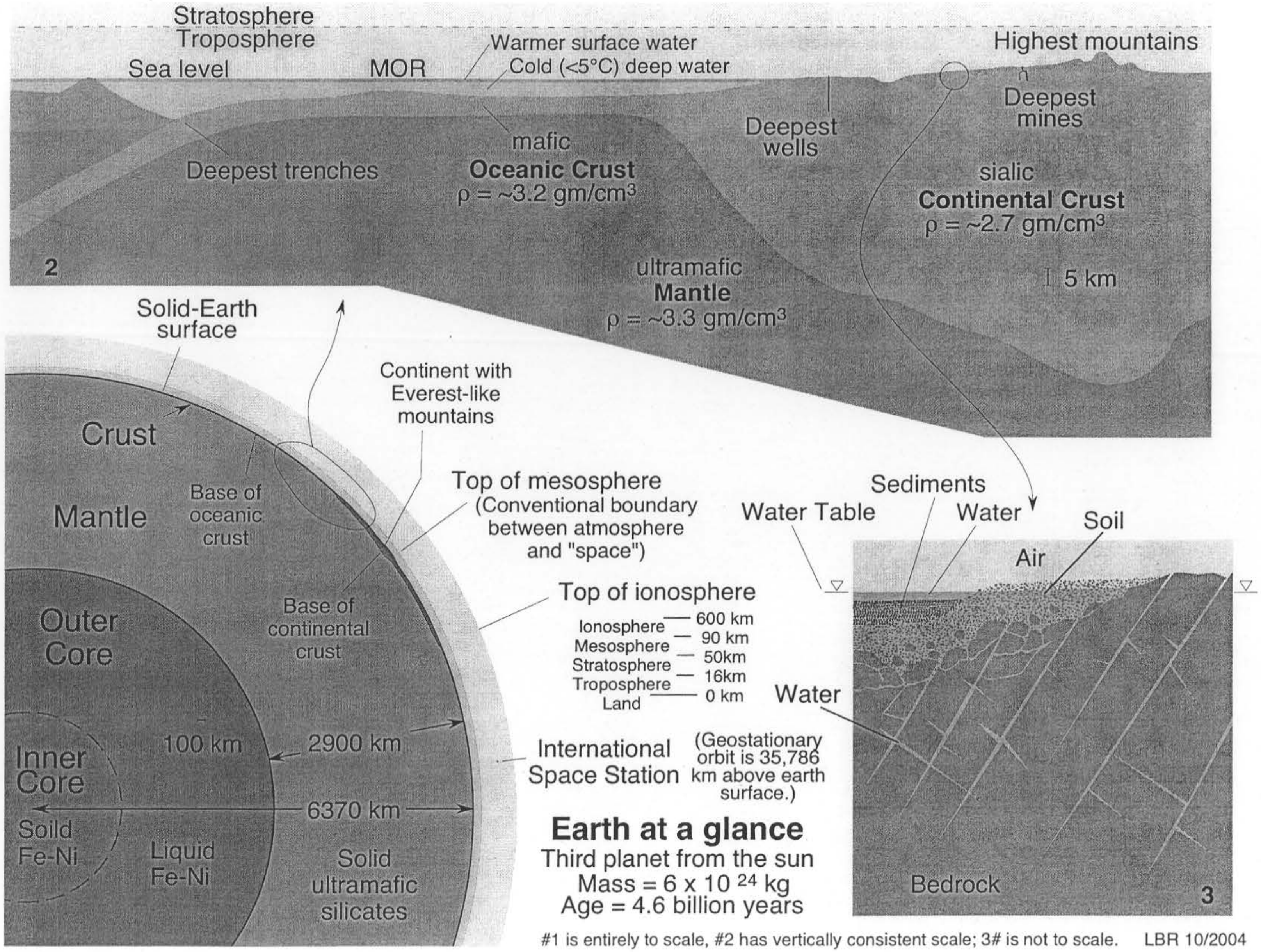


Figure 1-1: A cross-section illustrating basic features of the Earth

In 2003, the International Commission on Stratigraphy divided the Cenzoic into Paleogene and Neogene, eliminating the terms "Tertiary" and perhaps "Quaternary". "Paleogene" and "Neogene" are paleontologically meaningful terms, but the major paleoenvironmental breaks are at the Eocene-Oligocene boundary (a major global cooling event) and Pliocene-Pleistocene boundary (the onset of major continental glaciation in the Niorthern Hermisphere - hence the expression "Quaternary glaciation"). Earth Scientists interested in environmental change are thus prone to continue using "Tertiary" and "Quaternary" rather than "Paleogene" and "Neogene".

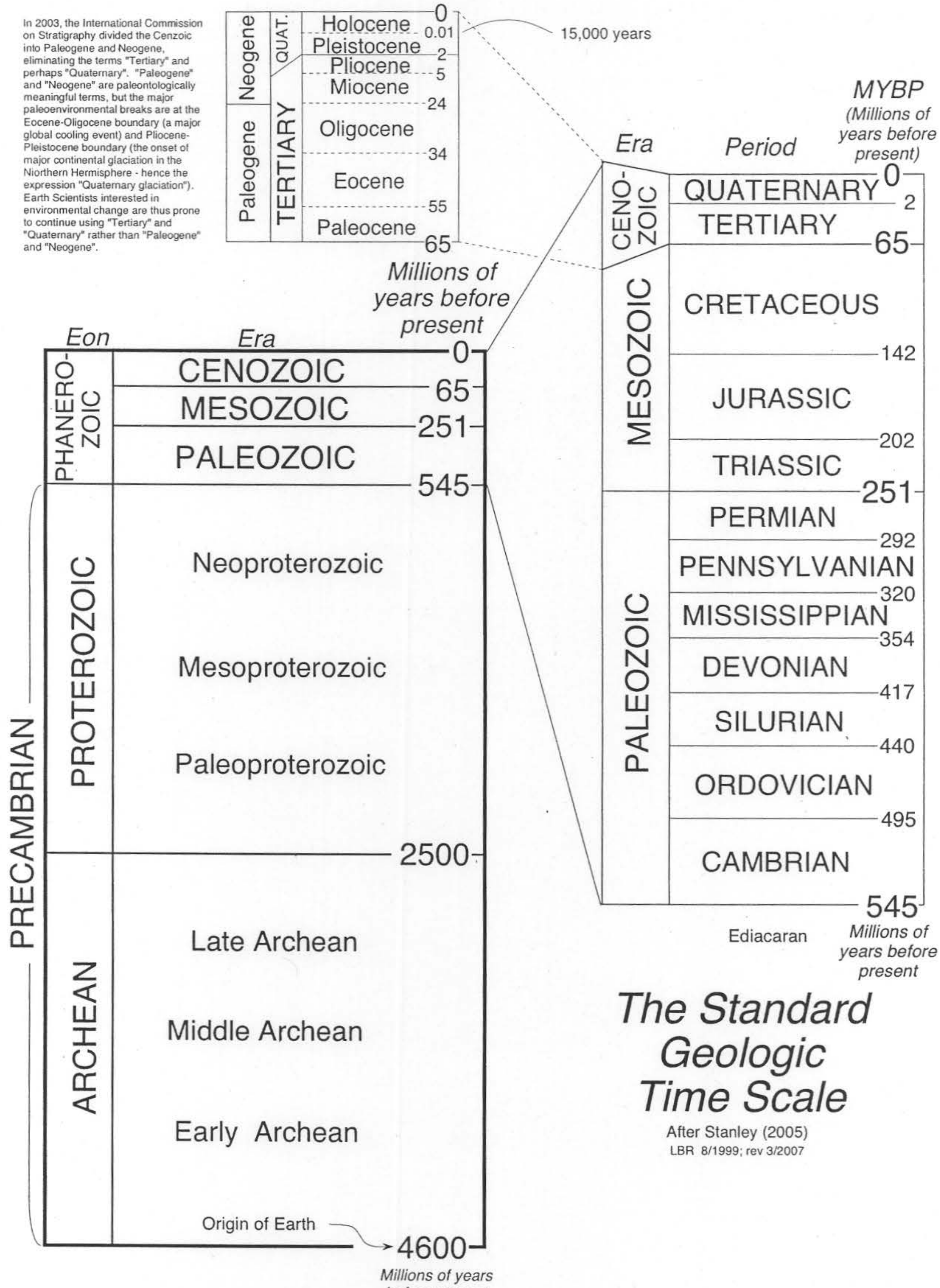


Figure 1-2: The geologic time scale.

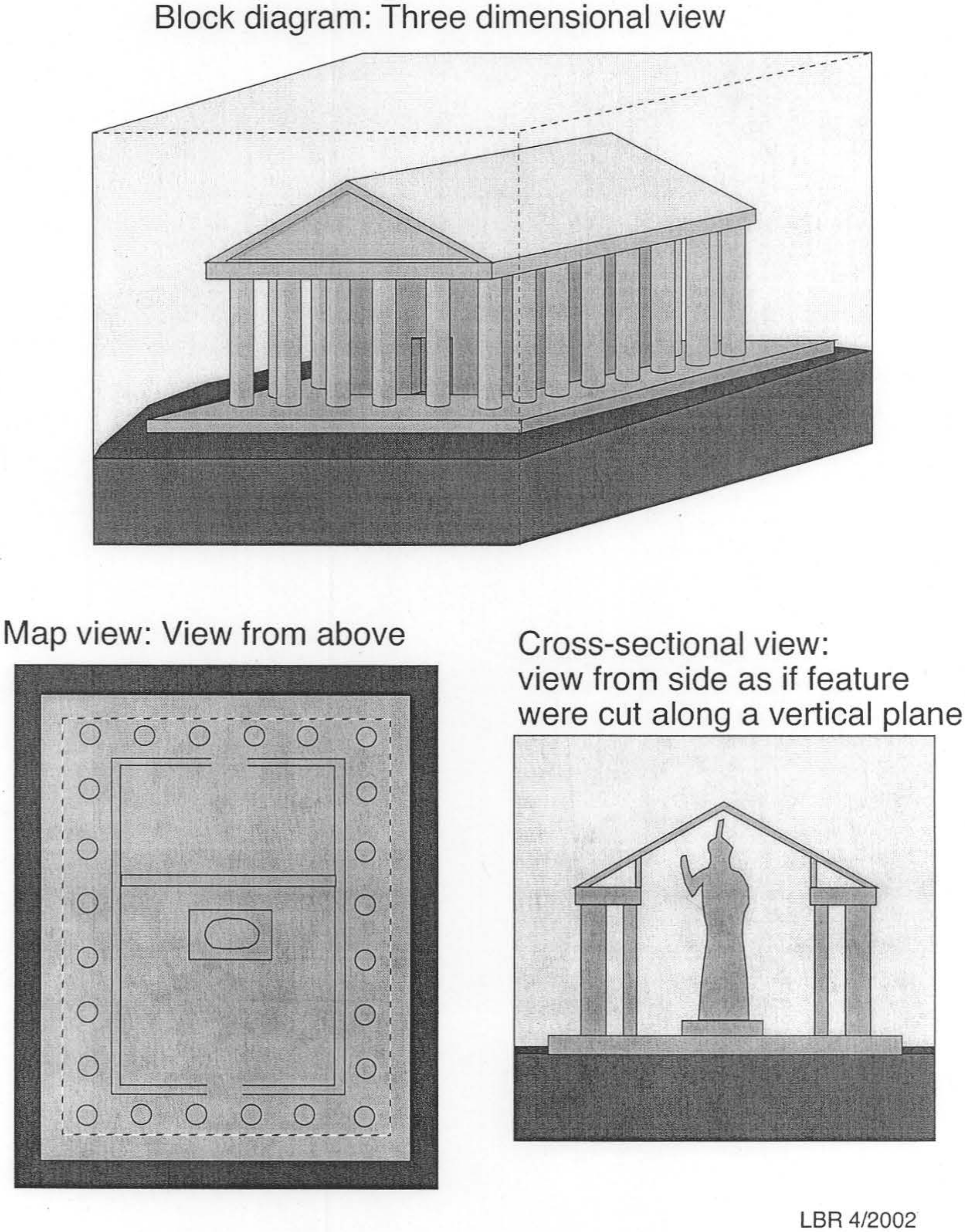


Figure 1-3: Kinds of illustrations used in geology.

CHAPTER 2: MINERALS

Minerals are defined as naturally occurring inorganic crystalline solids with set chemical composition and characteristic physical properties. That's quite a long definition and deserves some explanation. "Naturally occurring" means we're only interested in the solids found in nature, which means that we can ignore all the solid substances synthesized by chemists but not found in natural settings. "Inorganic" means that we can likewise ignore organic chemicals (things with carbon-carbon bonds, which in nature would include the constituents of coal and natural asphalt). "Crystalline" means that the atoms of minerals are arranged in some regular order, rather than being an amorphous jumble (Fig. 2-1). "Set chemical composition" means that we can write a chemical formula for every mineral (for example, SiO_2 for quartz and $CaCO_3$ for calcite), even though many minerals include minor or trace elements (for example, Mg^{2+} and Fe^{2+} can substitute for some of the Ca^{2+} in the $CaCO_3$ of calcite). That leaves us with "characteristic physical properties", which deserve a section of their own below.

Minerals are the fundamental building blocks of geology. Minerals consist of atoms, and a mineralogist or geochemist would take interest in those atoms. However, for a geologist, minerals are the smallest units of matter observed, and so they are the building blocks of which all larger geological entities consist (see the table below). That's why we need to develop a rudimentary understanding of minerals in this chapter, before proceeding to the larger-scale topics of this book.

Scales of observation in Geology,
from smallest to largest:
Atoms
Minerals
Rocks
Rock Layers & Structures
Geologic Provinces
Tectonic Plates
Core, Mantle, & Crust
Planets

Physical properties of minerals

Geologists working in the field identify minerals by the physical properties of those minerals – the physical characteristics that have been found to be typical and hopefully diagnostic of each mineral. Most geologists can identify several dozen and perhaps more than a hundred minerals, but even a beginning student can soon identify the twenty or thirty common minerals by recognizing the properties of those minerals.

One category of physical properties is shape, and the shape of a mineral can be the result of at least three factors. One is the **growth shape** of the mineral, the configuration of crystal faces or facets that form as the mineral forms. Another is **cleavage**, the sets of planes along which a mineral breaks. Thirdly there is **fracture**, the irregular rather than planar surfaces to which a mineral may break, rather than cleaving. For example, fluorite grows in cubes but breaks to yield an octahedral shape, and so it does not fracture. Quartz can grow with six parallel sides that are capped by inclined crystal faces that come to a point, but it does not cleave and instead breaks to give a conchoidal fracture (a fracture of curved or scalloped surfaces). Calcite, on the other hand, often grows to form steep three-sided pyramids, but when broken cleaves to give rhombs. Halite, by comparison, is simple: it grows to form three mutually perpendicular surfaces and cleaves to form three mutually perpendicular surfaces. These four different minerals thus have very different shapes, both in terms of how they grow and how they break.

Another category of physical properties has to do with how minerals reflect light. Seemingly the simplest of these is **color**. Some minerals have characteristic colors; for example, chlorite is always green (and thus gets its name for the Greek word for "green"). Many minerals have typical colors, such as the light green of olivine. However, color can be a fickle trait. For example, olivine can range from white to black, and its usual light green only results from the chemistry of its most common composition. Quartz is commonly colorless, but trace elements in quartz can give us rose quartz (a pink variety), smoky quartz (which can be almost black) and amethyst (the purple semi-precious gemstone).

More trustworthy than color, for some minerals, is **streak**. Streak is the color of a particular mineral when powdered, and the common way to generate that powder is to grind the mineral on a streak plate, a piece of unglazed porcelain. The most common and useful application of this property is to hematite (Fe_2O_3), a mineral that can look black, or metallic and shiny, or red and dull, depending on the size of its crystals and how they are configured. However, the streak of hematite is always red, because the powdered form consistently reflects red light and absorbs other wavelengths.

A third term expressing how a mineral reflects light is **luster**. Luster expresses the extent to which a mineral reflects light, absorbs light, or transmits light. Kinds of luster thus include metallic (reflects light a lot), glassy (reflects some but transmits some), earthy (absorbs much light and so is not shiny), and waxy (reflects at the surface and within).

Hardness expresses whether an individual crystal of a mineral (not a mass of crystals) can scratch another surface, or is scratched by other surfaces. It is thus a relative property that is expressed in terms of the Mohs scale of hardness, which establishes certain minerals, and other materials, as benchmarks on a scale of 1 to 10. A mineral with a greater hardness can scratch any mineral of lesser hardness.

Mohs hardness	*Benchmark mineral*	*Other benchmark material*
1	Talc	
2	Gypsum	Finger nail ~2.2
3	Calcite	Copper coin = 3
4	Fluorite	
5	Apatite	Pocket knife = ~5.2
6	K-Feldspar	Window glass = 5.5
7	Quartz	Steel file = 6.5
8	Topaz	
9	Corundum	
10	Diamond	

A property quantifiable for all minerals and striking for some is **specific gravity**, or (less technically) density. Density is weight per unit volume and so expresses the heft of a chunk of mineral when lifted – does it feel exceptionally heavy or exceptionally light? A good example is galena (PbS), which is very dense because of the lead in it.

Other properties are significant for one mineral or group of minerals. Magnetism is an example, in that magnetite (Fe_3O_4) is magnetic, whereas almost no other minerals are. Reaction with dilute acid is another, in that calcite reacts with dilute acid, a characteristic it shares to varying degrees with other carbonate minerals. For those persons willing to put minerals in their mouths, two more properties emerge, in that kaolinite consists of such a mass of tiny platelets that it adsorbs water sufficiently to stick to one's tongue, and halite (NaCl) tastes salty because it is the stuff of table salt.

Identification of minerals

Geologists in the field, collectors, and students all commonly identify minerals by the physical properties above. In the laboratory, however, at least two other major methods are used. One involves examining rocks and their constituent minerals when cut and polished in thin sections that are by convention 30 microns (0.03 mm) thick. These thin sections are examined under a microscope in transmitted polarized light, and the interaction of light with each mineral is sufficiently distinctive that almost all minerals can be identified, even when individual crystals are less than a millimeter across. This method, called "**optical mineralogy**", is of great use because it allows examination of minerals in their native context, in rocks where they are interlocked with other minerals, rather than after some mechanical extraction that removes them from their immediate geological context in a rock.

The second major method of mineral identification is to pass monochromatic X-rays (X-rays of just one wavelength) through a mineral and to measure the angles at which the X-rays are coherently diffracted, rather than scattered and dissipated. The magnitude of these angles in **X-ray diffraction** depends on the spacing of atomic planes in a crystal (Fig. 2-1), and the number of angles at which diffraction occurs depends on the symmetry of the crystal. This method thus identifies minerals by the spacing of their atomic planes, rather than by their chemical composition. The various configurations and spacings of atomic planes are sufficiently unique for each mineral that X-ray diffraction is considered the fool-proof scientific identifier of minerals, or the "gold standard" of mineral identification among professional geologists and mineralogists.

How minerals form

We said above that "minerals grow", but clearly crystals of minerals form and become larger by very different processes that those of living things. One major way that minerals form and enlarge is by crystallization from molten rock material, which is called magma. The details of that process will be a major topic of Chapter 3, which deals with igneous rocks. The other major way that minerals form and grow is by chemical precipitation from solution (for example, from seawater or groundwater). Every mineral has a fixed limit of concentration, or limiting solubility, in water, and water containing a concentration beyond that equilibrium limit will precipitate the mineral in question. One familiar non-natural example is precipitation of "lime" (really $CaCO_3$) in pipes, and one spectacular natural example is the precipitation of crystals from groundwater in geodes. Of greater significance in geology is precipitation of minerals from groundwater in the intergranular spaces of sediments, in the interparticle spaces of soils, and in fractures in rocks.

Categorization of minerals

Mineralogists have identified about 5,000 minerals, so clearly we need some capability to categorize among this remarkable diversity. We commonly distinguish groups of minerals on the basis of their chemical compositions (Fig. 2-2). These groups include **oxides**, where a cation, a positively charged atom, combines in more-or-less equal proportion with O^{2-}, as for example in hematite (where Fe^{3+} combines with O^{2-} to form Fe_2O_3). Similarly, cations can combine with negatively charged sulfur to make **sulfides**, as for example with galena (where Pb^{2+} combines with S^{2-} to form PbS). Cations can also combine with anions with –1 charge to form **halides**, as for example with halite, where Na^{1+} combines with Cl^{1-} to form NaCl. Ca^{2+} likewise combines with F^{1-} to make the halide mineral fluorite, CaF_2.

In several categories of minerals, O^{2-} anions cluster around a central atom to make a radical group or complex ion. For example, O^{2-} clusters in threes around carbon to make the CO_3^{2-} group that is then the building bock of the **carbonate** minerals like calcite and aragonite, which both have the formula $CaCO_3$ (Figs. 2-2 and 2-3). The same applies when O^{2-} clusters in fours around sulfur to make the SO_4^{2-} group that is then the building bock of the **sulfate** minerals like gypsum, with the formula $CaSO_4 \cdot 2H_2O$, and anhydrite, with the formula $CaSO_4$. There are many other analogous "-ate" groups of minerals, such as nitrates, phosphates, tungstates, arsenates, and selenates. However, because silicon is second only to oxygen in abundance in Earth's crust (Fig. 2-4), by far the largest such group is the silicates, which deserve a section of their own (Fig. 2-5).

Silicate minerals

All silicate minerals contain a fundamental building block consisting of four O^{2-} anions clustered around a Si^{4+} cation (Fig. 2-3). The four O^{2-} anions collectively form a tetrahedron, a three-sided pyramid that, along with its base, has four identical triangular sides. This configuration of silicon and oxygen is called the silicate tetrahedron. The arrangement of these tetrahedra in silicate minerals turns out to have major implications for the structure of the Earth and especially its volcanic mountains.

Silicate tetrahedra are linked in various ways to make different kinds of silicate minerals. **Nesosilicates**, for example, consist of isolated silicate tetrahedra whose O^{2-} anions bond to cations like Mg^{2+}, Fe^{2+}, and Ca^{2+} that lie between the tetrahedral (Fig. 2-3). The mineral olivine, with the formula $(Mg,Fe)_2SiO_4$, is a good example. In **sorosilicates**, on the other hand, every two silicate tetrahedra share one oxygen atom, so that pairs of silicate tetrahedra form dumbbell-like assemblages from which again bonds extend to interspersed cations (Fig. 2-3). An example is the mineral hemimorphite with the chemical formula $Zn_4Si_2O_7(OH)_2 \cdot H_2O$. The main thing to note in this chemical formula is that the ratio of Si to O is now 2 to 7, rather than the 2 to 8 to nesosilicates, because the sharing of one O^{2-} anion in the each pair of tetrahedra lessens the total number of oxygens.

This trend continues with chain silicates, or **inosilicates**, where tetrahedra share O^{2-} anions to form long chains of silicate tetrahedra.[7] In **pyroxenes**, silicate tetrahedra are arranged in lines wherein each silicate tetrahedron shares one O^{2-} with the silicon to one side and another O^{2-} with the silicon to the other side (Fig. 2-3). Each silicon is thus surrounded by two entire oxygens and effectively the halves of two others, giving a Si-to-O ratio of 1to 3, or 2 to 6. This is reflected in the

[7] This discussion of silicates has completely skipped cyclosilicates because they are not essential to understanding the progression of structures covered here and because their most common examples are beryl and tourmaline, neither of which is a common mineral.

chemical formulas of pyroxene minerals like diopside ($CaMgSi_2O_6$) and enstatite ($MgSiO_3$). The chain-like structure of pyroxenes is reflected in elongate crystals with growth faces and cleavages that meet at roughly right angles.

Pyroxenes form single chains, but another category of inosilicate minerals called **amphiboles** forms double or linked chains. Here, two pyroxene-like chains are linked, because every other silicate tetrahedron not only shares two of its O^{2-} anions with its neighbors in its own chain but also shares an O^{2-} with a tetrahedron of matching chain. Thus half of the tetrahedra have a Si-to-O ratio of 2 to 6, and half have a ratio of 2 to 5, yielding an average ratio of 4 to 11 and thus chemical formulas like that of grunerite, with the chemical formula $Fe_7Si_8O_{22}(OH)_2$. The chain-like structure of these minerals is also reflected in elongate crystals, but the double-chains cause growth faces and cleavages to meet at 60° and 120° angles, rather than the ~90° characteristic of pyroxenes.[8]

The next stage in this progression of structures is for every pyramidal tetrahedron to share its three basal O^{2-} anions with neighboring tetrahedra. The linked silicate tetrahredra thus form layers or sheets, and so they are called "**phyllosilicates**" ("leaf-silicates"). Each tetrahedron shares three of its four O^{2-} anions, so that the Si-to-O ratio is 1 to 2 1/2 or 2 to 5, yielding chemical formulae like that of kaolinite ($Al_2Si_2O_5(OH)_4$) and talc ($Mg_3Si_2O_5(OH)_4$). Al^{3+} can substitute for some of the Si^{4+} in the tetrahedra, leading to chemical formulas like that of the mica phlogopite ($KMg_3(AlSi_3O_{10})(OH)_2$). The sheet structure of phyllosilicates leads to strikingly planar growth shapes and cleavages best exemplified by the micas, of which muscovite and biotite are the most common. In fact, muscovite (the clear mica) cleaves in such planar sheets that it was used for window panes in the Middle Ages, when it was called "Muscovy glass" because it was mined near Moscow.

The last stage in this progression is to have each silicate tetrahedron share all four of its O^{2-} anions, so that the silicate tetrahedra form a three-dimensional framework (rather than the two-dimensional sheets in phyllosilicates or one-dimensional chains in inosilicates). These minerals with three-dimensional silicate frameworks are called "**tectosilicates**"[9]. Because each Si^{4+} shares all four of its O^{2-} anions with its neighbors, the Si-to-O ratio in tectosilicates is 1 to (4 x 1/2) or 1 to 2. This is seen in the chemical formula of the very common mineral quartz, SiO_2.

As in phyllosilicates, Al^{3+} can substitute for Si^{4+} in the tetrahedral sites of tectosilicates. This leads to the **feldspar** minerals. Orthoclase, the most common K-feldspar mineral, thus has the chemical formula $KAlSi_3O_8$, and its Na-bearing analog albite has the formula $NaAlSi_3O_8$. In those cases, one of every four tetrahedra is occupied by Al^{3+}. If two of every four tetrahedra are occupied by Al^{3+}, minerals like the feldspar anorthite ($CaAl_2Si_2O_8$) form. In fact, a whole range of compositions of feldspar exists between albite and anorthite, and those minerals are collectively called plagioclase feldspars.

Trends among the silicate minerals

The progression of silicate mineral structures discussed above dictates trends in the properties of these minerals that have great implications for geological processes (Fig. 2-6). One aspect of the progression above is that of chemical composition. For example, Fe^{2+} and Mg^{2+} are abundant in nesosilicate and inosilicate minerals, but they progress to absence in the common tectosilicates. On the other hand, Al^{3+} and Si^{4+} progress from less abundant in nesosilicates and inosilicates to more abundant in phyllosilicates and tectosilicates. We therefore speak of the minerals at the nesosilicate end of this progression as "mafic", for "magnesium-and-iron rich", and the minerals at the phyllosilicate and tectosilicate end as "sialic", for "silicon-and-aluminum rich".[10] "Mafic" and "sialic" are also used to characterize entire rocks, and in fact entire parts of Earth's crust, and even the magmas (molten mineral material) from which these minerals and rocks can form.

The progression from mafic to sialic silicate minerals is a progression from less linkage

[8] Pyroxenes and amphiboles are collectively "inosilicates" in the progression from neso- to soro- to cyclo- to ino- to phyllo- to tecto-silicates.

[9] Persons already familiar with the geological use of the word "tectonics" as the study of large-scale three-dimensional geological structures should see in "tecto-silicate" the corresponding reference to microscopic three-dimensional mineralogical structures.

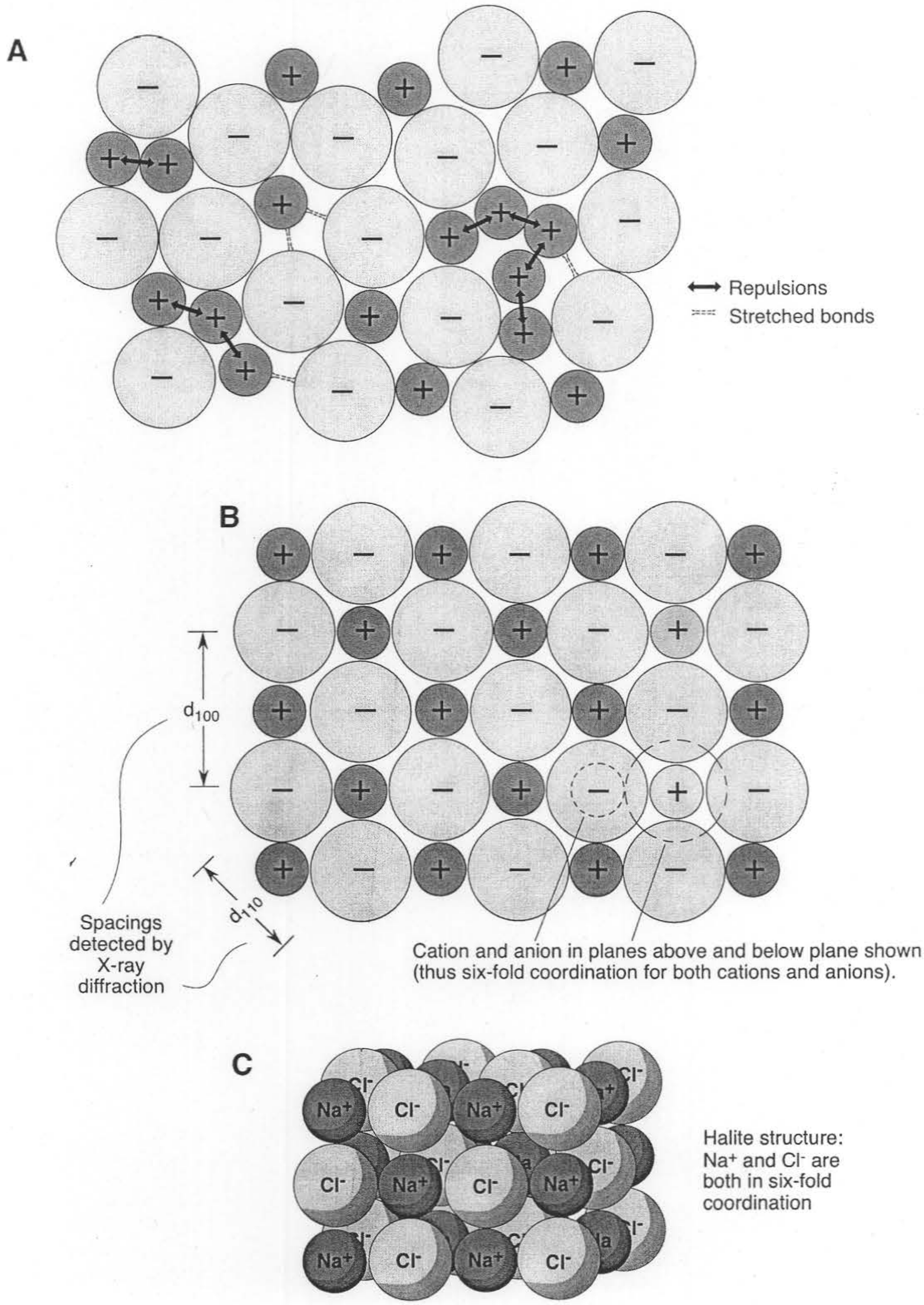
[10] A word essentially synonymous with "sialic" is "felsic", which originated as "feldspar-like".

of silicate tetrahedra to more linkage. The same is true in silicate magmas, where (despite the molten state of matter) Si^{4+} still forms tetrahedra that share O^{2-} anions and that thus link themselves together. These linkages control the magma's viscosity, or its resistance to flow, so that more sialic magmas are more viscous and more mafic magmas are less viscous. This has major implications for the formation of volcanoes, where the tendency of mafic magma to flow and sialic magma to congeal determines both the form and behavior of the volcano.

The progression from mafic to sialic silicate minerals is also one from minerals with higher melting temperatures (or temperatures of crystallization from magma) to minerals with lower melting temperatures (or temperatures of crystallization) (Fig. 2-6). Thus, as a magma cools, more mafic minerals tend to form first, and more sialic minerals tend to form later.[11] The melting of more sialic minerals at lower temperatures in the Earth allows more sialic magmas to form from mafic rocks in the mantle and crust, so that magmas rising to form Earth's crust are more sialic than the underlying mantle.

The progression from mafic to sialic silicate minerals is also one from more dense minerals (largely because of the presence of Fe) to less dense minerals. Where sialic magmas have risen and joined with other sialic rocks at Earth's surface, they have collectively formed the continental crust. By contrast, the oceanic crust, while still more sialic than the mantle, is nonetheless more mafic than the continental crust. The lesser density of the continental crust is what lets it sit higher across Earth's surface than the more dense oceanic crust of the sea floor. The difference in density between sialic and mafic minerals is thus why we have an Earth with continents rising above sea level and ocean basins below sea level. The difference in density between sialic and mafic crust is thus the reason for the global geography that we take for granted as land-dwelling animals perched on our concentrates of sialic rock.

[11] This temporal progression from crystallization of mafic to crystallization of felsic minerals is called "Bowen's Reaction Series" after American petrologist N.L. Bowen. The word "reaction" denotes Bowen's observation that the early-formed more mafic minerals do not sit unchanged in the magma that has become progressively more sialic as mafic minerals form. Instead, these early-formed minerals commonly react with the magma and commonly break down, at least in part, to form more sialic minerals. Thus olivine and pyroxene crystals formed early from magmas commonly have outer rims of pyroxene or amphibole, respectively. The same is true among plagioclase feldspars, where anorthite forms at higher temperatures than albite but may then react to form a more albitic plagioclase feldspar or at least to have rims of more albitic plagioclase.



A: A two-dimensional sketch of a hypothetical disordered arrangement of Na^+ and Cl^- ions. Repulsions between cations and large cation-anion distances (and thus weak bonds) make this arrangement unstable. B. Ordered halite structure provides a stable arrangement that maximizes cation-cation distances and thus cation-cation repulsions, with each Na^+ surrounded by six Cl^- ions. C. Three-dimensional representation of the halite structure.

Figure 2-1: Sketches illustrating crystallinity (B and C) or lack thereof (A).

Non-Silicate Minerals

Native elements
- Gold (Au)
- Silver (Ag)
- Copper (Cu)
- Sulfur (S)
- * Diamond (C)
- * Graphite (C)

Polymorphs

Oxides
- * Hematite (Fe_2O_3)
- * Magnetite (Fe_3O_4)
- Goethite (FeOOH)
- Ilmenite ($FeTiO_3$)
- * Bauxite ($\sim Al_2O_3$)
  (actually a mixture of Al oxides & hydroxides)
- Rutile (TiO_2)

Sulfides
- * Pyrite (FeS_2)
- Galena (PbS)

Sulfates
- * Anhydrite ($CaSO_4$)
- * Gypsum ($CaSO_4.2H_2O$)

Evaporite minerals

Halides
- ** Halite (NaCl)
- Fluorite (CaF_2)

Carbonates
- ** Calcite ($CaCO_3$)
- * Aragonite ($CaCO_3$)

Polymorphs

- * Dolomite ($CaMg(CO_3)_2$)

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Figure 2-2: A list of some non-silicate minerals

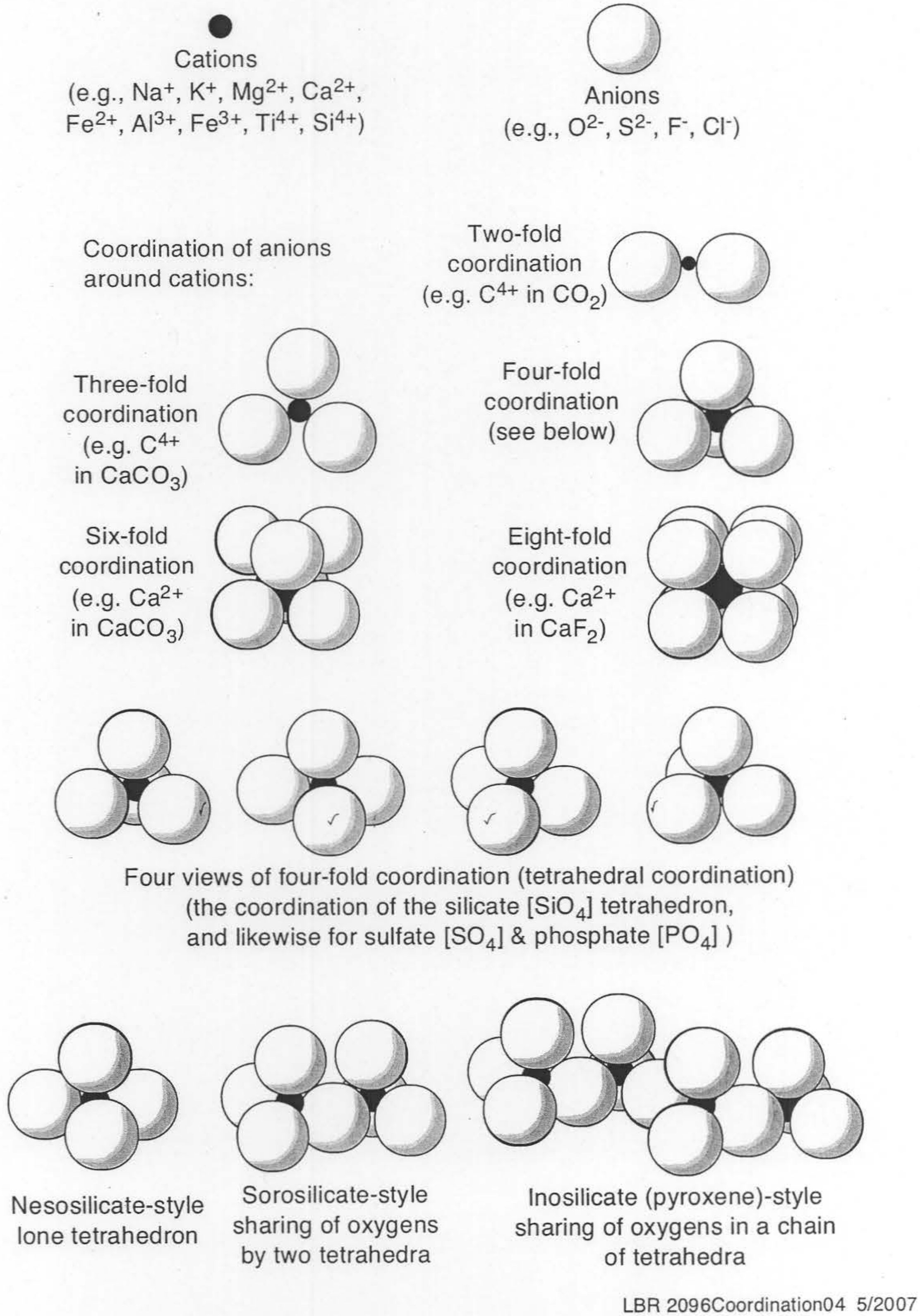

Figure 2-3: Coordination of cations and anions, and resulting silicate structures.

Composition of Earth's Crust

Element	Common charged Form	Wt% in Earth's crust	Resulting mineral groups
O (Oxygen)	O^{2-}	46.4	Oxides
Si (Silicon)	Si^{4+}	28.2	Silicates
Al (Aluminum)	Al^{3+}	8.1	
Fe (Iron)	Fe^{3+}	5.4	
Ca (Calcium)	Ca^{2+}	4.1	
Na (Sodium)	Na^{2+}	2.4	
Mg (Magnesium)	Mg^{2+}	2.3	
K (Potassium)	K^{+}	2.1	
Ti (Titanium)	Ti^{4+}	0.5	
H (Hydrogen)	H^{+}	0.14	
P (Phosphorus)	P^{5+}	0.11	(Phosphates)
Mn (Manganese)	Mn^{4+}	0.10	
F (Fluorine)	F^{-}	0.065	Halides
Ba (Barium)	Ba^{2+}	0.05	
Sr (Strontium)	Sr^{2+}	0.38	
S (Sulfur)	S^{2-} & S^{6+}	0.030	Sulfates & Sulfides
C (Carbon)	C^{4+} & C^{4-}	0.022	Carbonates
Zr (Zirconium)	Zr^{4+}	0.017	
Cl (Chlorine)	Cl^{-}	0.013	Halides

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Figure 2-4: The chemical composition of Earth's crust

Some Silicate Minerals

Nesosilicates

* Olivine -- $(Mg,Fe)_2SiO_4$ Garnet* -- $(Ca,Mg,Fe^{2+})_3(Al,Fe^{3+},Cr^{3+})_2Si_3O_{12}$

Pyroxene Group
e.g., augite: $(Ca, Na)(Fe, Mg, Al)(Si,Al)_2O_6)$

Chain silicates

Amphibole Group
e.g., hornblende: $(Ca,Na)_{2\text{-}3}(Fe, Mg, Al)_5Si_6(Si,Al)_2O_{22})(OH)_2)$

Minerals that form at high tempera-ture (>300°C)

Biotite (black mica) -- $K(Fe, Mg)_3(AlSi_3O_{10})(OH)_2$

Muscovite (white mica) -- - $KAl_2(AlSi_3O_{10})(OH)_2$

Sheet silicates

* K Feldspar -- $KAlSi_3O_8$

Framework silicates

** Quartz -- SiO_2

Minerals that form at low tempera-ture (<100°C)

* Kaolinite -- $Al_2Si_2O_5(OH)_4$

Smectite* --
$(Na, Mg)\cdot nH_2O(Mg,Al,Fe)_{2\text{-}3}(Al_{0\text{-}\frac{1}{2}}Si_{3\frac{1}{2}\text{-}4})O_{10}(OH)_2$

Clay minerals

* "Garnet" and "smectite" are actually names for groups of minerals

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Figure 2-5: A list of some silicate minerals

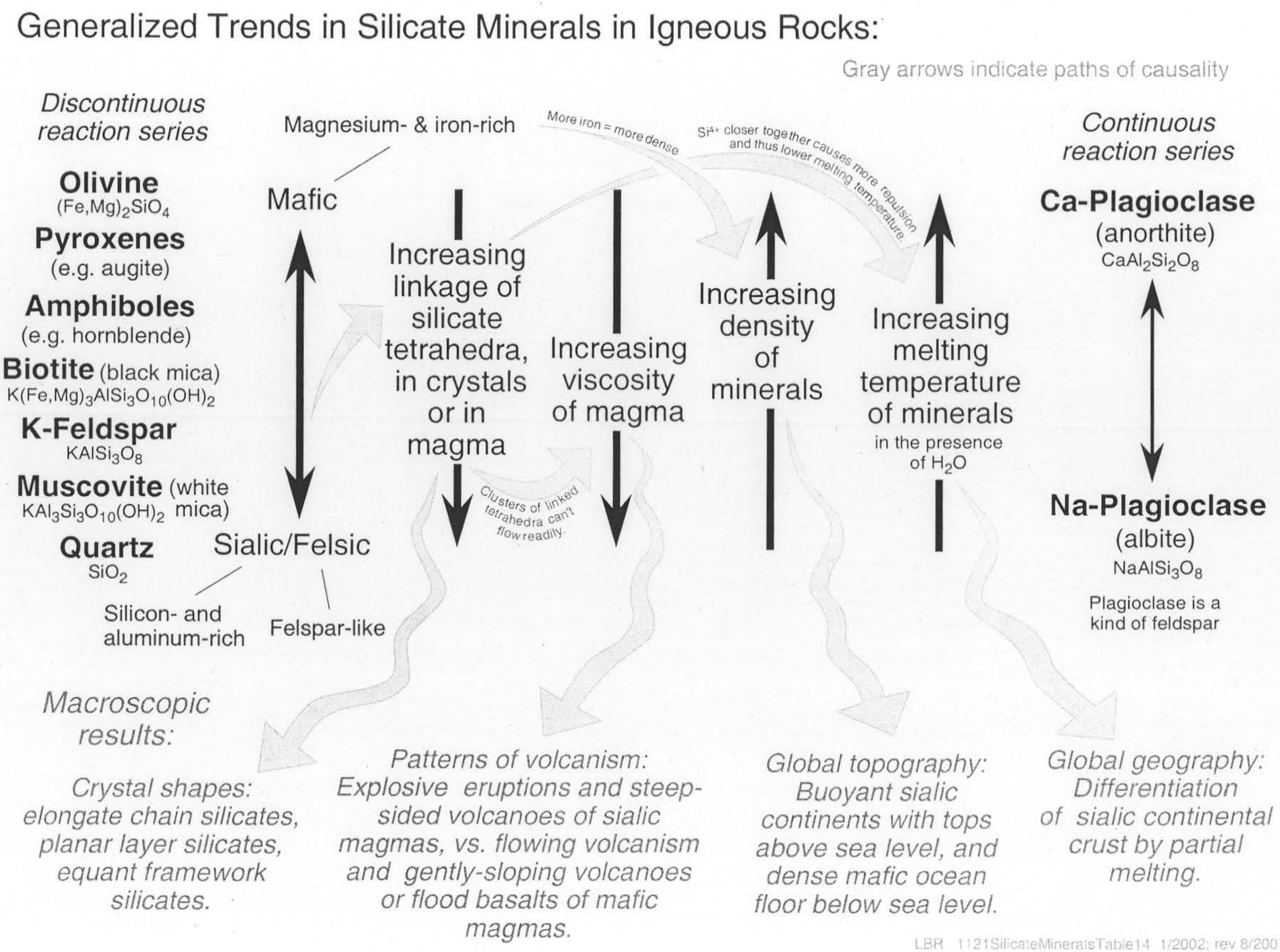

Figure 2-6: Trends in silicate minerals

CHAPTER 3: IGNEOUS ROCKS

Geologists recognize three fundamental kinds of rocks: igneous, sedimentary, and metamorphic. Petrology is the study of rocks, and petrologists are those geologists who specialize in studying the properties of rocks and the processes that form rocks.[12]

Igneous rocks are the rocks formed by the cooling of magma, which is molten mineral material.[13] In almost all cases, they consist largely of silicate minerals.[14] The magmas from which they form originate either in Earth's mantle or in the lower crust, and the buoyancy of those magmas relative to solid rock causes them to rise through the crust, seeping upwards through fractures. The results are both volcanoes, when magmas reach the surface of the Earth, and bodies of coarser-grained igneous rock that form when the ascent of magmas stops within the crust. Igneous rocks are categorized by their textures and compositions.

Textures of igneous rocks

Textures of igneous rocks range from rocks so fine-grained that crystals are invisible, and in some cases not present at all, to rocks consisting of visible crystals, some of which may be centimeters across. The principal control on variation in texture is cooling history. A magma cools slowly if its ascent stops within the crust and thus amidst hot rocks. With slow cooling, atoms in the cooling magma have time to migrate to the surfaces of growing crystals, and so crystals grow large (Fig. 3-1). The result is a coarse-grained rock in which all the crystals are visible. This kind of rock is termed "phaneritic" (from the Greek word for "visible"), "intrusive" (because the magma was intruded amidst pre-existing rocks), or "plutonic" (from the Roman god of the underworld, and in allusion to the rock's formation within the Earth). On the other hand, a magma cools quickly if it reaches Earth's surface, and atoms have little time to migrate to growing crystals. Instead, many small crystals nucleate, and none grow large (Fig. 3-1). The result is a fine-grained rock in which few or no crystals are visible to the naked eye. This kind of rock is termed "aphanatic" ("not visible" in Greek), "extrusive" (because the magma was extruded at Earth's surface), or "volcanic" (a common word because we see volcanoes, but a word derived from the name of the Roman god of the forge, and thus in parallel with "plutonic").

Some magmas have histories intermediate between the two cases above. A magma may reside for a while within the crust and thus cool to the temperature at which some crystals form, and then it may move upward and cool quickly at or near Earth's surface. The resulting rock thus has some visible crystals amidst a mass of fine-grained ones. Such a rock is a porphyry or is porphyritic.

Cooling of volcanic rocks may be so fast that few or no crystals form, and instead the magma freezes as a glass. The resulting rock is obsidian, the glass-like appearance of which reflects its glassy structure. A volcanic magma may also cool while trapping gasses that have exsolved as a separate phase in bubbles but not had to time to escape from the magma. Such bubbles are called "vessicles", and a rock rich in them is called a "scoria" (Fig. 3-2).

Compositions of igneous rocks

Almost all magmas are rich in silicates, and analyses of igneous rocks expressed in terms of weight per cents of oxides reveal that almost all are 30% to 80% SiO_2. Igneous rocks rich in silicon are also commonly rich in aluminum, and so they are referred to as "sialic".[15] On the other hand, rocks relatively poor in silicon (but still typically more than 40 % SiO_2 by weight) are commonly rich in magnesium and iron (Fe), and so they are referred to as "mafic".[16] Rocks extremely rich in Mg and/or Fe are called "ultramafic".

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[12] Petrology is the study of rocks; the word is derived from the Greek word "petros" for rock. Petrology has nothing to do with petroleum ("rock-oil"), other than that some sedimentary petrologists make their livings contributing to the effort to get oil out of sedimentary rocks.

[13] More accurately, magmas consist of molten mineral material and volatiles. These volatiles are gases such as water vapor, SO_2, CO_2, etc. Those gases are significant in the physical behavior of the magma and for their contribution to the atmosphere.

[14] The exception to this generalization is carbonatites, igneous rocks in which carbonate minerals are the dominant component.

[15] Another word essentially synonymous with "sialic" is "felsic", or "feldspar-like".

[16] In a manner of speaking no longer used by igneous petrologists, mafic rocks are called "basic" because their

Compositions of igneous rocks can be related in a general way to their color. Fe-bearing minerals are commonly very dark green to black, whereas minerals in which Si^{4+} and/or Al^{3+} are the only cations are typically white or clear[17]. Thus mafic rocks are commonly dark-colored, whereas sialic rocks are light colored, and those intermediate in composition are intermediate in color too.

The composition of a magma determines its viscosity, and thus the behavior of that magma, especially during volcanic eruption[18]. The relative scarcity of linked silica tetrahedra in mafic magmas causes them to have low viscosity, and so they erupt as flows. On the other hand, the greater linkage of silica tetrahedra in sialic magmas makes them viscous, and the high viscosity of such magmas causes them plug their eruptive vents and then to erupt explosively. As a result, many sialic volcanic magmas are erupted as volcanic ash, fine airborne volcanic material that settles to form non-porous rocks called tuffs. Less commonly, ash particles settle and collectively trap inter-particle space so that they make porous rocks called pumice that can contain so much enclosed gas-filled space that they float in water.

Classification of igneous rocks

Simple classifications of igneous rocks depend on the textural distinctions above between plutonic and volcanic rocks, and the compositional distinction above between sialic, intermediate, and mafic rocks. The result is a matrix of such fundamental rock names: granite, diorite, and gabbro for plutonic rocks, and rhyolite, andesite, and basalt for volcanic rocks[19] (Fig. 3-2).

The six-name system is used in introductory geology classes, but more advanced understandings of igneous petrology require more detailed classification schemes. The International Union of Geological Sciences has established a standardized quantitative nomenclature for igneous rocks, and this nomenclature is widely used by advanced students and professional geologists (Figs. 3-3 to 3-5). This scheme requires estimation of the abundance of different minerals in each rock, with the results plotted on ternary diagrams to determine rock names. In this scheme, granites are divided into several groups, including "true granites", monzonites, and syenites, and the same division occurs among rocks known in the simple classification as rhyolites. Gabbroes are divided according to a ternary diagram of their own into rocks like norites, harzburgites, dunites, etc. Basalts, being more inscrutable, are not.

Bodies of plutonic rock

Plutonic rocks can form irregularly shaped bodies known appropriately as "plutons" (Fig. 3-6). A relatively large such body is called a "stock", and a very large such body is a batholith (a "deep rock"). For example, a body of granitic rock occupying much of central Idaho is known as the Idaho Batholith, and the granites of the Sierra Nevada mountains are known collectively as the Sierra Nevada Batholith.

Plutonic rocks also commonly form sub-planar bodies where magma has been injected into fractures. Where these bodies cut across pre-existing rock structure, they are called "dikes" or "dykes".[20] Where these bodies follow pre-existing rock structure that was nearly horizontal (for example, along bedding planes), they are called

weathering can buffer the acidity of groundwater, and sialic rocks are called "acidic" because they are less effective at buffering such acidity as they weather.

[17] More generally, cations with some remaining outer-shell electrons give color to their minerals, whereas hard cations (cations with no outer-shell electrons) provide no color. Thus minerals with Mn^{2+}, Mn^{3+}, Fe^{2+}, Fe^{3+}, Ni^{2+}, Cu^{+}, Cu^{2+}, and Zn^{2+} commonly have colors characteristic of those cations, whereas minerals containing only Na^{+}, K^{+}, Mg^{2+}, Ca^{2+}, Al^{3+}, Si^{4+}, and Ti^{4+} are commonly white or colorless. However, other crystallographic features, such as defects, can generate color too.

[18] Viscosity is resistance to flow. Water is not very viscous; cold molasses is very viscous.

[19] The word "granite" is commonly used by builders to categorize any non-carbonate rock that can be polished, so that basalts and gneisses become "granites" in the building trade. "Andesite" logically takes its name from the Andes, where volcanic rocks of intermediate composition are common. "Basalt", on the other hand, has nothing to do with salt.

[20] The name originates because dikes of igneous rock cut across geology as discordantly as human-made hydrologic dikes can cut across a landscape. In fact, where igneous dikes erode more slowly than the rocks into which they were intruded, they commonly stand above landscape like walls, or like hydrologic dikes.

"sills" because of their resemblance to wooden beams or sills in old-fashioned wooden houses and barns. The pre-existing rocks into which plutonic rocks are intruded are called "county rocks" (Fig. 3-6).

Bodies of volcanic rock

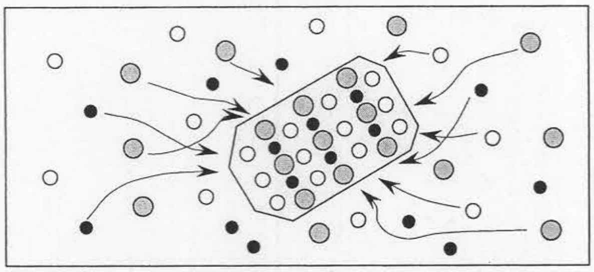
Bodies of volcanic rock obviously include volcanoes. Volcanoes formed by mafic (i.e., basaltic) magmas are typically broad with gentle slopes, because the mafic magmas flow so readily. These volcanoes are called "shield volcanoes" because their shape is like that of a shield facing upwards (Fig. 3-7). The volcanoes of Hawaii are the classic examples of shield volcanoes. Volcanoes formed by more sialic magmas (i.e., andesitic and rhyolitic volcanoes) are more steep-sided, because their magmas do not flow. These volcanoes are called stratovolcanoes or composite volcanoes, in both cases because they are more distinctly layered than shield volcanoes. Mt. Fuji is the classic example.

In addition to volcanoes, there are other bodies of volcanic rock (Fig. 3-7). The most common of these on land are flood basalts, which are widespread flows of basalt that form no cone at all. Instead, the basaltic magma flows so readily across a landscape that it "floods" the region. At sea, basalts are erupted at the mid-ocean ridge and make up the floor of the oceans, so that the most widespread, if least noticed, basalts are those of the deep sea-floor.

Cooling of magma to make an igneous rock:

Slow cooling of magma within the crust:

Time allows atoms to migrate to crystals, so crystals are large.



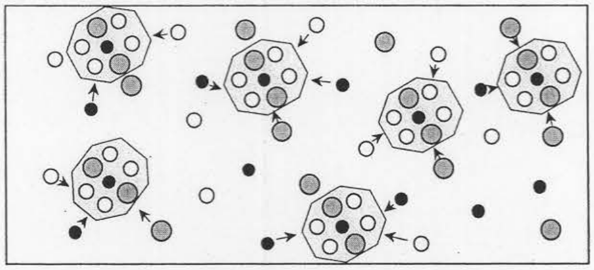
This makes a . . .

coarse-grained rock because the crystals grow large.

Synonyms { "**Intrusive**" rock because the magma is intruded into the crust.
"**Plutonic**" rock because Pluto is the god of the earth's interior.

Fast cooling of magma at the Earth surface:

Time does not allow atoms to migrate to crystals much, so crystals are small.



This makes a . . .

fine-grained rock because the crystals remain small.

Synonyms { "**Extrusive**" rock because the magma is extruded from the Earth.
"**Volcanic**" rock because Vulcan is the god of the forge.

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Figure 3-1: Development of textures of igneous rocks

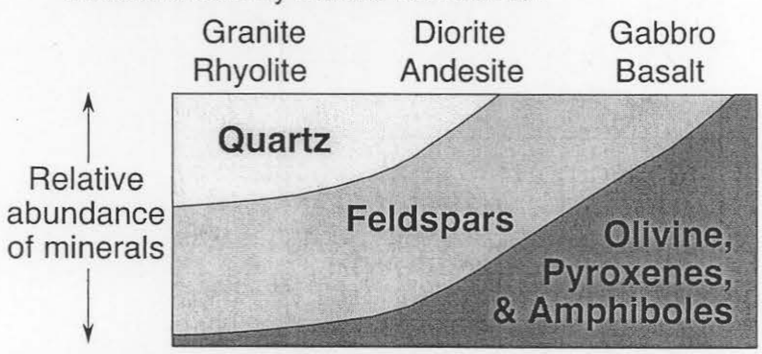
Simple Classification of Igneous Rocks

Six fundamental rock types defined by composition and texture, and some other names defined largely by texture

Composition:

Texture:	Sialic/Felsic Light color	Intermediate	Mafic Dark Color
Extremely coarse-grained	Pegmatite		
Coarse-grained or intrusive or plutonic	**Granite**	**Diorite**	**Gabbro**
Coarse crystals amidst fine		Porphyries	
Fine-grained or extrusive or volcanic	**Rhyolite**	**Andesite**	**Basalt**
Very fine-grained (glassy)	Obsidian		
Very fine-grained (glassy and bubbly)			Scoria
Consolidated pyroclastic material — Dense		Tuff	
Consolidated pyroclastic material — Porous		Pumice	

The same table by mineral abundance:



after Fig. 5.3 of Montgomery's *Environmental Geology* (5th edn)

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Figure 3-2: Simple classification of igneous rocks

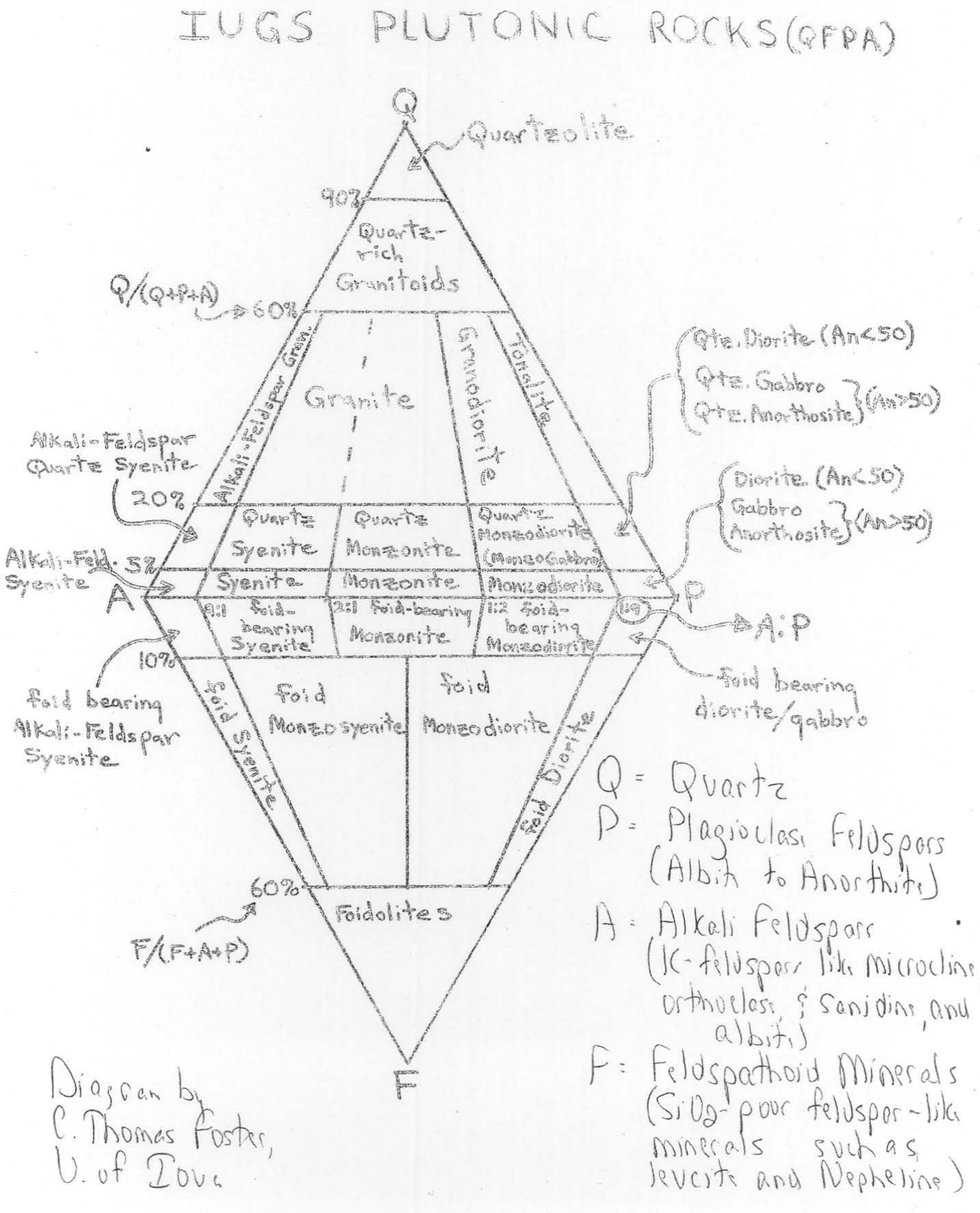

Figure 3-3: Detailed classification of plutonic igneous rocks

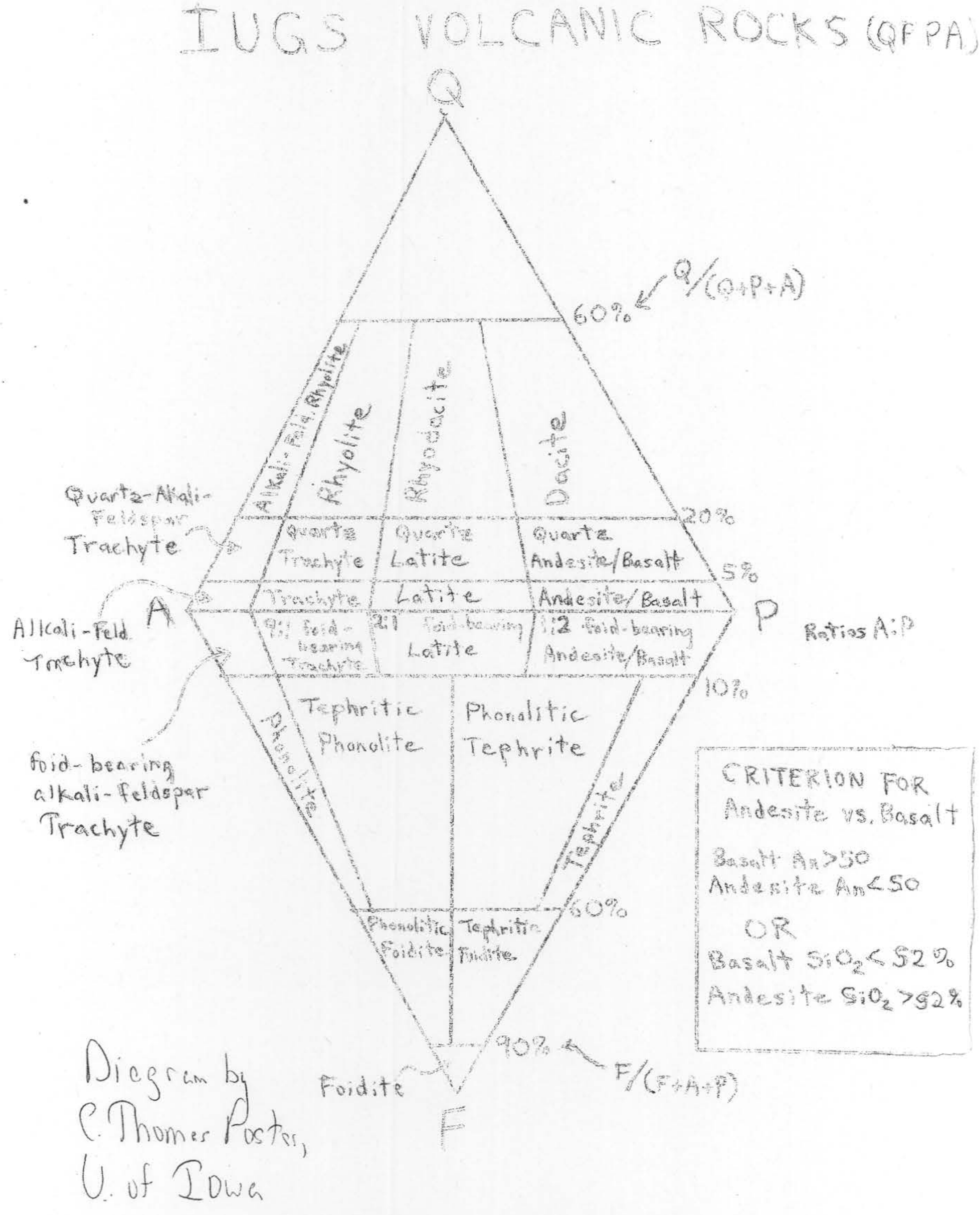

Figure 3-4: Detailed classification of volcanic igneous rocks.

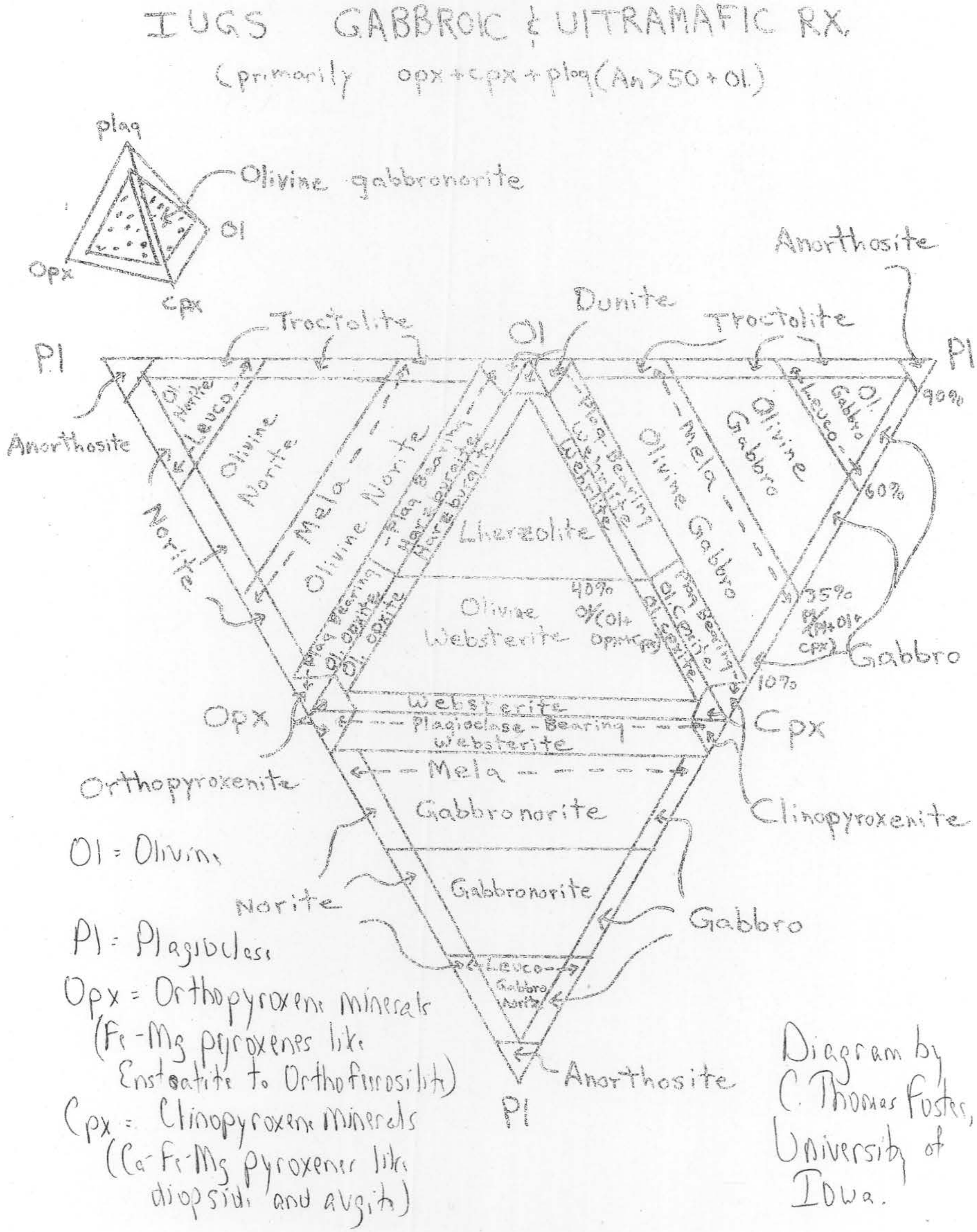

Figure 3-5: Detailed classification of mafic to ultramafic plutonic igneous rocks

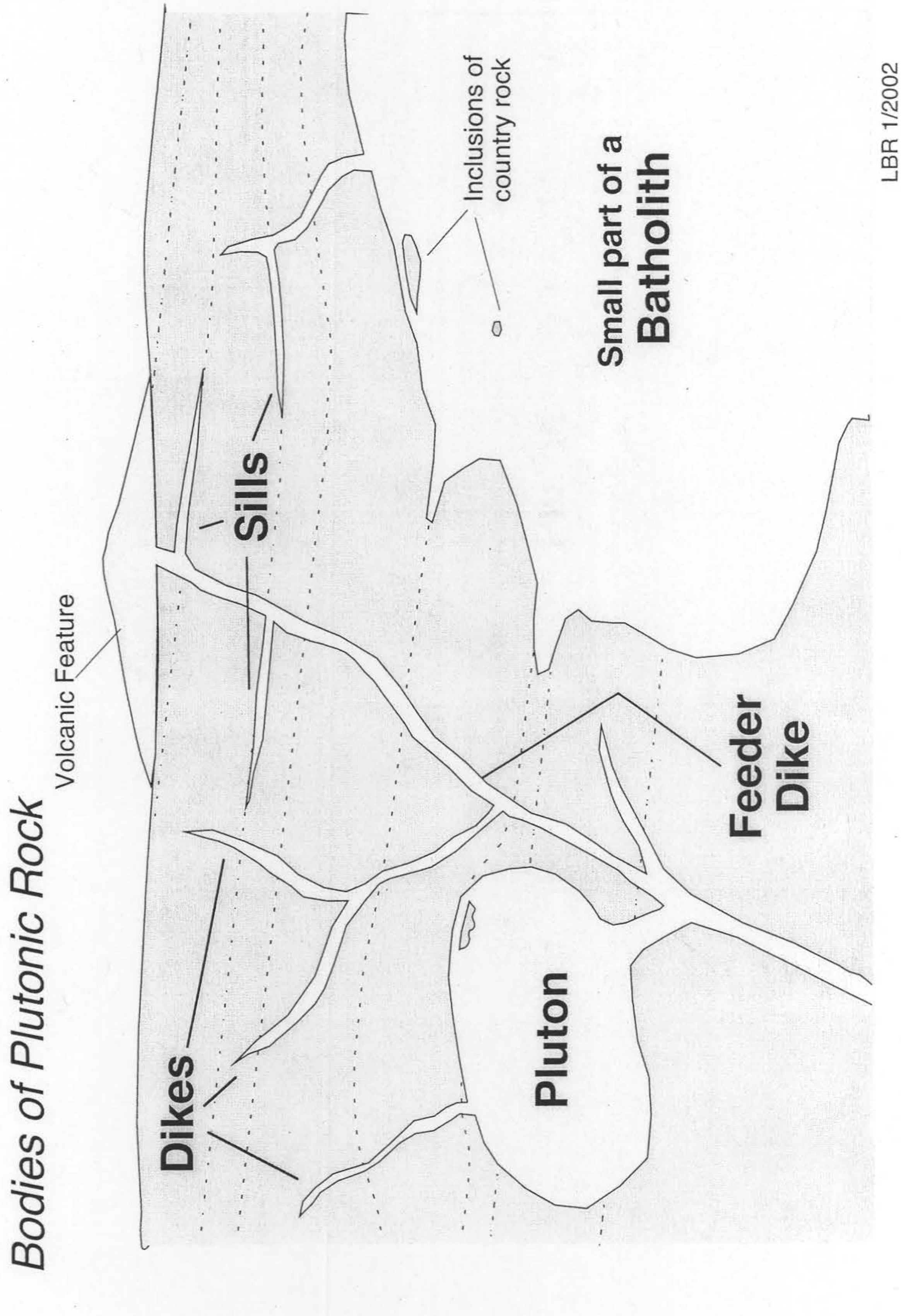

Figure 3-6: Sketches illustrating bodies of plutonic igneous rocks and their generic names.

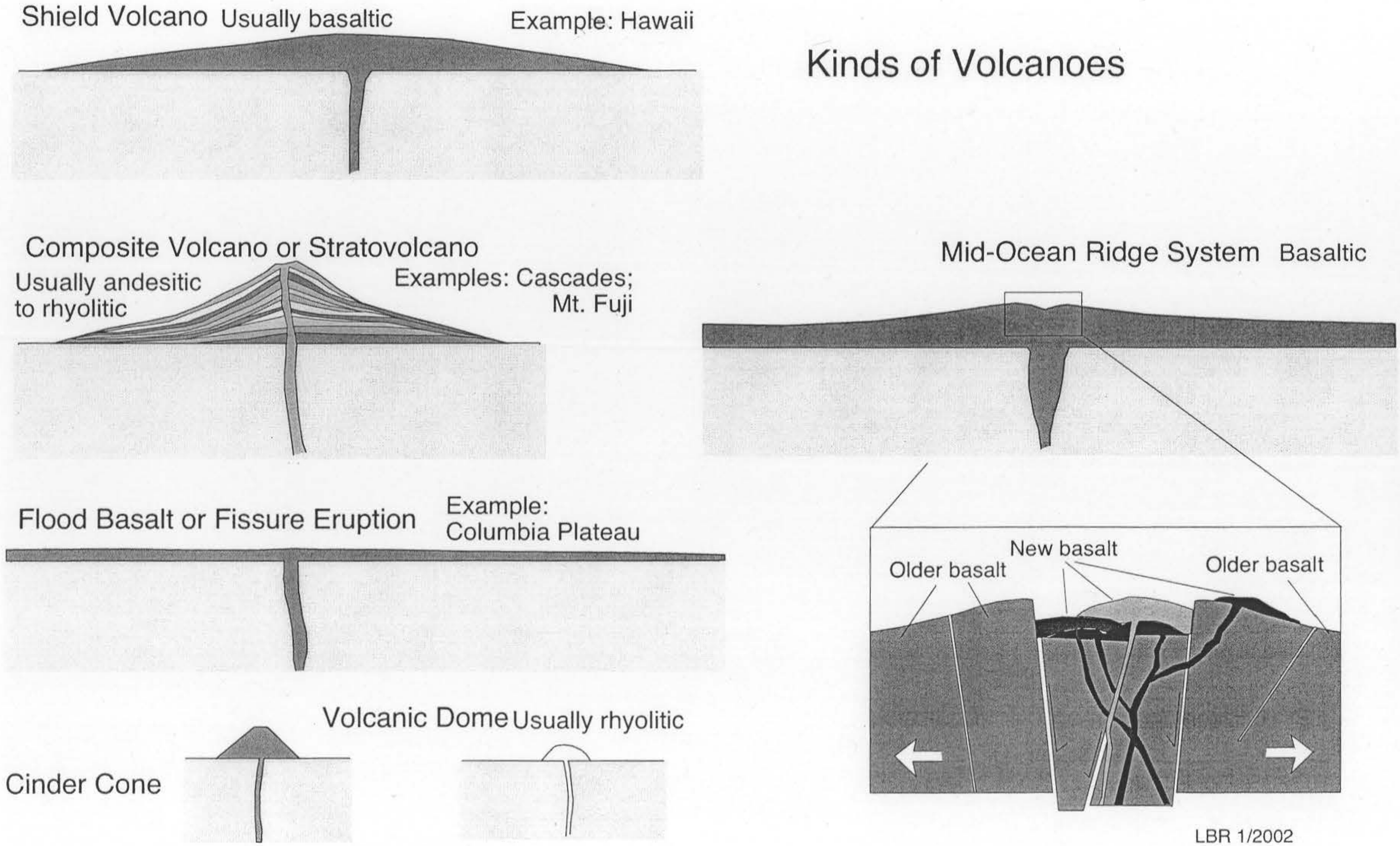

Figure 3-7: Sketches illustrating bodies of volcanic igneous rocks and their generic names.

CHAPTER 4. SEDIMENTARY ROCKS

Sediments are naturally occurring deposits, usually of some mineral material, that are deposited by air, water, or ice onto the solid surface of the Earth. Dust settling from wind, sand deposited by rivers, shell beds in the ocean, and tills left by glaciers are all examples of sediments. Sedimentary rocks are the rocks formed by consolidation of these sediments into solid masses.

The most obvious characteristic of sedimentary rocks is their layering. Layering develops because one kind of sediment is typically deposited across a broad area, and then a different kind of sediment is deposited atop the first, and then a third atop the second, to yield distinct layers. The differences may be subtle, but the result is layers that are commonly visible miles away (Fig. 4-1).

Sediments and sedimentary rocks are first divided by their composition, which has much to do with their origin. **Siliciclastic sediments,** as the name implies, consist of clasts of pre-existing silicates, the common minerals of igneous rocks (and of other types as well). These clasts originate in the weathering and erosion of pre-existing silicate rocks, and the physical transport of those particles or clasts down rivers or streams to sites of deposition (Fig. 4-1). Muds, sands, and gravels are examples. In contrast to siliciclastic sediments, **biochemical sediments** are formed by organisms using the solutes dissolved in seawater or less commonly lake waters, and/or are formed inorganically by chemical precipitation of those solutes. Deposits of seashells and their debris (as found on the sandy beaches of the world), deep-sea deposits of the mineralized tests of plankton, and sea salt are examples of these biochemical sediments.

Given this first-order division by composition, these groups are secondarily categorized by particle size. This is inevitably true among the siliciclastics, and in part true among the biochemical sediments called limestones.

Siliciclastic sediments and sedimentary rocks

Siliciclastic sediments range in particle size from clay minerals less than four microns (0.004 mm) in size, to silt-size particles (0.004 mm to 0.06 mm in size), to sand (0.06 mm to 2 mm), to gravel (particles more than 2 mm in size, and sometimes far larger) (Fig. 4-2). Clays, or clays and silt, collectively make up mud, both in the common sense and in the technical geological sense of that word.

Consolidation of these siliciclastic sediments generates siliciclastic rocks. In the simplest classification, clays lithify to form **shales**, silts lithify to form **siltstones**, sands lithify to form **sandstones**, and gravels lithify to form **conglomerates** (Fig. 4-3). In more rigorous approaches, sandstones are divided into mud-free sandstones (arenites) and muddy sandstones (wackes, pronounced "wack-ees"), each of which is further subdivided by the mineralogy of their grains. Thus any one sandstone might be a quartz wacke, and another might be feldspathic arenite (Fig. 4-4).

Biochemical sediments: Limestones

The most abundant biochemical sediments are limestones, rocks consisting of $CaCO_3$ (calcium carbonate). Many of these are biological in origin, in that they consist of pieces of shells (or sometimes whole shells) of marine invertebrate organisms. In other cases, they seem to be more purely chemical in origin, where seawater has precipitated $CaCO_3$ either to form ooids (small spherical concretions of $CaCO_3$) or fine particles of $CaCO_3$ that collectively form carbonate mud.

Simple classifications of **limestones** recognize "limestone" (usually a featureless rock consisting of carbonate mud or small pieces of fossils), fossiliferous limestone (a limestone with invertebrate fossils, or at least recognizable pieces thereof), and coquina (a recently formed limestone consisting only of small shells). More serious work uses the Dunham classification of limestones, which distinguishes whether mud is present between the grains and, if so, whether the mud supports the grains (a mud-supported fabric) or if the grains rest on each other with mud in the resulting interstices (a grain-supported fabric) (Fig. 4-4).

Another kind of limestone outside both of these classifications is chalk. Chalks are rocks consisting of the microscopic $CaCO_3$ tests of marine plankton (Fig. 4-3). Their fine particle size and softness made pieces of chalk useful as writing instruments. The analogous rock consisting of microscopic SiO_2 tests (siliceous tests) of marine plankton is chert.[21]

Dolomites and dolostones

Limestones consist largely of calcite, a $CaCO_3$ mineral. $CaCO_3$ can be altered with the addition of magnesium to the mineral dolomite, which has the chemical formula $CaMg(CO_3)_2$. This dolomitization of calcite is common in nature, and thus many limestones have been converted to dolomite, seemingly via processes in which Mg-rich water dissolves small amounts of calcite and in its place precipitates dolomite, until over millions of years entire stratigraphic layers of limestone hundreds of feet thick are converted to dolomite. The resulting rocks are commonly called "dolomites", although the word "dolostone" better keeps separate the distinction of mineral dolomite and bodies of rock consisting largely (but rarely entirely) of that mineral.

The extent of dolomitization can be seen in the lateral extent of Ordovician and Silurian dolostone strata that cover much of the midwestern U.S. from Minnesota to Ohio. The significance of dolomitized limestone to alpine geology can be seen both in the Bighorn Dolomite, the Ordovician dolostone that mantles much of Wyoming's Bighorn Mountains, and in the entire range of Europe's southern Alps, the famously rugged mountains known as the Dolomites.

Other biochemical sediments

One obviously biological sediment is the remains of plants and trees that accumulate as peat and lithify to form coal. On the other hand, the most clearly chemical sediments are the salts precipitated inorganically from sea water. These commonly include the minerals gypsum ($CaSO_4 \bullet 2H_2O$) and halite (NaCl), the two most abundant evaporite minerals. When lithified, they are known simply by their mineral names, or halite is known as "rock salt".

The significance of sedimentary rocks

Sedimentary rocks are significant for at least four reasons. Firstly, they cover about three quarters of Earth's land surface, so they are the rocks we encounter most commonly. Secondly, they are the rocks that house fossil fuels, in that petroleum is extracted from the pores of sandstones and limestones, and coal is in itself a major energy resource.

Thirdly, sedimentary rocks are records of what the Earth surface was like in the past. Each sediment has particular conditions under which it forms, and so each sediment is an indicator of a particular environment of deposition (Fig. 4-3). Lateral distributions of sediments from one particular time thus provide a paleogeographic map of the Earth surface at that time, and vertical sequences of marine or near-shore land sediments provide a record of changes of sea level. These concepts combine to provide much of our understanding of Earth's paleoenvironmental history. To alpine geologists, an appreciation of the depositional history of sedimentary rocks is additionally important when sedimentary rocks deposited on the sea floor are found high in inland mountain ranges.[22]

Fourthly, compositions of sandstones are commonly used to infer sources of sediments and thus to make paleogeographic inferences. For example, quartz-rich sands are observed today to originate largely from the weathering and erosion of lowland landscapes consisting of sedimentary rocks. Sands rich in potassium feldspar commonly originate in uplifted regions of granitic rocks. Ancient examples of these thus imply source areas of these rock types and landscapes. Sandstones rich in fragments of pre-existing rocks are wealth of such information. For example, a sandstone rich in fragments of metamorphic rocks implies an uplifted area of metamorphic rocks, and perhaps a mountain range. Sandstones shed by mountain ranges now lost to erosion may thus be our best, and perhaps only evidence, of those ancient mountain ranges and of the kinds of rock that were exposed in them.[23]

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[21] The word "chert" is also used to characterize silicified limestone. Because silicification of limestone commonly produces nodules of chert, the expression "layered chert" is sometimes used to distinguish planktic cherts from nodular cherts generated by silicifcation.

[22] This is, for example, the case in the Alps north of Innsbruck, where marine limestones make up the Nordkette.

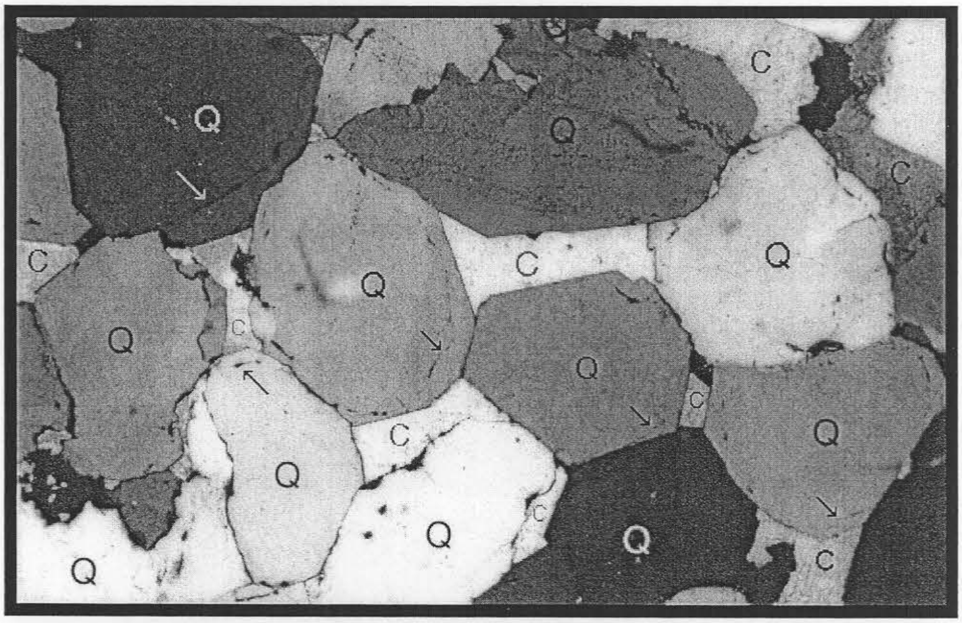
[23] The source of a sediment is commonly called its provenance. Provenance refers collectively to the rock type, topographic relief, and climate of the source area, because all of these affect the composition of the sediment produced by weathering and erosion.

Lithification of sediments to form sedimentary rocks

Two processes account for the solidification of sediments to form sedimentary rocks (Fig. 4-5). One is **compaction**, the physical rearrangement of sedimentary particles to reduce inter-particle space and to pack grains together more closely. In the process, grains may be squeezed together sufficiently to weld them at their contacts, making sediment into solid rock.

The second major process in lithifying sediments is **cementation**. The porewaters in sediments are commonly supersaturated with respect to at least some minerals, and the porewaters precipitate those minerals onto the surfaces of the sedimentary particles. These newly formed minerals fill interparticle space and cement the grains together, forming a solid rock. The most commonly cementing minerals are quartz (common in sandstones) and calcite (common in both sandstones and limestones).

Compaction, cementation, and other changes in rocks in the subsurface are collectively called "diagenesis". These processes take place at temperatures of 20 to 200°C within the Earth. When rocks are subjected to greater temperatures, we typically begin to think of them as metamorphic rocks, which will be the topic of Chapter 6.



Photomicrograph of a calcite-cemented sandstone (specifically, a quartz arenite). Well-rounded quartz grains (Q) have euhedral quartz overgrowth cements beyond their original grain edges (arrows). Calcite (C) has filled remaining pore space. Field of view is 2.3 mm wide. Sample is from the Mesozoic of South America and donated by to the University of Georgia Department of Geology by Professor Gilles O. Allard.

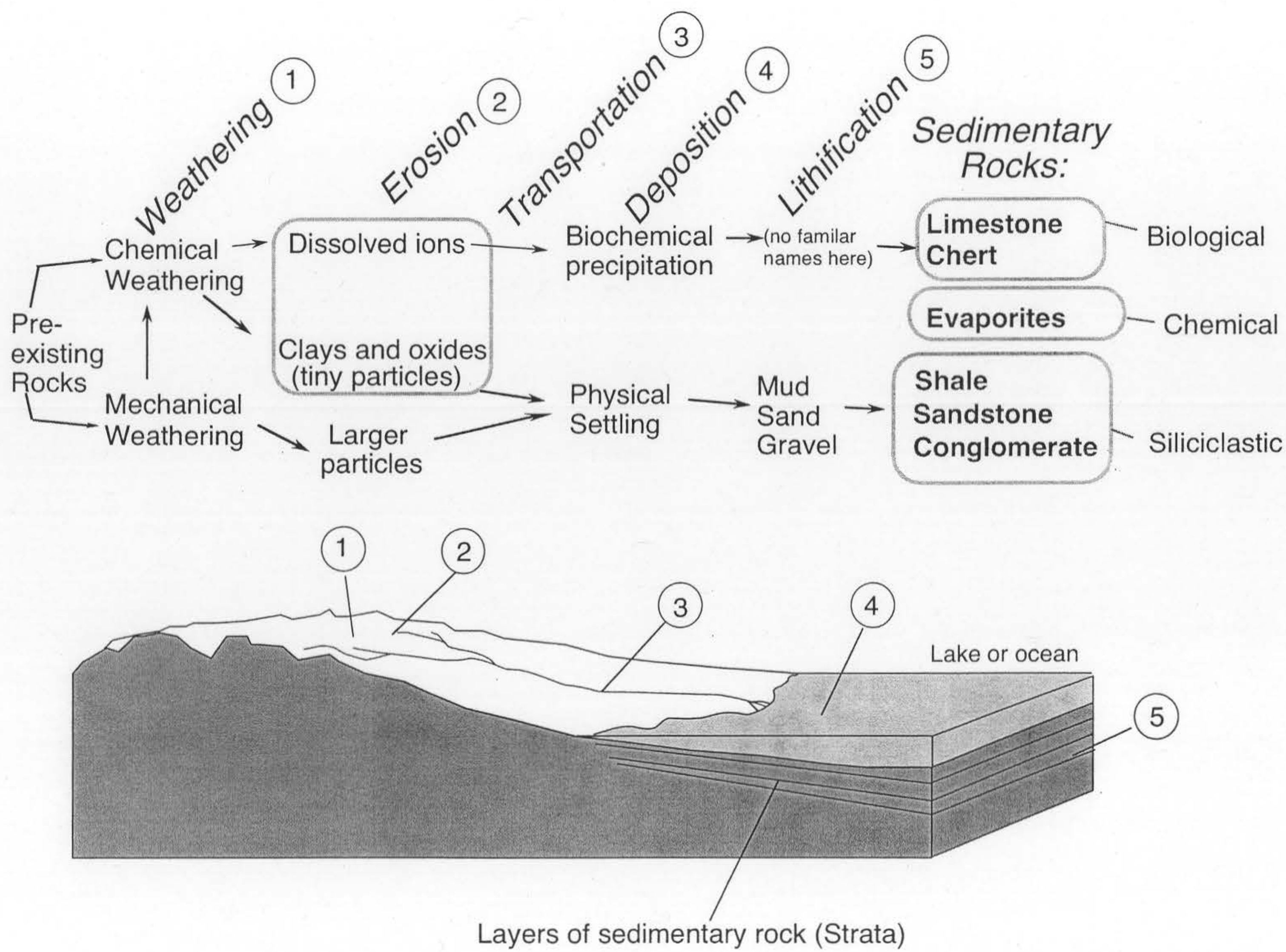


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Figure 4-1: A schematic illustration of how sediments originate from weathering of pre-existing rocks, and subsequent transportation of resulting particles and dissolved solids to sites of deposition.

Sedimentary Grain Sizes and Rock Names

φ[a]	Size(mm)	Grain[b]	Sediment	Rock
		Boulder		
-8	256			
-7	128	Cobble		
-6	64			Conglomerate
-5	32		Gravel	
				<Breccia>[c]
-4	16	Pebble		
-3	8			
-2	4			
		Granule[d]		
-1	2			
		VC Sand		
0	1			
		Coarse Sand		
1	$\frac{1}{2}$			Sandstone
		Med. Sand	Sand	
2	$\frac{1}{4}$			<Grit[d]>[c]
		Fine Sand		
3	$\frac{1}{8}$			
		VF Sand		
4	$\frac{1}{16}$			
5	$\frac{1}{32}$			
6	$\frac{1}{64}$	Silt	Silt	Siltstone
7	$\frac{1}{128}$			
				Mudstone
8	$\frac{1}{256}$		Mud	Shale[e]
				Clay Shale[e]
9	$\frac{1}{512}$	Clay[f]	Clay[f]	Claystone

Notes

[a] φ scale of Krumbein (Jo. Sed. Petrol., v.4, p. 65-77, 1934). $\phi = -\log_2(\mathrm{diam}_{(mm)})$

[b] Wentworth scale (Jo. Geol v.30, 377-392, 1922).

[c] "< >" used for rocks with angular clasts

[d] Term with increasingly limited modern usage

[e] Usually used for laminated rocks.

[f] "clay" is also a mineralogical term.

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Figure 4-2: Terms for grain sizes of sediments, and names of corresponding siliciclastic rocks.

Sediments and Corresponding Sedimentary Rocks

Sediment	*Sedimentary Rock*	*Where the sediment accumulates*
Gravel	Conglomerate	Alluvial fans, river channels, wave-swept coastlines
Sand	Sandstone	Desert dunes; river channels, shorelines, deltas, shallow seas
Mud	Shale	Lakes. river floodplains, tidal flats, distal deltas, deep sea
Shells and lime mud	Limestone	Warm shallow seas
$CaCO_3$ produced by marine plankton	Chalk	Deep sea
SiO_2 produced by marine plankton	Chert	Deep sea
Woody plant matter: Peat	Coal	Swamps
Salt	Rock salt	Lagoons or marginal seas in horse latitudes

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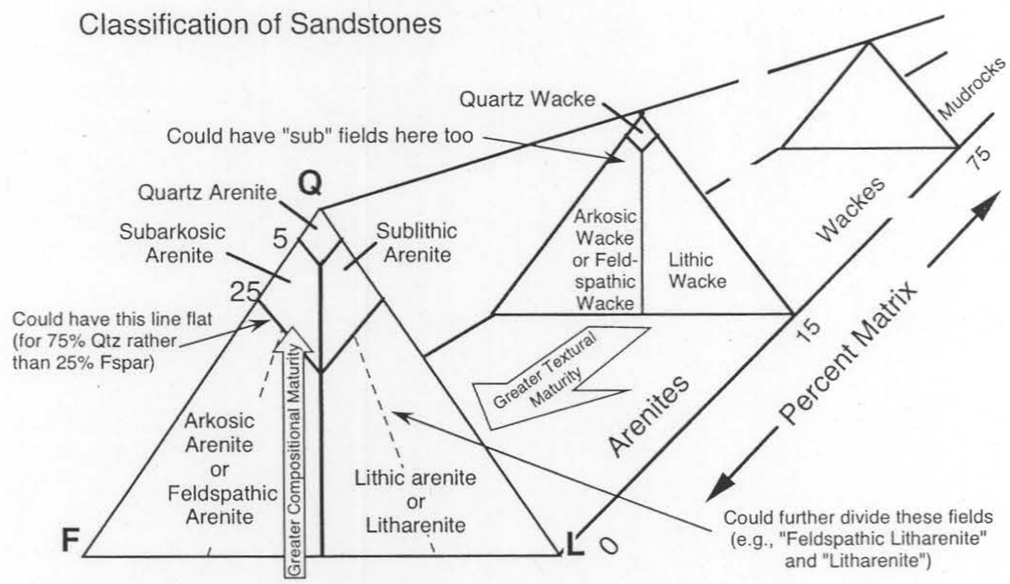
Figure 4-3: A generalized list of common sedimentary rocks and their depositional environments.



Q = Quartz grains (but not chert grains)

F = Feldspar grains (and commonly granitic rock fragments)

L = "Lithic" grains: polymineralic grains (i.e., rock fragments), and monomineralic grains other than those of quartz and feldspar

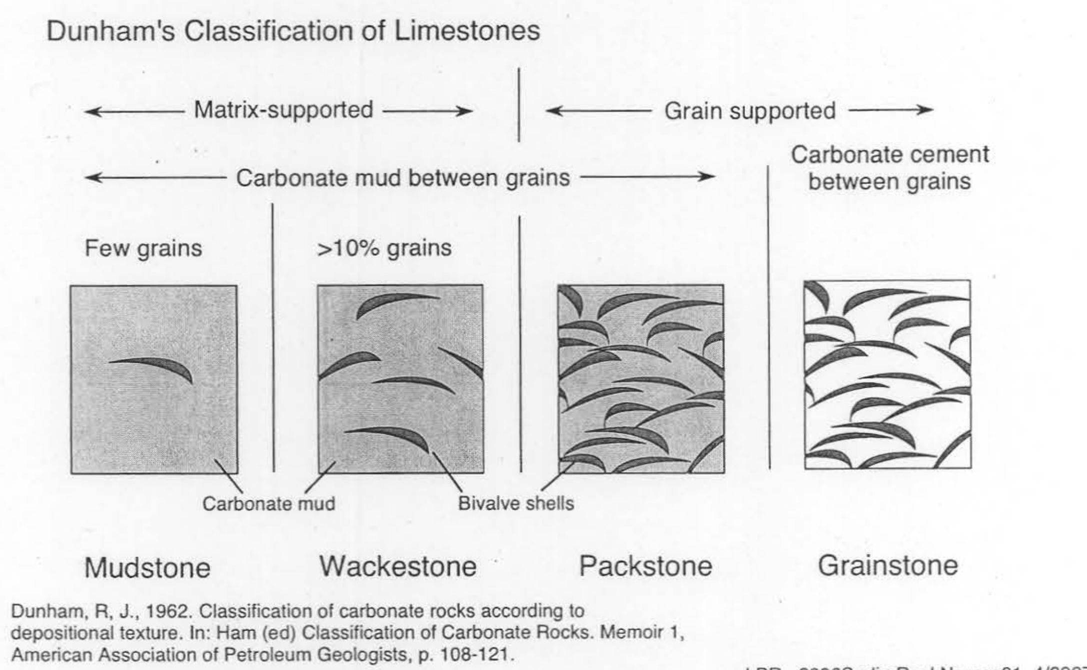


Figure 4-4: Advanced nomenclature for sandstones and limestones.

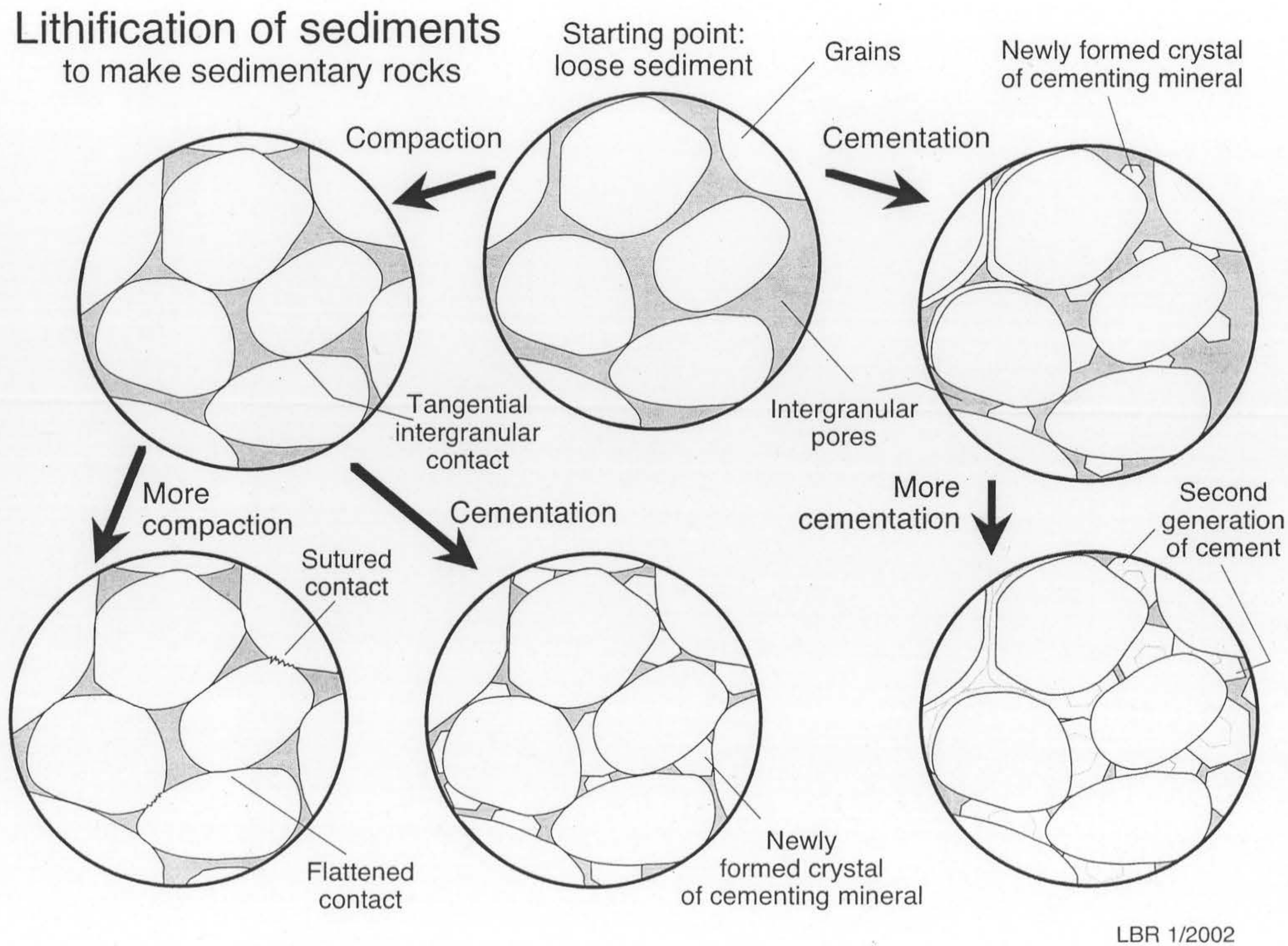


Figure 4-5: Processes in the lithification of sandstones and limestones.

Chapter 5: Coarse-grained Sedimentary Rocks, and Breccias

On relatively flat landscapes, weathering breaks down bedrock to generate sand-to-clay-sized material that is eroded by streams and rivers. As steep-sided landscapes, mountains are places where weathering has little time to break down rock before gravity carries that rock to lower elevations. As a result, geologists in mountains are commonly confronted with sediments and sedimentary rocks that consist of large clasts. This chapter therefore briefly examines coarse-grained rocks to provide a context not usually found in texts less focused on mountains and glaciation.

Defining characteristics

The clasts about which this chapter is concerned are technically those particles more than 2 mm in diameter, because smaller particles are defined as sand and a component of sandstones (See Fig. 4-2). In actual practice, we are commonly concerned with sediments and rocks with clasts that are at least a few centimeters across: fist-sized to refrigerator-sized, and sometimes even larger. The fine (<2 mm) particles between these clasts are called "matrix".[24]

One characteristic used to distinguish among these sediments is the roundness or angularity of the clasts – whether the clasts have smooth rounded surfaces or jagged sharp edges. The significance of this distinction is that rounding of detrital clasts requires abrasion, and thus requires considerable transport in a mode where clasts bang against each other. This abrasion may take place on the beds of streams and rivers, or perhaps in a slowing flowing mass of sediment and ice (a glacier). On the other hand, angularity requires that the clasts not be transported far.

A second characteristic used to distinguish among coarse-grained sediments is whether the clasts are clast-supported or matrix-supported. In a clast-supported sediment, clasts are in contact with each other and rest on each other. Thus, if we could remove all of the matrix, the clasts would not fall from their present positions. In a matrix-supported sediment, clasts are commonly not in contact with each other, and removal of all the matrix would cause the clasts to fall. The significance of this distinction is that flow of a non-viscous medium like water sorts grains by size sufficiently to ensure a clast-supported fabric. In contrast, flow of a viscous medium like ice or a mudflow does not sort the clasts by size and thus commonly deposits a matrix-supported sediment.

Descriptive vs. Genetic Classifications

Classification schemes can distinguish between clearly defined characteristics, like those above, to assign names to things. Such classification schemes are descriptive. On the other hand, classification schemes can assign names to things on the basis of the origin inferred for those things.[25] Descriptive classification schemes provide an objective name that should not change through time, but the name provides a reader or listener with no statement about the origin of the thing classified. Genetic classification schemes are more informative about the inferred origin of the thing classified, and are more useful if that inference was correct. If it was not, the name provides disinformation and must ultimately be changed. The treatment that follows distinguishes between names in these two kinds of systems because names of both sorts are commonly used to classify coarse-grained sediments and rocks, with occasional resultant problems.

Matrix-supported sediments and rocks

Diamictons are matrix-supported sediments with large clasts. Diamictites are lithified diamictons. The two most common origins of such sediments are as deposits by glaciers (i.e., glacial till) and as mudflows. Distinguishing between the two is often not straightfoward. Tills commonly consist of a range of particle sizes and so are very poorly sorted, and the clasts are commonly somewhat rounded. Tills of continental glaciers usually have clasts of many different rock types. Mudflows, on the other hand, may be more bimodally sorted because they consist of soil (a

[24] Matrix is material deposited as a sediment between the larger grains, either as the large grains are deposited or soon thereafter. Cement, on the other hand, is mineral material precipitated chemically by groundwater after, and commonly long after, deposition of the grains.

[25] Imagine classifying chocolate-chip cookies. A descriptive classification might distinguish between "hard cookies" and "soft cookies". A genetic classification might correspondingly distinguish between "dried-out cookies" and "water-soaked cookies".

fine-grained component) and chucks of more-or-less unweathered bedrock (the coarse component). The clasts of mudlows may be relatively angular, because most mudflows do not travel great distances. Despite these clues, distinguishing between the two can be difficult, and many ancient deposits have been identified as tillites (lithified tills) and thus as evidence of glaciation, when in fact less certainly was justified.

Conglomerates

Conglomerates are clast-supported rocks with large rounded clasts. They are the lithified equivalent of gravels, which most commonly accumulate in stream beds. Gravels also accumulate on wave-swept beaches like those of the New England coast and the coast of the northwestern United States.

Because gravels are deposited in localized settings like stream channels and beaches, conglomerates rarely form widespread or thick rock layers in the stratigraphic record. When they do form widespread or thick stratigraphic units, they typically are the result of erosion from an ancient mountain range that provided a major source for such sediment.

Breccias

Breccias are clast-supported rocks with large angular clasts. As sedimentary rocks, they are most commonly the lithified equivalents of talus or scree, jumbles of chunks of rock that have fallen from cliffs or mountainsides to make rocky slopes.[26] One thus commonly hears the expression "landslide breccia", a combination of genetic and descriptive terms.

Breccias can also originate in three ways that do not involve sedimentation, and thus some breccias are not sedimentary rocks. One origin is in the jumble of rock fragments that forms along a fault as motion of the two fault blocks grinds the rock along the fault. Such breccias can be recognized in part because the clasts commonly have undergone internal deformation visible in thin section and often in hand specimen. Such breccias may also have zoned cements of exotic minerals precipitated by waters moving along the fault. Such breccias may also be recognized because they are found along faults, but geologists more commonly use the opposite logic, using a breccia as evidence for the presence of an otherwise unapparent fault.

Breccias can also form where evaporites have undergone subsurface dissolution. Evaporites consist of relatively soluble minerals like gypsum and halite, and subsurface waters can dissolve layers of these minerals. The result is tabular caverns into which overlying layers of sedimentary rock collapse. This collapse generates a breccia of tabular clasts of the overlying strata, with the clasts roughly parallel to the rock layers above and below. The space between clasts is devoid of fine matrix and instead is commonly filled with cementing minerals precipitated by groundwater.

Finally, breccias can form as the result of impacts by meteorites or asteroids. The diagnostic evidence of such breccias is shocked quartz, which is quartz with microscopic deformation lamellae that only form with rapid application of intense pressure. Another identifying characteristic are shatter cones, conical fractured surfaces a few centimeters in size that likewise form under the sudden intense pressure of an impact.

Meteorite impacts are not common, and collapse breccias are no more common in alpine settings than anywhere else. However, fault breccias, landslide breccias, mudflow deposits, and glacial tills are all common in mountains. Application of the criteria above is thus commonly needed to understand the coarse-grained materials found by geologists in mountains.

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[26] The difference between talus and scree is essentially one of clast size, with scree for most people consisting of clasts less than ten centimeters in size and talus consisting of clasts more than 30 centimeters in size. From the standpoint of hazards to hikers, scree is the loose rock that slides under one's feet and sends one hurtling downslope, whereas talus is the jumble of rocks with interstices into which one can step and break a leg.

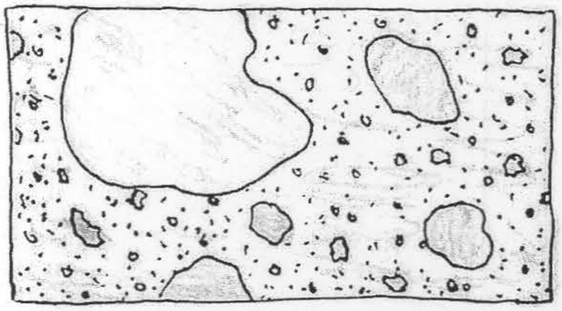
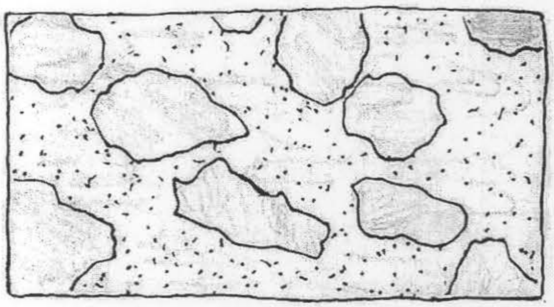
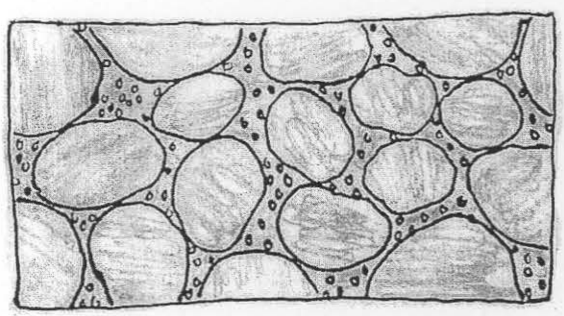
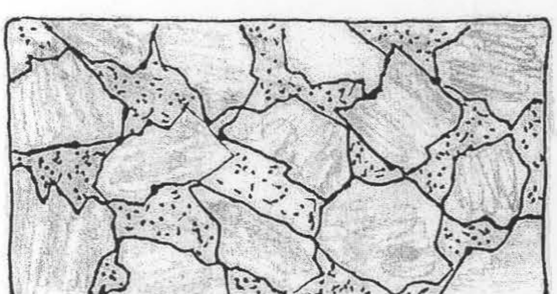
		Descriptive name		Genetic name	
		for unlithified sediment	for lithified sedimentary rock	for unlithified sediment	for lithified sedimentary rock
	Very poorly sorted, matrix-supported, variable clasts that are commonly somewhat rounded	**Diamicton**	**Diamictite**	**Till**	**Tillite**
	Bimodally sorted, matrix-supported, variable or not variable clasts that are commonly somewhat angular	**Diamicton**	**Diamictite**	**Mudflow**	**none**
	Bimodally sorted, clast-supported, rounded clasts	**Gravel**	**Conglomerate**	**none**	**none**
	Bimodally sorted, clast-supported, angular clasts	**none**	**Breccia** See also "Origins of Breccias" for non-sedimentary origins	**Talus** or **Scree**	**none**

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Figure 5-1: Terminology, or lack thereof, for coarse-grained sediments and rocks.

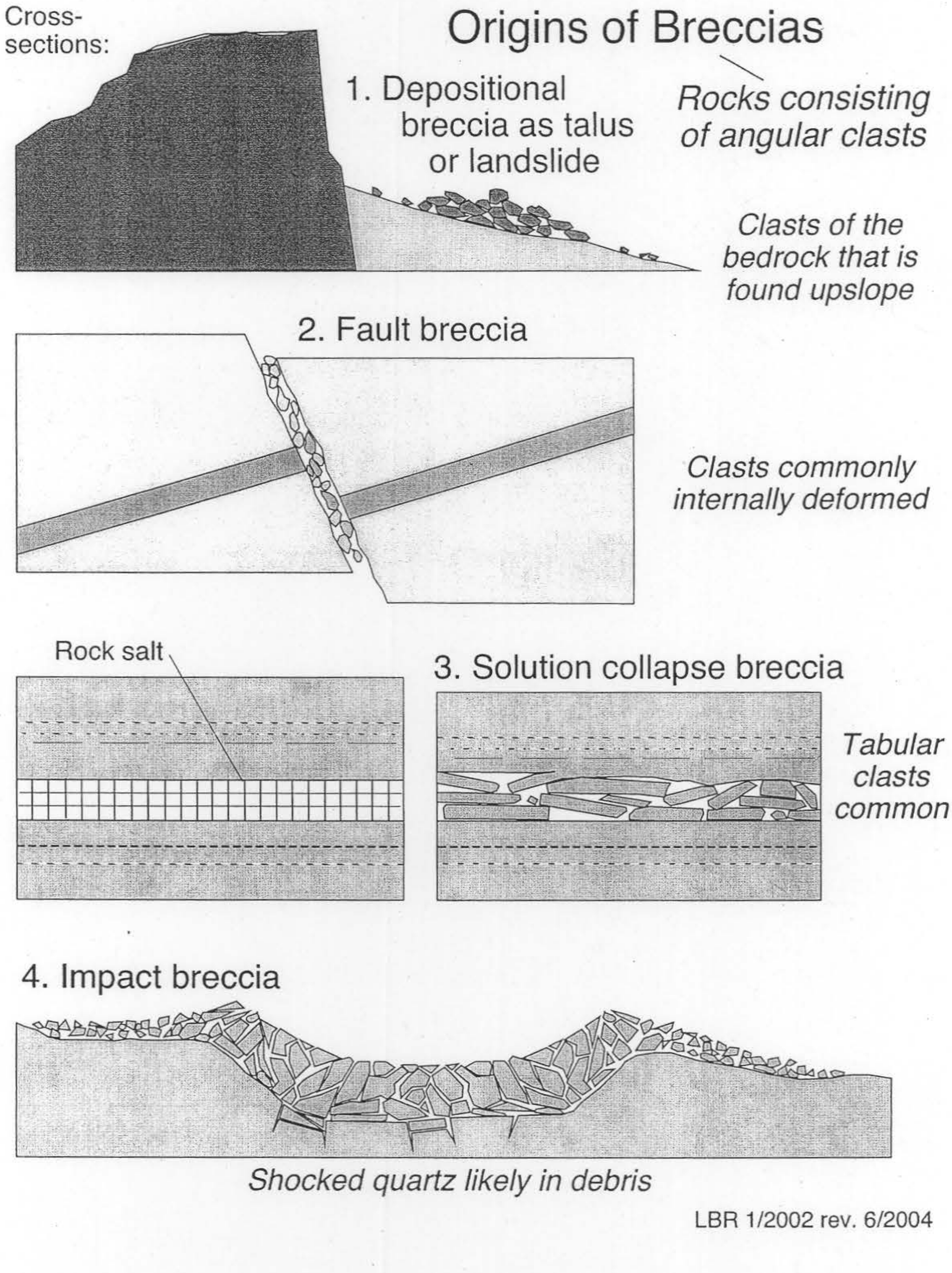

Figure 5-2: Four origins of breccias.

CHAPTER 6: METAMORPHIC ROCKS

Metamorphism, in the geological sense of the word, is the set of changes that rocks undergo when subjected to high temperature and/or pressure. Those changes include breakdown of some minerals, growth of new minerals from the chemical constituents of those breaking down, growth of crystals with preferred orientation (a concept discussed further below), and growth of progressively larger crystals. These processes produce metamorphic rocks, which are defined as those rocks that have undergone a change of mineralogy and fabric from some pre-existing rock.[27] The extent of such change is called metamorphic grade.

Metamorphism typically takes place in one of two contexts. One is where bodies of magma give off so much heat that the surrounding rocks are altered by temperature, but not particularly by pressure. This process is called *contact metamorphism* because it takes place at the contact, or boundary, between igneous rocks and surrounding rocks. Because the magma body gives off only a finite amount of heat, the metamorphic event is relatively short-lived. The other context is deep within the continental crust, where sedimentary rocks, or igneous rocks that formed at shallow depths, are subjected to great temperature *and* pressure. This process is called *regional metamorphism* because it produces great volumes of metamorphic rock that may then be exposed across broad regions, in contrast to the localized distribution of rocks altered by contact metamorphism. Because Earth's interior is consistently hot, regional metamorphism is a long-term event. Thus one might define these kinds of metamorphism not by their context but instead say that contact metamorphism is localized high-temperature low-pressure geologically brief metamorphism, and that regional metamorphism is widespread high-temperature high-pressure long-term metamorphism.

Metamorphic rocks are classified first by texture, on the basis of whether they have a preferred orientation or not. This proves to be largely a compositional distinction as well, but further subdivision is made among the non-foliated rocks in terms of their composition.

Metamorphic rocks with preferred orientations

If the platy or linear crystals of a rock are not arranged randomly but instead are parallel to each other, the rock is said to have a preferred orientation (Fig. 6-1). Preferred orientation of platy crystals is called foliation (from the parallel packing of leaves), and preferred orientation of linear minerals is a lineation. Preferred orientations form when metamorphism takes place under directed (non-isostatic) pressure, so that certainly crystal orientations are favored and some are disfavored. Because the only common platy minerals are layer-silicates and the only common linear minerals are chain-silicates, foliated and lineated metamorphic rocks are almost inevitably silicate rocks.

The foliated rocks are in general categorized by the size of their crystals. **Slates** are foliated rocks with such fine crystals that the crystals can not be seen with the naked eye. Slates break smoothly, which is why slate blackboards were for centuries the surface on which people wrote with chalk. The crystals are sufficiently small that smooth surfaces of slate do not reflect light. **Phyllites** are foliated rocks with crystals large enough to be seen, at least with a handlens if not with close examination by the naked eye, and the smooth surface of a phyllite will reflect light to some extent.[28] **Schists** are foliated rocks with crystals large enough to be seen easily with the naked eye, and those large crystals commonly reflect light well. Schists may contain large crystals of non-platy minerals like garnets, so that the foliation wraps around those crystals like water in a stream swirling past a boulder. Finally,

[27] This definition would seemingly include limestones, which commonly undergo a change from aragonite to calcite (the two common polymorphs of $CaCO_3$) and would include coal, which has undergone great change in fabric from peat. However, most geologists consider limestones and coal to be sedimentary rocks, and they only include as metamorphic rocks those rocks altered at temperatures above 200 to 300 °C (Fig. 6-2).

[28] The term"phyllite" is often neglected in introductory presentations of names for metamorphic rocks, but it is needed for anyone studying the geology of the Alps, and especially the Tirolean Alps, because phyllite is abundant in the Alps southeast of Innsbruck.

gneisses are foliated rocks with large crystals, and thus like schists, but with distinct compositional layers. For example, many gneisses consist of foliated layers of micas and non-foliated layers of feldspar and quartz.

The only common lineated rocks are **amphibolites**, which consist of elongate crystals of amphibole without or without other minerals.

Metamorphic rocks without preferred orientations

Equant crystals cannot form lineations or foliations, so rocks consisting of minerals that generally take equant forms have no preferred orientation. For example, quartz generally forms equant crystals, so **quartzites**, which are metamorphic rocks consisting largely of quartz, have no preferred orientation. Quartzites are commonly the result of metamorphism of sandstone. **Marble**, on the other hand, consists of calcite, another equant mineral. Marbles are rocks resulting from metamorphism of limestone, and their lack of a preferred orientation and the relatively low hardness of calcite are the reason marble has been so popular as a stone for sculpting and building. The third non-foliated metamorphic rock is **anthracite coal**, the result of metamorphism of normal (bituminous) coal. Anthracite is typically shinier and harder than bituminous coal, and it burns to give greater heat.

There is one exception to the generalization that silicate metamorphic rocks other than quartzites have preferred orientations. If silicate rocks undergo contact metamorphism, where there is little directed pressure to induce a preferred orientation, the result is a fine-grained non-foliated rock called a "**hornfels**". In addition, contact metamorphism of limestone can produce rocks rich in both silicate and carbonate minerals, because magmas commonly expel silicate-rich waters into the surrounding limestones. These rocks are called "**skarns**" and can include large crystals of various exotic minerals, including but not limited to garnets.[29] Skarns are commonly of interest to the mining industry because they may host economically significant minerals.

The significance of metamorphic rocks

Metamorphism, other than contact metamorphism, takes place at considerable depth in the Earth. Exposure of metamorphic rocks at the Earth surface thus requires removal of the rock that overlay those metamorphic rocks. Exposure of metamorphic rocks in mountain ranges thus implies considerable uplift and/or isostatic rebound[30], and thus considerable orogenic activity. This is especially true where higher-grade metamorphic rocks are exposed in the heights of mountain ranges: rocks metamorphosed many kilometers deep in the Earth are now exposed kilometers above sea level.

When observed in greater detail, metamorphic rocks provide information about the conditions under which metamorphism took place and thus about the history of the crust in which the rocks are found. Many minerals are stable only over a limited range of temperature and pressure, and they break down to form different minerals with increasing temperature and pressure. Laboratory research has determined the transforming temperatures and pressures of many mineral assemblages, so that metamorphic petrologists can infer temperatures and pressures from the minerals found in the rocks of a particular locality (Fig. 6-3).

This concept can be applied to specific minerals or to entire rock types. At the scale of specific minerals, three minerals with the composition $Al_2Si_2O_5$ are diagnostic of different temperature-pressure regimes, with kyanite forming at highest pressures (Fig. 6-3). At the whole-rock scale, metamorphic petrologists recognize different facies of rock typical of general ranges of temperature and pressure. The typical progression of these facies with increasing pressure and temperature is from zeolite facies (characterized by certain zeolite minerals) to greenschist facies (characterized by the green mineral chlorite) to amphibolite facies (charac-

[29] The "silicate-rich waters" expelled from magmas also commonly contain elements that are not readily included in the common minerals of igneous rocks (the "incompatible elements" of igneous petrology). Thus the "exotic minerals" of skarns may consist of these more rare elements, which accounts for the economic significance of some skarns.

[30] Isostatic rebound is upward movement of rock in response to removal of an overlying mass, in this case by erosion. Imagine an ice cube floating in water. As the top of the ice cube melts, the underlying ice rises. The same applies to rocks at depth in the crust as overlying rock is eroded.

terized by amphiboles) to granulite grade (characterized by coarse feldspars). The upper limit of the granulite facies is melting to generate magma. Some granulite-facies rocks called migmatites record heating to this boundary, in that they consist of intensely metamorphosed layers or zones interspersed with rock that appears to have undergone melting to resolidify as igneous rock.

Some closing thoughts about rocks

Having discussed igneous, sedimentary, and metamorphic rocks, we can now look for some larger perspectives on them.

One concept commonly introduced in introductory geology texts is the rock cycle (Fig. 6-4). The rock cycle is a useful concept in reminding one that mineral matter can be transformed from one kind of rock to another. Thus the atoms of modern rocks have existed previously in other kinds of rocks. However, the word "cycle" is misleading, in that a diagram of the possible transformations of rocks looks more like spaghetti than like a Cheerio (Fig. 6-4). Few atoms progress neatly from igneous to sedimentary to metamorphic to igneous rocks.

The relative abundance of these rocks is a matter of perspective. Across the continents, sedimentary rocks are the bedrock of 75% of the land surface, and igneous and metamorphic rocks are the bedrock of the remaining 25%. On the other hand, the continental crust by volume is overwhelmingly igneous and metamorphic rocks, and the sedimentary rocks necessarily emerge as a veneer at the top.[31] The same is true of the oceans: sediments cover almost all of the seafloor, but they are a thin veneer over the basalts and gabbroes that make up most of the thickness, and thus most of the volume, of the oceanic crust. Someone interested in the land on which we walk and the seafloor that we can sample sees the abundance of sedimentary rocks; someone interested in the entire crust envisions the abundance of igneous and metamorphic rocks (Fig. 6-5).

A third point to consider about our discussion of rocks is that, like all users of language, we have tried to erect categories labeled with words. However, intermediate forms exist at the boundaries between all these groups (Fig. 6-5). For example, slaty shales lie at the boundary between sedimentary and metamorphic rocks, and migmatites at the boundary between metamorphic and igneous rocks. These rocks constitute no great mystery, but they are a reminder that our assignment of words to things in the business of classification does not represent the entire spectrum of reality.

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[31] "Veneer" is a relative term: oil wells have penetrated sedimentary rocks at depths as great as ~30,000 feet (~10 km) in sedimentary basins, a great depth by human standards but only a small proportion of the thickness of the continetnal crust.

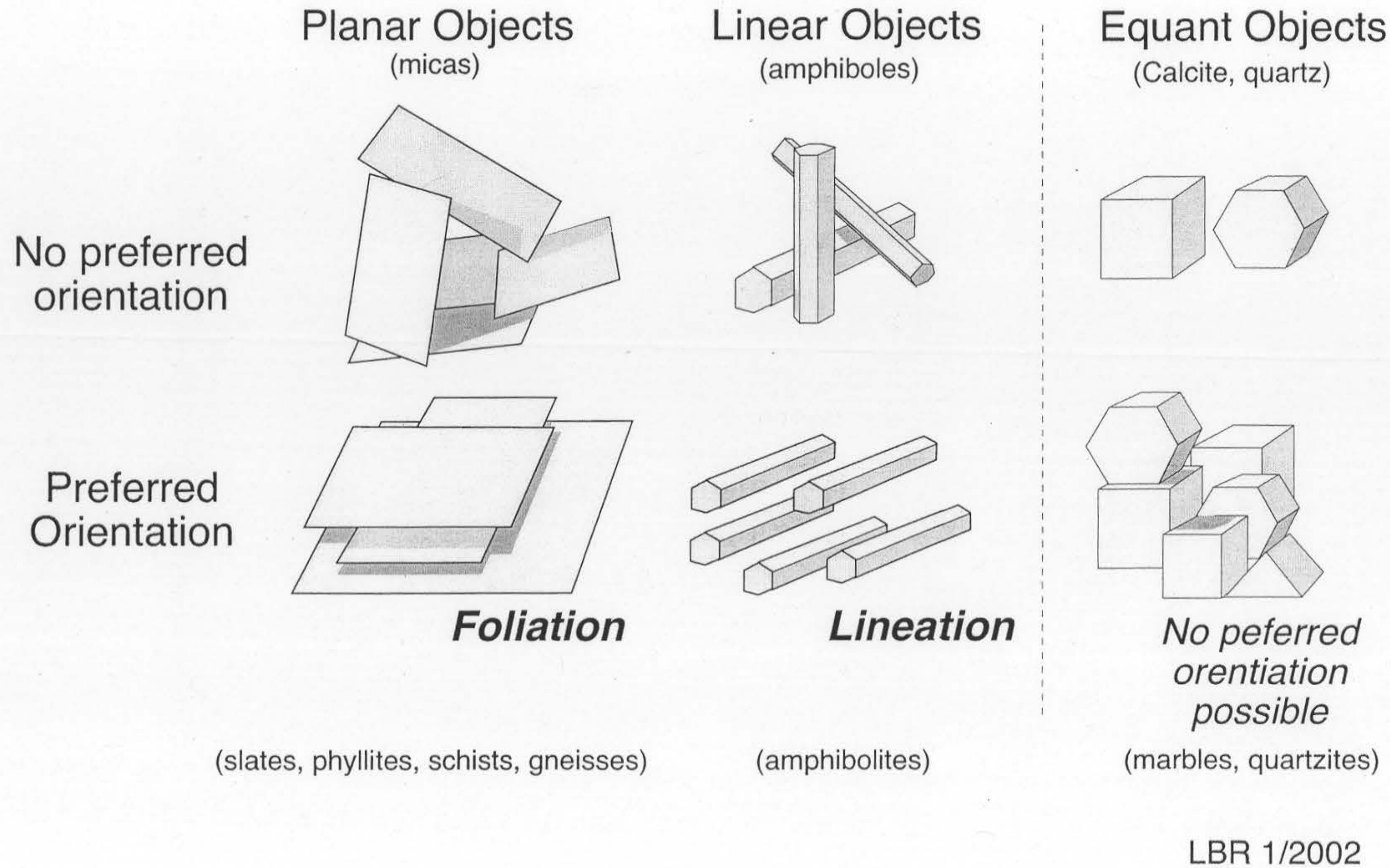


Figure 6-1: Preferred orientations of objects, with application to minerals in metamorphic rocks.

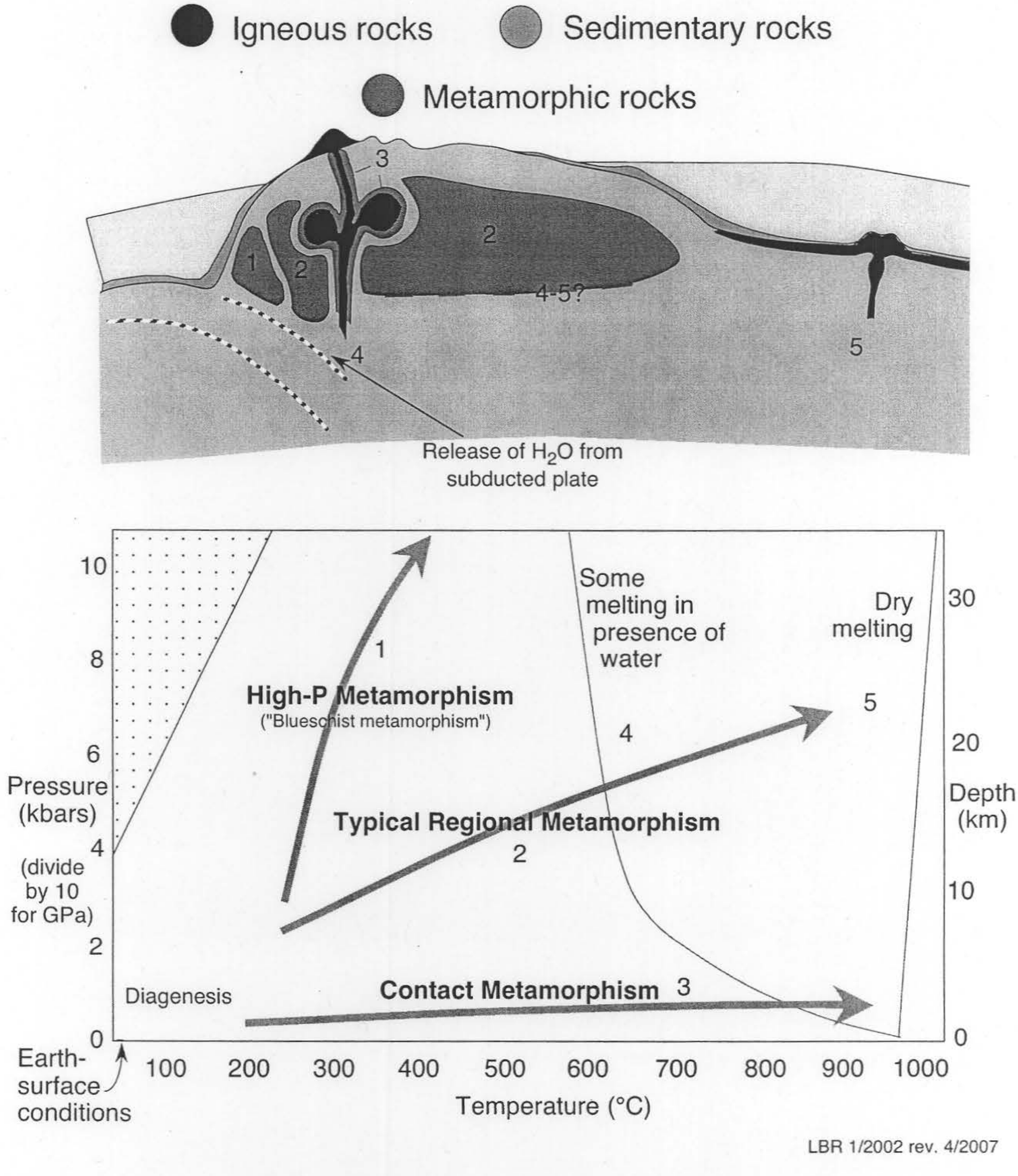
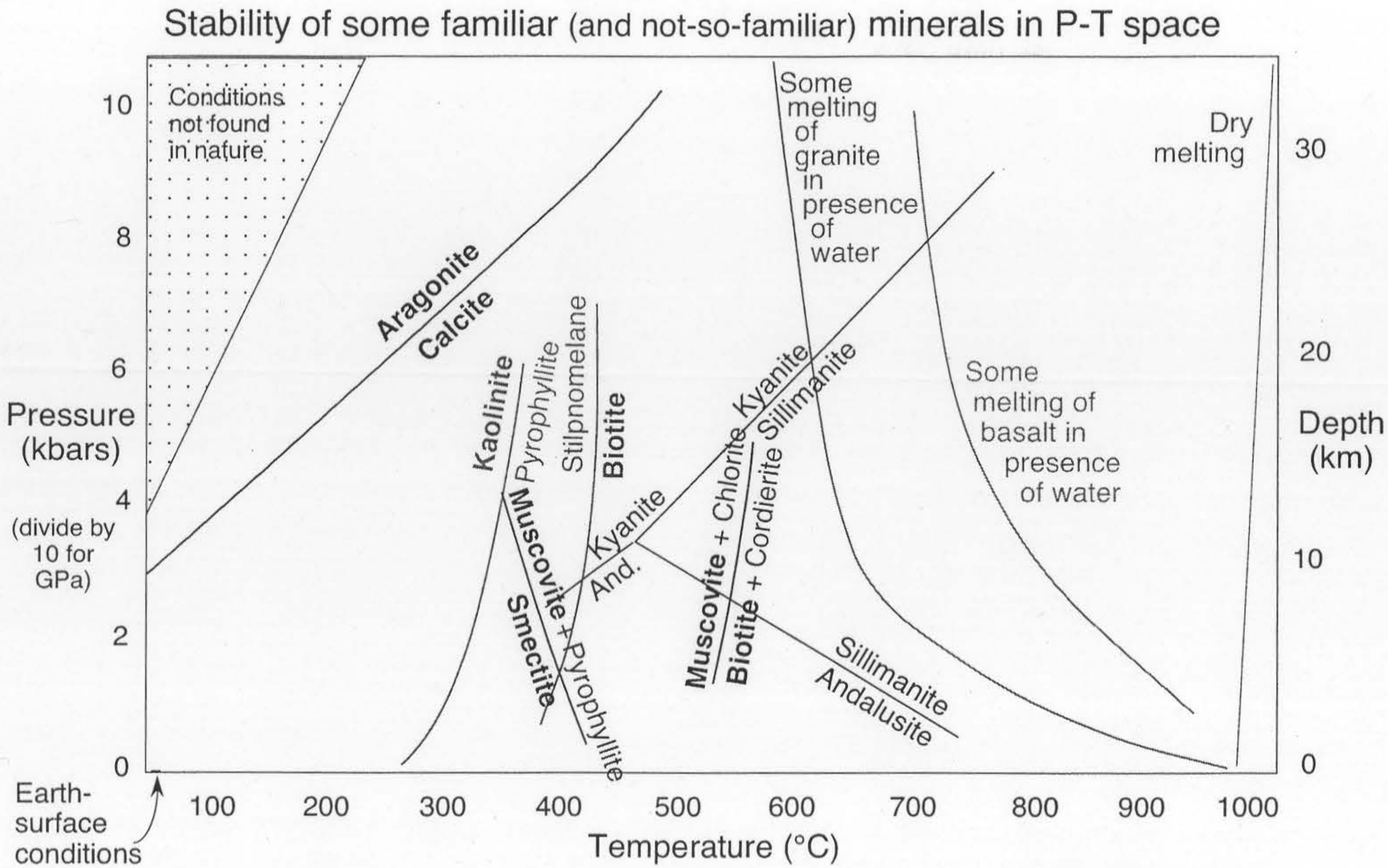


Figure 6-2: Relationship of locations to metamorphic rocks to their pathways through P-T space.



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Curves are approximated from Figure 7-11 and 7-12 of Hyndman's (1972) *Petrology of Igneous and Metamorphic Rocks*

Figure 6-3: Stability boundaries for some diagenetic and metamorphic minerals.

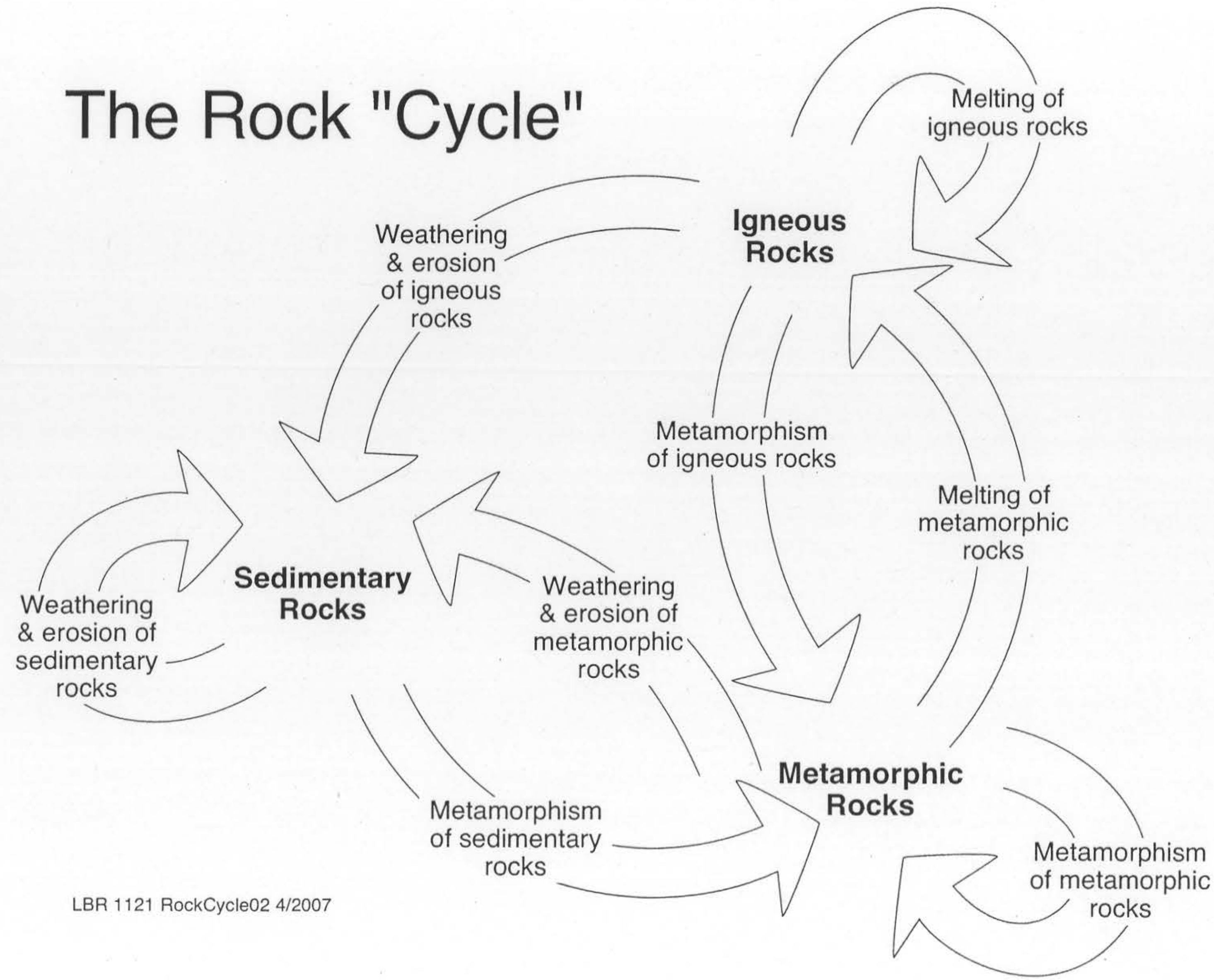
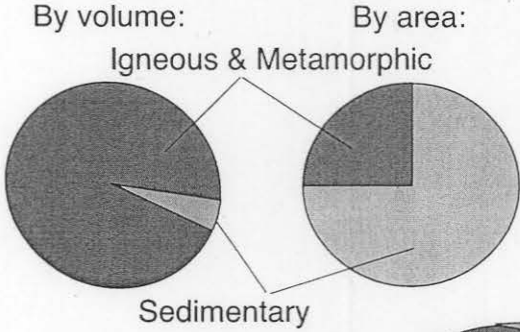
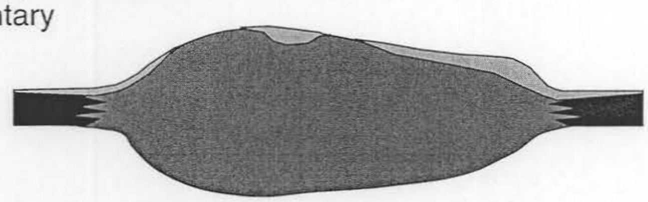


Figure 6-4: The rock cycle. The three central arrows making a circle are the most simple presentation of the rock "cycle"; the entire diagram more accurately represents reality.

- Proportions of rock types in the continental crust:





- Relative abundance of sedimentary rocks:

	Measured	Calculated
Shales:	47%	77%
Sandstones	31%	14%
Limestones	22%	9%

- The continuity of the three kinds of rocks:

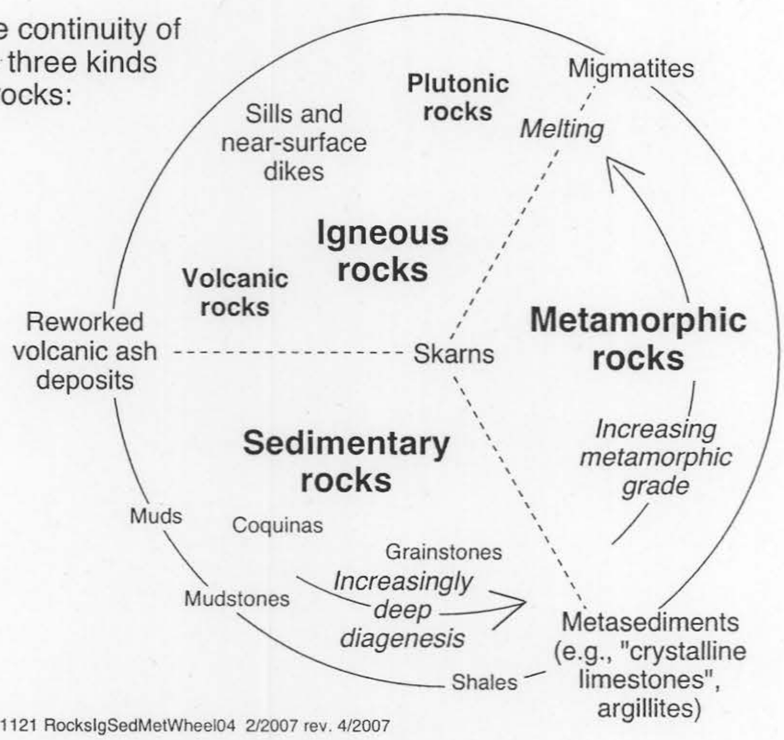


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Figure 6-5: Relative abundances of the various rock types, and their intergradations.

Chapter 7: Weathering, Erosion, and Mass Wasting

The previous three or four chapters have focussed on rocks and their formation, and thus on processes that make the solid substrate under our feet and of our mountains. This chapter instead focuses on some of the processes that break down rock and, at the larger scale, tear down our mountains. That's not a bad thing, because those processes generate the soil on which our food depends and from which mountain flowers grow, and they account for much of the sculpting that makes mountains so spectacular.

Weathering

From a destructive standpoint, weathering is the set of processes that break apart solid rock to produce small particles, some of the same mineralogy as their precursor rock and some of different mineralogy. From a more constructive or progressive standpoint, weathering is the set of processes that brings rocks formed at depth in the Earth to equilibrium with the Earth-surface environment. Those rocks, after all, formed in environments characterized by high temperatures, high pressures, low abundances of O_2 and H_20, neutral to alkaline concentrated groundwaters, and the absence of life. However, at the Earth surface those rocks must equilibrate with an environment characterized by low temperatures, atmospheric pressure, abundant O_2 and H_2O, acidic dilute solutions, and an environment greatly influenced by life and the presence of organic carbon. From that perspective, it's hardly surprising that rocks must undergo great changes when exposed at Earth's surface. The processes that change them are divided into physical weathering and chemical weathering

Physical weathering consists of at least three processes. One is **exfoliation**, the development of horizontal planar fractures. Rocks in the Earth are under pressure both horizontally and vertically. However, when a rock is exposed at Earth's surface, pressure is released upwards, so not surprisingly the upper most rock is prone to spring upwards. The result is the development of horizontal fractures a few feet to tens of feet below the land surface. Such fractures provide some of the first avenues for further weathering processes.

A second major process in physical weathering is "**freeze-thaw**", the repeated freezing and thawing of water in tiny cracks in the rocks. Most accounts of this process assume that the growth of ice in cracks in rock, and the expansion of the H_2O as it freezes, forces open pre-existing cracks further and so breaks rock apart.[32] The result is breakage of the rock into smaller pieces with more surface area, which promotes other weathering processes too.

The third major process in physical weathering is **root wedging**. As plant roots grow into pre-existing cracks in rock, they exert enormous pressures that open those cracks further. Plant roots can thus break rock apart, further contributing to the breakage of previously solid rock into progressively smaller pieces.

Chemical weathering is the set of chemical processes that transform pre-existing rock or rock particles into new minerals or into dissolved solids. The simplest of these processes is **dissolution**, wherein minerals break down entirely into dissolved solutes. This happens in the dissolution of halite (NaCl) to give dissolved Na^+ and Cl^-, and likewise the dissolution of gypsum, ($CaSO_4 \bullet 2H_2O$) and anhydrite ($CaSO_4$) to yield dissolved Ca^{2+} and SO_4^{2-} in solution that are then carried away by groundwater and riverwater to the oceans (Fig. 7-1).

A more extensive and important set of chemical reactions is called "**hydrolysis**", for the reaction of minerals with H_2O, but it might be better called "acidization". That's because all these processes involve the reaction of CO_2 in soil gas with H_2O to form H_2CO_3, or carbonic acid (Fig. 7-1). Carbonic acid is not a strong acid, but given years to millennia across which to work and constant resupply of CO_2 from plants, it attacks both carbonate minerals (the constituents of limestone and marble) and virtually all silicate

[32] A new view holds that the expansion of H_2O from water to ice is not the key the breakage of rock in the freeze-thaw process, but instead that ice crystals and surrounding water molecules propagate along cracks and force them open. This process has been observed to break open soils even with non-H_2O liquids for which the crystallizing solid takes up less volume than the liquid. See Hallet, B, 2006, Why do freezing rocks break?: *Science*, v. 314, p. 1092 - 1093, and Murton, J.B., et al., 2006, Bedrock fracture by ice segregation in cold regions: *Science*, v. 314, p. 1127-1129.

minerals other than quartz. When the various silicate minerals dissolve, many of their cations (e.g., Mg^{2+}, Ca^{2+}, Na^+, and K^+) go into solution, and much Si^{4+} goes into solution as H_4SiO_4 or, more accurately, $Si(OH)_4$.[33] However, the Al^{3+} ion is so insoluble that it instead forms some solid phase. In wet climates, all other cations are leached away, so the Al^{3+}-bearing soil mineral is commonly gibbsite ($Al(OH)_3$. In more moderate climates, some Si^{4+} remains so that the Al^{3+}-bearing new mineral is kaolinite ($Al_2Si_2O_5(OH)_4$). In dry climates, both Si^{4+} and other cations like Mg^{2+} and Ca^{2+} remain in the soil with Al^{3+} to form a group of minerals called smectites.

A third set of chemical reactions involves **oxidation** by atmospheric O_2. In the weathering of silicate minerals, the most common oxidation reaction is the conversion of the Fe^{2+} of mafic minerals to Fe^{3+}. Fe^{3+} is like Al^{3+} in being very insoluble, and it commonly stays in soils as an iron oxide or hydroxide. Oxidation of iron also commonly causes the physical breakage of rocks or their constituent minerals, because inclusion of more oxygen atoms creates a greater volume that ruptures the mineral or rock. Abundant Fe^{3+} oxide in soils gives them a red color, as in the soils of the Piedmont of the southeastern United States.

Another ion for which oxidation is very important is the S^{2-} of sulfide minerals, which oxidizes to the S^{6+} of SO_4^{2-} (sulfate). Oxidation of sulfide minerals inevitably produces sulfuric acid (H_2SO_4), and the oxidation of sulfides in mines and mine tailings can produce intensely acidic groundwater and stream water called "acid mine drainage".[34]

Soil is the accumulated layer of clay minerals, organic matter, and (in at least some cases) other minerals found at Earth's surface. Weathering and plant growth thus combine to produce soil.

[33] The two chemical formulae represent exactly the same solute, but they emphasize different things. $Si(OH)_4$ emphasizes the physical reality that the Si^{4+} ion is bonded to the four oxygen atoms of four OH^- groups. H_4SiO_4 emphasizes the potential behavior of this entity as an acid, called silicic acid, because it can release a H^+ ion at high pH.

[34] Chemistry students will remember that acidity is expressed with the pH scale, which is said to range from 0 (very acidic) to 14 (very basic). However, acid mine drainage can be so acidic as to have values of pH of –2 and –3 (as can industrially purified acids).

Erosion

Erosion is the removal of soil, sediment, or small pieces of rock by flowing water or air.[35] From a hypothetical smooth and only slightly inclined land surface, erosion can produce small rills that develop into gullies, which in turn can develop into canyons. Canyons like the Grand Canyon and the Black Canyon of the Gunnison are evidence of the extent to which erosion can cut down through solid rock. However, even more powerful evidence of erosion is seen in landscapes like that of Monument Valley, where only a few pillars of rock remain but are evidence of the vast volume of rock that has been removed by erosion.

Erosion most effectively removes sand-sized to silt-sized soil and sediment. Larger particles have sufficient weight to make their movement unlikely. Masses of clays, on the other hand, have sufficient inter-particle cohesion to inhibit removal of individual particles.[36]

Erosion is most effective in areas of high rainfall, because more water moves across the landscape; on steep slopes, where water runs off at greater velocity; and in regions with little vegetation, because the absence of roots and plant cover lets soil and rock fragments slip away more easily. Mountainous regions are thus prone to erosion because of their steep slopes and, in many cases, because mountains force updrafts that induce rainfall onto those slopes.

Degree of erosion also varies with the kind of rock eroded. Almost all igneous rocks, most metamorphic rocks, and sandstones are relatively resistant to erosion and thus can form cliffs in landscapes. Shales, on the other hand, are easily eroded and thus commonly form slopes and valleys.[37] Limestones can be cliff-forming strata in dry or cold regions, but in warm wet climates chemical weathering causes them to be eroded too quickly to survive as topographic prominences.

[35] There is, in addition, erosion by flowing glacial ice, a topic to which we will turn in later chapters about glaciation.

[36] As evidence of the cohesion of clay, consider a ball of potting clay or sculpting clay, and then consider the unlikelihood of a ball of sand.

[37] Preferential erosion of shale is how shale can be the most abundant kind of sedimentary rock, which is the most abundant kind of rock at Earth's surface, but to be observed so little.

Mass Wasting

Mass wasting is the movement of soil or rock downslope as the result of gravity. It might be better termed "gravity-driven movement". It may in some cases be promoted by the presence of some water, but it does not depend on flowing water, which separates it from erosion. It is also significantly different from erosion in that, whereas erosion only moves relatively small particles up to at most the size of pebbles or small boulders, mass wasting can move individual contiguous bodies of rock or soil that are kilometers in size.

"**Landslide**" is the most common term for the results of mass wasting, but geologists aspire to distinguish between different styles of mass wasting on the basis of two criteria. One is whether the material that moves is all rock, or if it is mostly or entirely soil or sediment. The other criterion is whether the moving mass remains intact with no internal movement, or it undergoes significant deformation and parts of the mass move relative to each other.

With those two criteria, one can attempt to divide the results of mass wasting into at least four categories (Fig. 7-2). For example, if a mass of rock moves downslope as one block and thus with no internal rearrangement, the result is called a "**glide**". If, on the other hand, a large amount of rock moves in individual pieces, it generates a "**rockfall**" (for example, at the base of a cliff) or perhaps a "**rockslide**" (if the chunks of rock move significantly horizontally).

When fine-grained material (i.e., soil or sediment) moves with little internal deformation, the result is a "**slump**", and when it moves with considerable internal deformation the result is a "**flow**". These two categories clearly intergrade. Both commonly produce a scarp (a small cliff) at their upper boundary. Their lower ends are commonly characterized by rumpled hummocky landscape, particularly in the case of flows, and hummocky terrane is a tell-tale clue of slumps and flows. Scarps are also good clues. A slump may be sufficiently well defined that one can envision its base in its upper reaches (near the scarp) as a very small and shallow normal fault. At the other end, one can envision its base in its lower reaches as a small and shallow thrust fault, or a series of small thrust faults.

Mass wasting is a relatively common process, and careful examination of almost any sloping landscape provides evidence of mass wasting. However, several factors can increase the likelihood of mass wasting. **Oversteepening** of slopes, either by fluvial erosion or by human activities like road-building, can release material to move downslope. Increased **water content** of soils and sediments, either from increased rainfall or a rise of the water table, can let particles slide past each other and promote mass movement. **Loading** on a landscape, either by other landslides or by human placement of masses, can increase the mass to a failing point and promote mass wasting. **Planes of weakness**, such as layers of mud or shales that dip into a valley, commonly fail and thus allow glides and slides. Finally, a **triggering impact**, commonly an earthquake but conceivably an anthropogenic explosion, can set off mass wasting.

Three examples illustrate these factors. The first is generic, rather than specific, and comes from the United States. In California, highway construction or clearing for development of real estate commonly involves cutting away slopes to make flat land surfaces, with resulting steep roadcuts or scarps at the bases of the remaining natural slopes. These slopes remain stable during dry years, but during occasional wet years, commonly El Niño years, pore pressure of soil water increases and lets slopes fail. The results are landslides, most commonly flows, that sweep down to block roads and cover homes and businesses.

The second example comes from Peru. On May 31, 1970, an offshore earthquake measuring 7.9 on the Richter Scale triggered sliding on the north side of Huascarán, a 6768-meter peak of granite in the western Andes. A mass of rock and glacial ice began the slide down the mountain, and it accumulated glacial deposits from the mountainside on the way down. The resulting mass is estimated to have been about a kilometer wide and 1.5 kilometers long, to have contained 61 million cubic meters of water, ice, and rock, and to have moved at more than 100 miles an hour.[38] It buried the town of Yungay, 11 miles away, and other towns as well. Yungay's entire population of 20,000 was buried and died in the landslide. Yungay's location in a valley below the mountain caused its destruction, but any concentration of human population in the area would have been in a valley below one of the region's mountains.

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[38] Details from Wikipedia.

The third example comes from the Dolomites, in the Alps in modern northern Italy. There, several days of heavy rain set in motion a large landslide into a reservoir behind the recently constructed Vaiont Dam. The landslide, on the night of October 9, 1963, slid into the reservoir, and it went up the opposite side of the valley sufficiently to cover one village. However, it also displaced about 50 million cubic meters of reservoir water that swept over the dam and moved down-valley as a ~220-meter-high wall of water. That water swept away several villages and killed about 1500 people there, contributing to a total human toll of about 2000. The dam itself was not significantly damaged and was a structural or engineering success. However, failure to take seriously the geological issues - dipping clay layers and the effect of water (both reservoir water and rain water) - allowed the building of a reservoir that was almost certain to generate a disaster.

The talus of a rockfall over which the Zirbenweg hiking trail passes
east of the Patscherkofel on the south side of the Inn Valley above Innsbruck and Hall-in-Tirol.

Mineral	*Weathering Reaction*	
Dissolution reactions		
Halite	$\underset{\text{Halite}}{NaCl} \text{ --> } Na^+_{(aq)} + Cl^-_{(aq)}$	
Gypsum	$\underset{\text{Gypsum}}{CaSO_4{\cdot}2H_2O_{(s)}} \text{ --> } Ca^{2+}_{(aq)} + SO_4^{2-}{}_{(aq)} + 2H_2O$	
Hydrolysis (acidization) reactions (all begin with $CO_2 + H_2O \text{ --> } H_2CO_3$)		*HCO_3^-/H_4SiO_4 in solution*
Mafic silicates: Olivine	$\underset{\text{Carbonic acid}}{4H_2CO_3} + \underset{\text{Olivine}}{MgFeSiO_4} \text{ --> } \underset{\text{Dissolved magnesium \& iron}}{Mg^{2+}_{(aq)} + Fe^{2+}_{(aq)}} + \underset{\text{Bicarbonate}}{4HCO_3^-{}_{(aq)}} + \underset{\text{Dissolved silica}}{H_4SiO_{4(aq)}}$	4 : 1
Pyroxene	$\underset{\text{Carbonic acid}}{4H_2CO_3} + 2H_2O + \underset{\text{Pyroxene}}{MgFe(SiO_3)_2} \text{ --> } Mg^{2+}_{(aq)} + \underset{\text{(See more about dissolved iron below in the oxidation reactions below)}}{Fe^{2+}_{(aq)}} + \underset{\text{Bicarbonate}}{4HCO_3^-{}_{(aq)}} + \underset{\text{Dissolved silica}}{2H_4SiO_{4(aq)}}$	2 : 1
Amphibole	$\underset{\text{Carbonic acid}}{14H_2CO_3} + 8H_2O + \underset{\text{Amphibole}}{(MgFe)_7Si_8O_{22}(OH)_2} \text{ --> } 7(Mg,Fe)^{2+}_{(aq)} + \underset{\text{Bicarbonate}}{14HCO_3^-{}_{(aq)}} + \underset{\text{Dissolved silica}}{8H_4SiO_{4((aq)}}$	1.75 : 1
Feldspars: Anorthite	$\underset{\text{Carbonic acid}}{2H_2CO_3} + \underset{\text{Anorthite}}{2CaAl_2Si_2O_8} + H_2O \text{ --> } 2Ca^{2+}_{(aq)} + \underset{\text{Bicarbonate}}{2HCO_3^-{}_{(aq)}} + \underset{\text{Kaolinite}}{Al_2Si_2O_5(OH)_{4(s)}}$	∞
Plagioclase (An_{50})	$6H_2CO_3 + \underset{\text{Plagioclase}}{2CaNaAl_3Si_5O_{16}} + 11H_2O \text{ --> } 2Ca^{2+}_{(aq)} + 2Na^+_{(aq)} + \underset{\text{Bicarbonate}}{6HCO_3^-{}_{(aq)}} + \underset{\text{Kaolinite}}{3Al_2Si_2O_5(OH)_{4(s)}} + \underset{\text{Dissolved silica}}{4H_4SiO_{4(aq)}}$	1.5 : 1
Albite (same stoichiometry for K-feldspar)	$\underset{\text{Carbonic acid}}{2H_2CO_3} + \underset{\text{Albite}}{2NaAlSi_3O_8} + 9H_2O \text{ --> } 2Na^+_{(aq)} + \underset{\text{Bicarbonate}}{2HCO_3^-{}_{(aq)}} + \underset{\text{Kaolinite}}{Al_2Si_2O_5(OH)_{4(s)}} + \underset{\text{Dissolved silica}}{4H_4SiO_{4(aq)}}$	1:2
Quartz	$\underset{\text{Quartz}}{SiO_{2(s)}} + 2H_2O \text{ --> } \underset{\text{Dissolved silica}}{H_4SiO_{4(aq)}}$	0:1
Calcite	$\underset{\text{Carbonic acid}}{H_2CO_3} + \underset{\text{Calcite}}{CaCO_3} \text{ --> } Ca^{2+}_{(aq)} + \underset{\text{Bicarbonate}}{2HCO_3^-{}_{(aq)}}$	
Oxidation reactions		
Dissolved iron	2 x 2 = 4 e⁻ gained; 4 x 1 = 4 e⁻ lost $\underset{\text{Dissolved iron (see products of reactions above)}}{4Fe^{2+}} + O_2 + 10H_2O \text{ -> } 4Fe(OH)_{3(s)} + \underset{\text{acidity}}{8H^+}$	
Pyrite	30 x 2 = 60 e⁻ gained; (4x6) + (4x8) = 56 e⁻ lost; 4 x 1 = 4 e⁻ lost $\underset{\text{Pyrite}}{4FeS_2} + \underset{\text{Atmospheric oxygen}}{15O_2} + 14H_2O \text{ -> } 4Fe(OH)_{3(s)} + \underbrace{8SO_4^{2-}{}_{(aq)} + 16H^+_{(aq)}}_{H_2SO_4\ =\ \text{sulfuric acid}}$	

Note the production of dissolved bicarbonate ($HCO_3^-{}_{(aq)}$) in most of the reactions on this page. This is why bicarbonate is commonly the most abundant solute in groundwater, even in non-carbonate terranes.

Figure 7-1: Some reactions in chemical weathering.

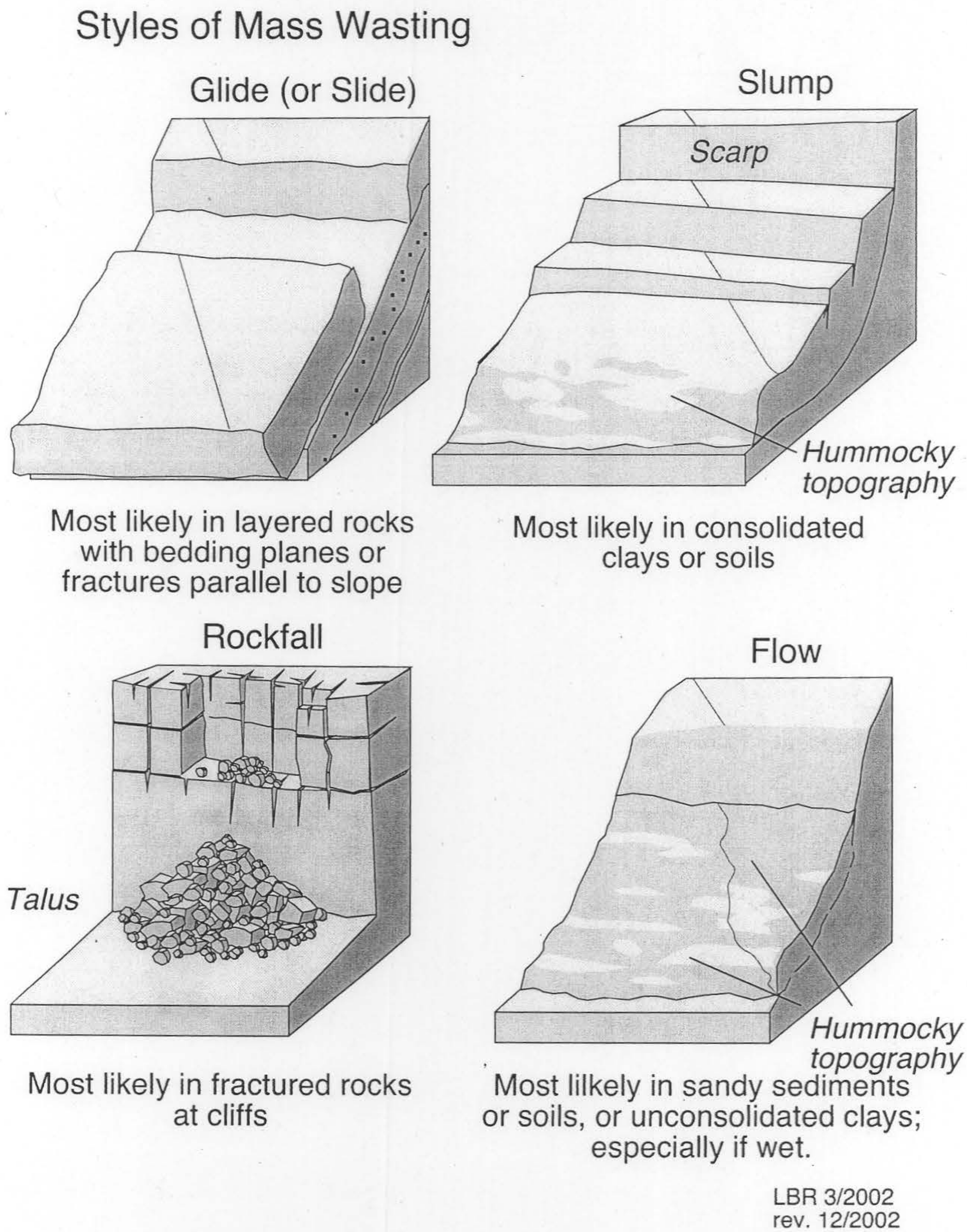

Figure 7-2: Styles of mass wasting

Chapter 8: Geologic Structures

In Chapter 4, we noted that sedimentary rocks are layered as a result of their deposition. By the 1600s, Nicholas Steno had noted three principles that should be true of such layers when deposited: they should be more-or-less horizontal, they should be laterally continuous, and they should progress upwards from older layers to younger. However, anyone who has hiked in mountains has seen rock layers at all angles, and has seen rock layers bent, broken, or sheared off. Furthermore, astute observers will have noted that the age or "up direction" of layers is sometimes inverted.

These observations of sedimentary rocks suggest that considerable deformation of rocks has taken place, and that any understanding of mountains will require understanding such deformation. Structural geology is the study of how all types of rocks deform, and of the structures that result from such deformation.

Stress, strain, and styles of deformation

Stress is the pressure put on a body of rock by the surrounding or overlying rock. It may be uniform in all directions, and it is then called "hydrostatic" stress because a body immersed in water is pressed equally by the water from all directions. Within the earth, stress is commonly not hydrostatic. It may be greatest in one horizontal direction because two regions of crust moving toward each other can cause compression, or it may be least in one horizontal direction because two regions of crust move apart. Near the Earth surface, it may be least in the vertical direction because little overlying rock mass weighs down to generate pressure, but at depth it may be greatest in the vertical direction because of the weight of the overlying rock.[39]

Strain is the deformation of rock that results from stress. If the strain is permanent, and it commonly is, then strain is the feature we can observe in a rock long after the stress has been released by erosion or other action. Thus, just as fossils are the remaining physical evidence from which we make inferences about ancient ecosystems, strain is the remaining evidence from which we make inferences about ancient arrangements of stress and thus about ancient crustal movements, and thus about ancient plate tectonic systems.

Deformation can take a variety of forms. One form, not terribly important for us, is elastic deformation. "Elastic" here is like the elastic behavior of a rubber ball: the rock is deformed slightly while under stress, but it returns to its original shape when the stress is released. Such deformation is of little importance to us because it, by definition, leaves no record of the stress that caused the deformation.[40] Of greater importance are brittle deformation and ductile deformation (Fig. 8-1). In brittle deformation, the rock breaks, and we see either fractures (breaks along which no relative motion occurred) or faults (breaks along which the two sides have moved relative to each other). In ductile deformation, the rock bends and at some scale flows, resulting in folds. Faults and folds are therefore two structures to which we will devote considerable attention in this chapter.

Brittle and ductile deformation take place in different settings and in different rocks (Fig. 8-1). Brittle deformation is more typical of low-temperature environments and thus of deformation nearer the Earth surface, whereas ductile deformation requires flow and so is more typical of high-temperature environments and thus of deformation deeper in the crust. Brittle deformation is more typical of faster deformation (or of a "higher strain rate") because rocks have no time to flow, whereas ductile deformation is more typical of slow deformation, which allows time for flow at microscopic scale. Finally, some rocks like limestones, quartzites, and granites consist of interlocking crystals that make a framework so rigid that such rocks are prone to brittle deformation. On the other hand, shales consist of fine platy clays that slide past each other to favor ductile deformation, and halite will flow in response to relatively small stresses. Thus we may

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[39] Any given stress field (the distribution of stress in three dimensions) can be described as three mutually orthogonal stresses σ_1, σ_2, and σ_3, where σ_3 is the greatest and σ_1 the least. See Figure 8-2.

[40] Elastic deformation is, however, very important in storing energy along faults that is released as earthquakes. Rocks along faults bend (deform elastically) with gradual non-seismogenic movement and then release their elastic strain abruptly to cause earthquakes.

generalize that faults are typical of near-surface deformation and folds are typical of deeper deformation, but one must also consider strain rate and rock type before reaching any reliable conclusion (Fig. 8-1).

Brittle deformation: faults and fractures

Characterization of faults requires a geometrical definition of the fault and its relationship to the bodies of rock it separates. Miners have given us a terminology for non-vertical faults that expresses geometry relative to the position of a miner in a mine tunnel that has intersected a fault. The body or block of rock below the fault is the block on which the miner stands, so it is called the "foot wall". On the other hand, the block of rock above the fault hangs menacingly over the miner's head, and so it is called the "hanging wall" (Fig. 8-2).

With this concept, we can define some simple faults. If the hanging wall above a fault has moved down relative to the foot wall, or the foot wall has moved up relative to the hanging wall, the fault is called a "**normal fault**".[41] Because motion on a normal fault increases the distance between points on opposite sides of the fault, normal faults are typical of regions of extensional stress (Fig. 8-2). Many normal faults flatten at depth, so that their motion does not require infinite displacement into the depths of the Earth. Such normal faults are called "listric faults".

If, in contrast to normal faults, the hanging wall above the fault has moved up relative to the foot wall, or the foot wall has moved down relative to the hanging wall, the fault is called a "**reverse fault**".[42] Many reverse faults have sufficiently shallow slope that the relative motion of the two blocks is horizontal more than vertical, and they are given the special designation "**thrust faults**" (Fig. 8-2). The horizontal direction of the motion of the hanging wall is the direction in which the thrust fault is said to "verge". Because motion on reverse faults, in contrast to that of normal faults, brings two points on opposite sides closer together, these faults are typical of regions of compressional stress.

If a fault is vertical, the concepts of hanging wall and foot wall loose their meaning. If the motion of the two blocks is horizontal, the fault is called a "**strike-slip fault**" (Fig. 8-2). Relative motion along a strike-slip fault is defined from the perspective of a person standing on one side of a fault and observing the motion of the opposite block. If the opposite block has moved to the right, the fault is a right-lateral or dextral fault (Fig. 8-3). In contrast, if the opposite block has moved to the left, the fault is a left-lateral or sinistral fault.[43] Strike-slip faults commonly develop in regions of lateral shear stress, or stress that wrenches in a horizontal direction (Fig. 8-2).

The definitions above and the accompanying sketches might suggest that it should be easy to identify faults in the field and to categorize them as normal, reverse, right-lateral, or left-lateral. In fact, soil cover and vegetation commonly make recognition of faults, let alone categorization of faults, difficult. One key to recognizing faults is simply to observe straight valleys between contrasting rocks, because the grinding of rock in fault zones generates a pulverized material (fault "gouge") that is more easily eroded than the surrounding rock.[44] A second clue in recognizing a fault is to find breccia in a valley between contrasting rocks, where the breccia has formed from the rock fragments in the fault zone. The direction of motion along faults can sometimes be recognized from slickensides, which are scratches or linear structures on the fault surface itself (the slickensides parallel the direction of motion). Another clue may be drag folds, which are bends in the rock along faults, where those bends have been caused by dragging of rock layers along the fault.

Finally, one must appreciate that faults need not be as simple as presented above (Fig. 8-4). Motion along a fault may take place at oblique angles, rather than conveniently parallel to or perpendicular to the strike of the fault. Also, faults

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[41] One rationalization of the name "normal" is that displacement of sedimentary strata along normal faults always leaves younger strata atop older, the normal arrangement (cf. the next footnote).

[42] One rationalization of the name "reverse fault" is the motion is opposite of, or the reverse of, a normal fault. The motion also puts older strata over younger and thus reverses the stratigraphic order.

[43] For example, the most famous strike-slip fault in North America is California's San Andreas Fault. Along the San Andreas, the southern California coast has moved northwest relative to the rest of North America (or, from the perspective of a Losangelino or Santa Barbarian, North America has moved southeast). The San Andreas Fault is thus a right-lateral fault.

[44] Note that a valley formed by the linear outcrop of a dipping shale bed can easily be mistaken for a fault.

may be reactivated, so that a normal or reverse fault may become a strike-slip fault or vice versa, and a normal fault developed in the opening of a basin may be reactivated as a reverse fault in subsequent closing of the basin.

Ductile deformation: folds

Ductile deformation bends rocks to generate folds, which can be categorized in two general types. **Anticlines** are folds in which older rocks are folded between younger ones (older rocks are the meat, and younger rocks are the taco shell). In the simplest case, anticlines are concave downward, or "A"-shaped. On the other hand, **synclines** are folds in which younger rocks are in the center of fold, between older rocks (the older rocks are the taco shell). Synclines are commonly concave upwards, or "U" shaped (and thus "Y" shaped). However, both anticlines and synclines can be rotated to positions wherein the direction of concavity or convexity is no longer relevant (Fig. 8-5).

Because points on opposite sides of a fold are brought closer together, both anticlines and synclines are typical of regions of compressional stress (and thus they commonly co-occur with thrust faults). Small folds can also develop in regions of lateral shear, where wrenching stress induces en-echelon folds.

In general, the smaller the fold, the more ductile was the rock in which it formed. Thus small (a few centimetres) and tight folds are typical of high-grade metamorphic rocks. Moderate-scale folding (several centimetres to meters) is typical of metamorphic rocks, and ductile sedimentary rocks like halite and weakly-lithified sedimentary strata. Large-scale folds that are several meters to hundreds and even thousands of meters in scale are typical of less ductile rocks, like most sedimentary rocks.

The significance of faults and folds

As alpine geologists, we will largely view faults and folds as historical documents that will help us decipher the tectonic movements that built our mountains. However, fractures, faults, and folds have great practical significance. Fractures are the storage places and conduits for groundwater in many regions of igneous and metamorphic rock, as in the Piedmont of the southeastern United States. Upward folds (antiforms and domes) are the most common structures that trap upwards-migrating petroleum, and thus they are the foremost target of the petroleum industry in supplying our modern energy-hungry society with fuel. Faults are conduits for the movement of both groundwater and petroleum, and sometimes they act to trap petroleum at depth too. As conduits for deep waters rich in dissolved solids, faults are loci for the precipitation of minerals that are commonly calcite and quartz but may include ores for silver and gold (Fig. 8-6). Thus many of the world's mines follow faults in the subsurface.

Finally, faults are of great significance to society because motion along faults releases energy that has been stored as elastic strain, and such energy moves away in the form of earthquake waves. Perhaps no part of the world better epitomizes the significance of faults and folds than California. That is where ores localized along faults and fractures inspired the gold rush of 1849, and where oil housed in anticlines in southern California fueled the petroleum industry,[45] but where an earthquake originating along the San Andreas fault destroyed San Francisco in 1906, and where earthquakes continue to threaten vast economic and social devastation.

[45] Chevron began as Standard Oil of California, and many other oil companies likewise got their start in the oil boom of southern Calfornia in the late 1800s and early 1900s.

Types of Rock Deformation

Type of deformation	*Typical action*	*Geologic result*	*Favorable geologic environment*	*Favorable rock types*	*Favorable strain rate*
Brittle	Breakage	Faults	Near surface (Low P & T)	Sandstone, Limestone, Igneous Rocks	Fast
Ductile	Bending & flowing	Folds	Deep (High P & T)	Salt, Shale, Slate, Schist	Slow

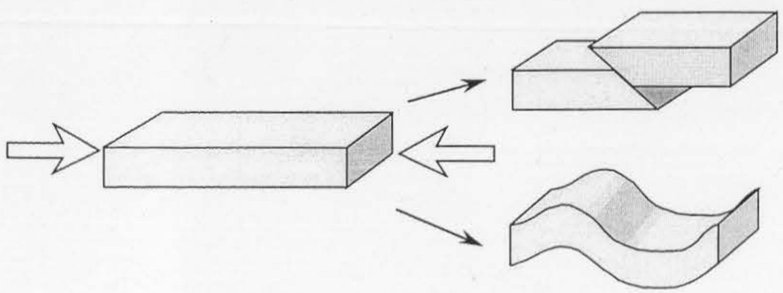

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Figure 8-1: Types of rock deformation and the factors that favor each.

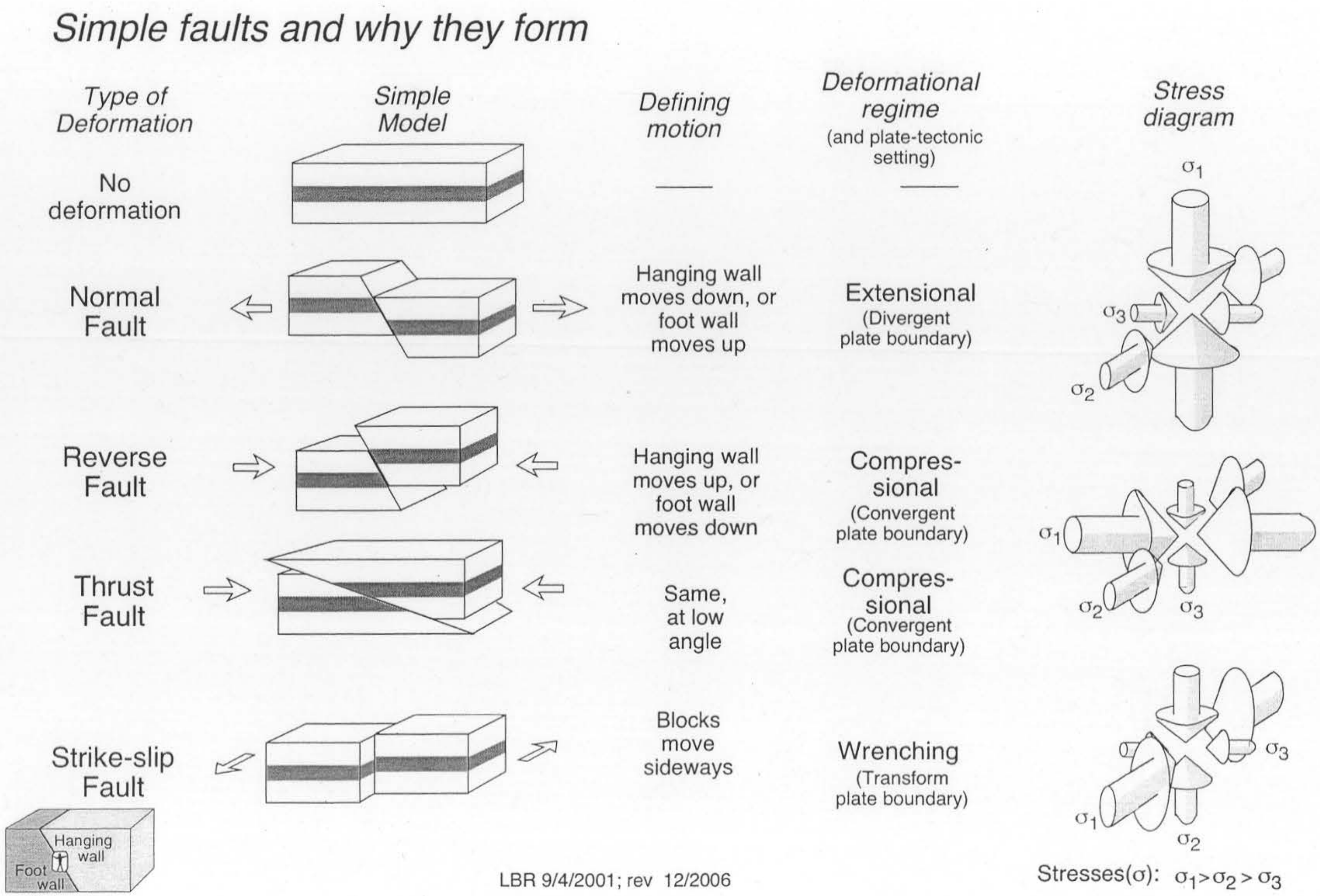

Figure 8-2: Sketches of simple faults, and the stress regimes in which they typically form.

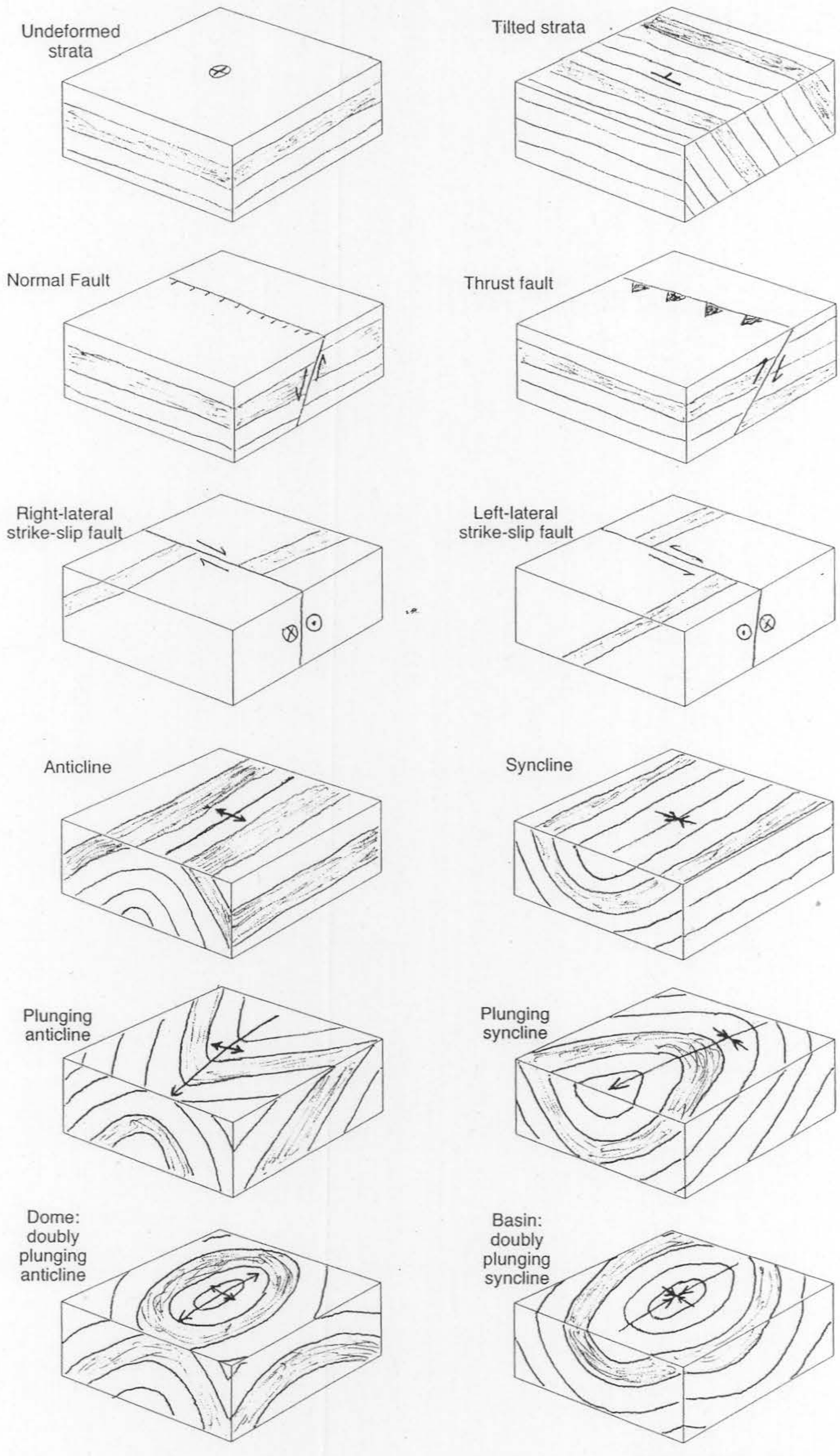

Figure 8-3. Sketches of geologic structures, with their conventional map symbols.

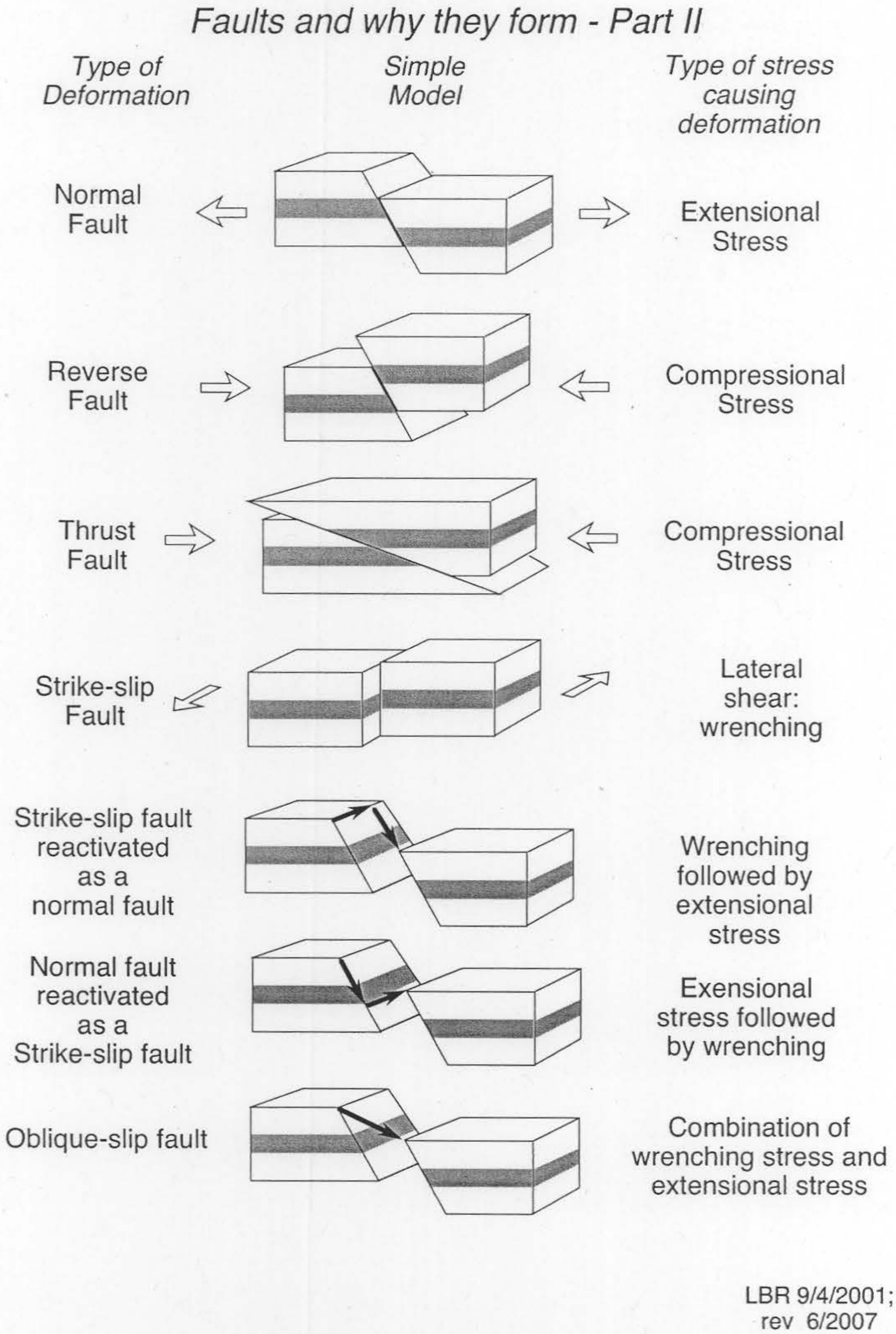

Figure 8-4: Sketches of simple faults at top and more interesting faults at bottom.

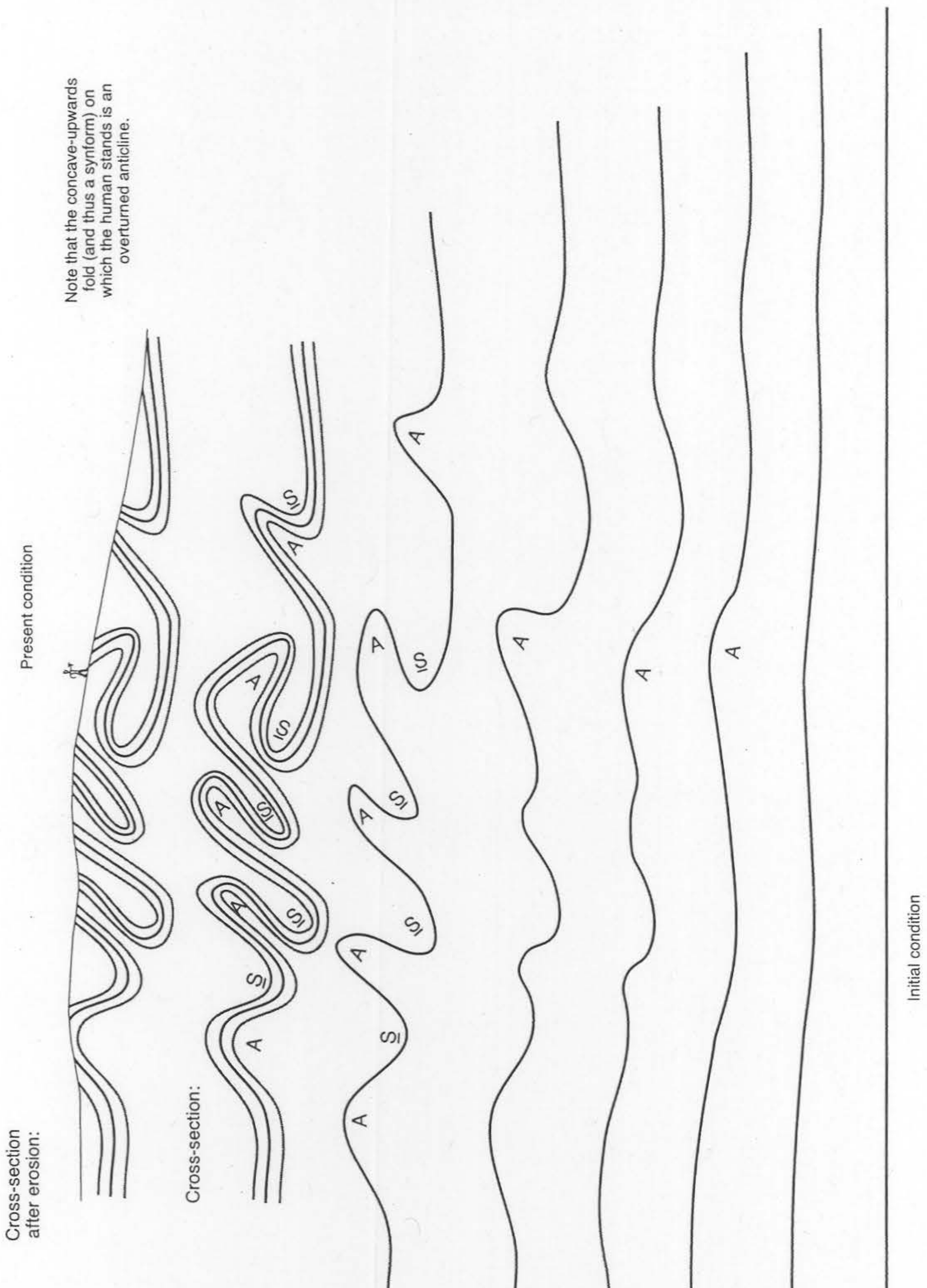

Figure 8-5: Sequential sketches of the origin of a series of folds, from the earliest and thus undeformed at the bottom to the latest and most deformed at the top. The six lowermost and earliest increments show just one layer, whereas the uppermost and latest two show three layers, solely for the sake of visual satisfaction.

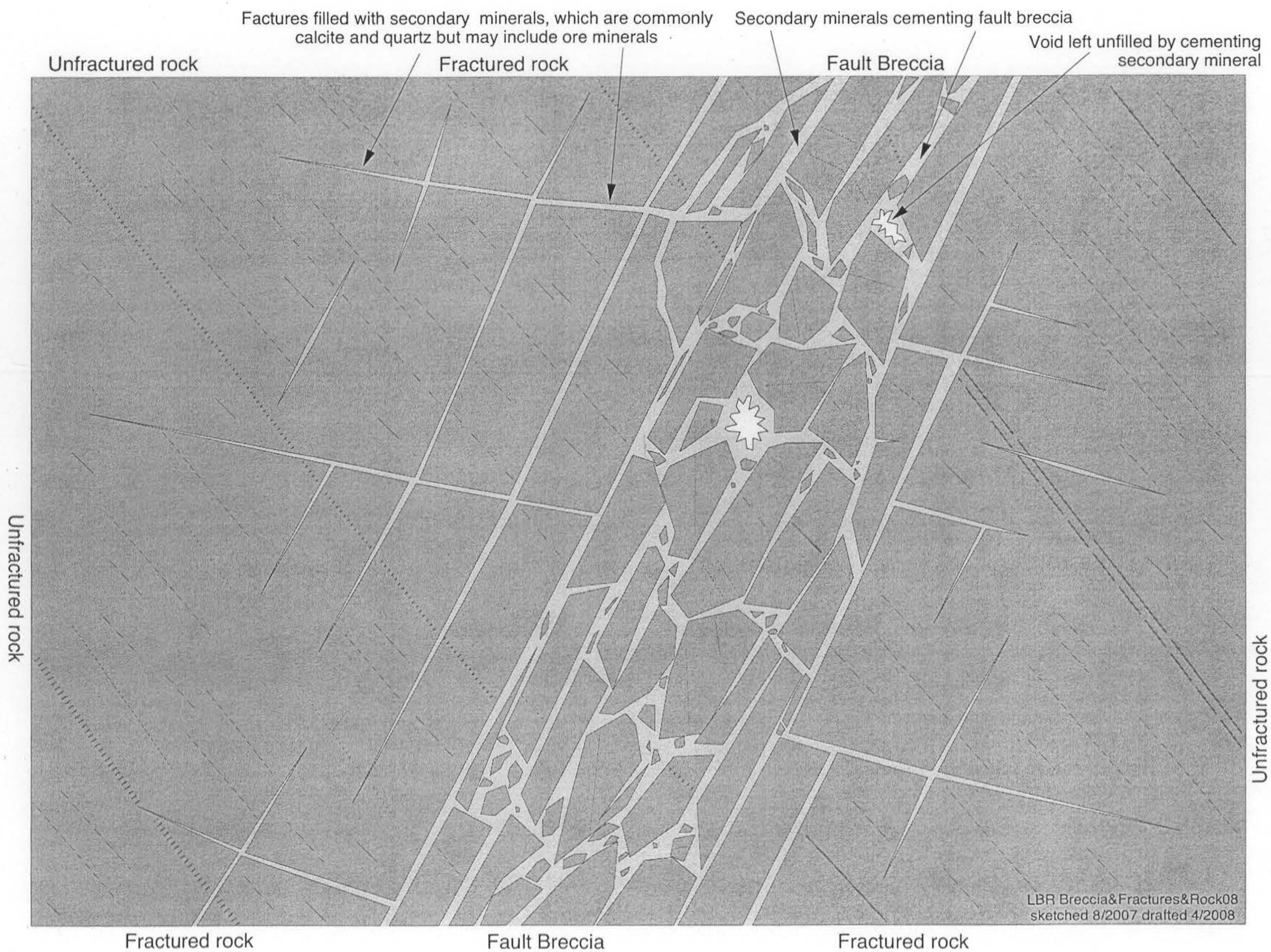

Figure 8-6: A sketch of the continuum from unfractured bedrock to fractured rock to fault breccia. This sketch compresses the continuum, in that the zone of fractured rock would be broader in reality.

CHAPTER 9. PLATE TECTONICS AND MANTLE FLOW

Geologic thinking about the motions of the continents and oceans has changed greatly in the last hundred years. We began with the experientially reasonable assumption that the continents and oceans don't move. When geologists, especially those working in the Southern Hemisphere, found evidence requiring that the continents have been joined at some times, we reluctantly concluded that the continents move via "Continental Drift", but ignorance of the ocean basins precluded a model involving them too. With deep-sea drilling and other probing of the oceans in the later 1900s, we incorporated the oceans into a model with moving plates of crust that consist of oceanic and continental crust, or of oceanic crust alone. That model, one of the two great triumphs of geology in the twentieth century, is called Plate Tectonics.[46]

Plate Tectonics, in the strict sense, is a theory of the behavior of the rigid outermost rind of the Earth - the crust or, a little more generously, the lithosphere (a term defined below). However, the oceanic crust (or lithosphere) forms from the vast mantle below and later sinks into that mantle. Any full understanding of the behavior of the Earth's surface must therefore also take into account the underlying mantle (or asthenosphere), which emerges as the ductile or *slowly* flowing mass on which the rigid plates float and move. That is why this chapter, which might be called "Plate Tectonics" in most geology textbooks, is instead called "Plate Tectonics and Mantle Flow".[47]

The significance of Plate Tectonics (and mantle flow) to alpine geology is that all the known mechanisms of mountain-building involve movement of the plates. We will see mountain ranges most commonly built at the boundaries between moving plates, but we will also see some chains of mountains in plate interiors, where plate motion has been critical to the formation of those mountains. Thus any understanding of mountains requires an understanding of plate tectonics and mantle flow.

Crust & mantle, and lithosphere & asthenosphere

The word "crust" has the implication of a rigid brittle surface, like the crust of a pie. The same applies to Earth's crust, which is indeed at least the rigid outer surface of the Earth. However, and perhaps more importantly in Geology today, "crust" also has a compositional implication of material richer in silicon than the underlying interior of the earth. Thus we speak of a relatively siliceous crust, which we divide into the sialic thicker continental crust and the more mafic and thinner oceanic crust. This contrasts with the underlying ultramafic mantle.[48]

"Crust" and "mantle" thus define compositional zones of the Earth, but they do not correspond to zones of physical behavior of the Earth. Instead, seismic analysis of the Earth has shown that the crust and uppermost mantle combine to form a zone of rigid behavior that is called the "lithosphere" (the "rocky-sphere"). This lithosphere is what constitutes the plates of Plate Tectonics. In contrast, the rest of the mantle exhibits ductile behavior and flows, and this much thicker zone is called the "asthenosphere" (the "weak sphere") (Fig. 9-1).[49] The concept of the asthenosphere as slowly flowing solid rock is difficult for us to grasp at the Earth surface, but the great temperature and pressure of Earth's interior

[46] The word "tectonics" comes from the Greek word "tekton" for carpenter or builder, in parallel with the Sanskrit word "taksan" for carpentry and axe-work (K. Frampton, Columbia University, 2001). The word thus implies the study of rigid structures (alá archi<u>tecture</u>) like those of the crust/lithosphere, but not flowing things like the mantle/asthenosphere.

The other great triumph of twentieth-century geology was establishing the age of the Earth as 4.6 billion years, in contrast to religious estimates in the thousands of years and estimates by the British physicist Lord Kelvin in the tens of millions of years.

[47] A purist might argue that the more accurate title would be "Lithospheric Slip and Asthenospheric Flow".

[48] "Sialic", "mafic", and "ultramafic" were defined in Chapter 3's section on "Composition of igneous rocks".

[49] In discussing strcutural geology, we distinguished between brittle deformation to give faults and ductile deformation to give folds. That disinction was made with regard to meso-scale structures (meters to kilometers) forming in the crust over thousands to millions of years. Here, in distinguishing between lithosphere and asthenopshere, "brittle" and "ductile" disintinguish between deformation and flow across thousands of kilomters in the crust and mantle over hundreds of millions of years.

allow rock to flow at slow rates that become significant, and in fact staggering, across geologic time.[50]

The crust, or the lithosphere, is very thin compared to the underlying mantle and asthenosphere (Fig. 9-2). Some geophysicists analogize the relationship of lithosphere to asthenosphere to that of a piece of tissue paper floating on the surface of a swimming pool, until the piece of paper takes on so much water that it becomes dense and sinks, with one of its edges sinking first and leading the rest downward. This proves to be a good analogy for the behavior of oceanic lithosphere, and especially for Earth's largest plate of lithosphere, the Pacific Plate.

Plate tectonics and plate boundaries

The basic concept of plate tectonics is that plates of lithosphere move across Earth's surface, with the direction of each plate's motion different from that of its neighbors.[51] These motions require that plates interact with each other at their margins: they must either approach each other (i.e., converge), move away from each other (i.e., diverge), or slip past each other. Thus we observe convergent, divergent, and transform (or conservative) plate boundaries (Fig. 9-3).

At **convergent plate boundaries**, lithospheric area must be eliminated. Most convergent plate boundaries are located at oceanic trenches, the deepest parts of the oceans, suggesting that one plate descends into the asthenosphere there (Fig. 9-3). The distribution of earthquakes supports this idea, because we observe planes of earthquake foci that dip away from the trenches into the earth, suggesting that some rigid body bumps and grinds its way down from the trench.

The descent of a lithospheric plate into Earth's interior is accomplished by two mechanisms that usually work together as one. The first is *subduction*[52], wherein the plate slides across Earth's surface toward the trench, bends at a hingeline, and slides downward at an angle into and through the trench (Fig. 9-4). The second is *rollback*, wherein the aforementioned hingeline migrates in the direction from which the slab[53] comes (i.e., oceanward). Thus the descending inclined portion of slab sinks directly downward at the same time that it slides down the inclined slope of the trench or subduction zone (Fig. 9-4).

The lithosphere subducted at trenches is always oceanic, rather than continental, presumably because the greater density of oceanic crust allows oceanic lithosphere to sink, whereas the buoyancy of continental crust precludes the sinking of continental lithosphere. Where two oceanic plates converge, the older oceanic plate is the one subducted, presumably because its greater age has allowed more time for it to cool from its initially very hot condition and thus become more dense.

Subduction zones are commonly overlain by volcanoes (Fig. 9-5). The reason for this relationship is that the sinking oceanic plate releases water that seeps upward into the overlying non-sinking plate. The presence of water lowers the melting temperature of most silicate minerals, especially the more sialic ones, so that this upward-moving water induces melting in the lower reaches of the overriding plate. The resulting magmas move upward, and at least some of the magmas reach the Earth surface to generate volcanoes. Where the overriding plate is oceanic, the mafic crust yields magmas that are largely basaltic, whereas if the overriding plate is continental, the sialic crust yields more variable magmas that result in volcanoes ranging from basaltic to rhyolitic. The more sialic of these magmas may also stop within the crust to form batholiths of diorite (or more precisely granodiorite) to granite.

If we have introduced oceanic-oceanic and oceanic-continental convergence here, the remaining possibility is continental-continental convergence, where the continental parts of two plates collide (Fig. 9-5). The relative buoyancy of

[50] Staggering: the building of one large mountain range commonly implies the flow of volumes of asthenosphere hundreds to thousands of times greater than the volume of that mountain range.

[51] As an analogy, one might envision a smooth ball on which pieces of thin plastic, adhering to the ball by electrostatic attraction, move across the surface of the ball. However, we will need to imagine our pieces of plastic (the plates) emerging from the ball and diving into it.

[52] The word "subduction" comes from the Latin for leading (ducere) under (sub).

[53] The word "slab" is commonly used to refer to the part of an oceanic plate that is being or has been subducted, and thus to oceanic lithospere inclined in a subduction zone or somewhere at depth in the mantle.

continental crust precludes sustained subduction, and so the only possibly result is intense compression and stacking of continental crust atop continental crust, largely via thrust faults. The result is some of the world's highest mountain ranges, and most notably the Himalayas.

At **divergent plate boundaries**, plates move apart, and the underlying asthenosphere flows up into the resulting gap (Fig. 9-3 and 9-6). The melting temperature of most solids decreases with a decrease in pressure (because the melt is less dense than the solid), and so the rising asthenosphere is prone to melting that produces mafic magmas. Much of this vast volume of magma cools to form gabbro, but some rises higher and is erupted as basalt. Most divergent plate boundaries are at the edges of oceanic plates, and so this gabbro and overlying basalt move away to become the seafloor of the oceans.[54] This process is called sea-floor spreading. The most newly-formed seafloor is thermally expanded because of its recent origin as igneous rock, whereas older seafloor has cooled and contracted. This is why almost all divergent plate boundaries are at mid-ocean ridges: the elevated nature of the ridge is caused by thermally expanded young crust there, relative to the cool and contracted crust of the older seafloor of the rest of the ocean.

Convergent plate boundaries and divergent plate boundaries combine to balance the amount of lithospheric area at a constant value, the total area of Earth's surface.[55] Every square kilometer of lithosphere destroyed by subduction at convergent plate boundaries is seemingly balanced by generation of new lithosphere by seafloor spreading at divergent plate boundaries (Fig. 9-3). The two kinds of boundary are also complementary in that old cooled contracted dense ocean lithosphere sinks into trenches at convergent plate boundaries, whereas new hot expanded buoyant lithosphere sits high at the mid-ocean ridges of divergent plate boundaries. The sinking of old cooled dense oceanic lithosphere seemingly pulls behind it oceanic plates, so that the opening of divergent plate boundaries is a passive response causing, rather than caused by, upwelling of asthenosphere below mid-ocean ridges.[56]

At **transform plate boundaries**, plates slip past each other (Fig. 9-3). Crust is thus neither created nor destroyed, leading to the more logical, if less used, name "conservative" for such boundaries. The characteristic features of such boundaries are strike-slip faults, and the boundary itself is in fact a major strike-slip fault.

One should appreciate that positions of plate boundaries are not fixed (Fig. 9-7). Rollback allows trenches, and thus convergent plate boundaries, to move seaward. One result is extension of the plate above the trench (thus leading to fore-arc basins or backarc-basins) (Fig. 9-7). Mid-ocean-ridges also commonly migrate, when symmetrical addition of new lithosphere along a divergent boundary involving a non-mobile plate means that the edge of non-mobile plate, and thus the divergent boundary, must move[55] (Fig. 9-7).

Plates, large and small

Geologists currently recognize several plates of different size across the present surface of the Earth. The largest is the Pacific Plate, a plate consisting almost entirely of oceanic lithosphere. It extends from the East Pacific Rise (a mid-ocean ridge) across most of the Pacific Ocean to the trenches near Japan. Its age, as determined by deep-sea drilling and paleomagnetic studies, progresses from very young along the East Pacific Rise to about 170 million years old, the oldest oceanic lithosphere, in the western Pacific.

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[54] Bodies of rock with this sequence from gabbro to basalt and then to marine sediments are called "ophiolites". It was traditionally assumed that all ophiolite sequences found on land were slivers of abyssal sea floor shoved onto the edges of continents, but more recently researchers have realized that many ophiolites sequences have other origins not at mid-ocean ridges.

[55] It is generally assumed that Earth's volume and surface area are essentially constant. Some theorists early in the development of plate tectonics focused on convergent plate boundaries and inferred a shrinking Earth, whereas other theorists focused on divergent plate boundaries and inferred an expanding Earth. However, the lack of evidence for, or a mechanism for, volume change leads virtually all Earth scientists today to assume constant volume and surface area.

[56] The non-dependence of mid-ocean ridges and divergent boundaries on particular locations of underlying asthenosphere is demonstrated by the migration of the mid-ocean ridges away from the African and Antarctic continents. Oceanic lithosphere cannot have been added at the edges of these plates without movement of the mid-ocean ridges relative to the asthenosphere.

Other large plates include the Eurasian Plate, the North American Plate, the South American Plate, the African Plate, the Indian-Australian Plate, and the Antarctic Plate, all of which consist of both continental and oceanic lithosphere. There are other smaller plates, descending in scale to so-called "microplates". Microplates are commonly found in the complex regions where three or more plate boundaries come together to create a complicated plate-tectonic landscape.

The evidence for plate tectonics

At this point, a reader might reasonably ask, "Why should I believe all these claims about moving continents, and about seafloor that forms here, moves, and sinks into the Earth over there?" The reason is that there is much evidence for these ideas, or for plate tectonics:

- *The continents fit together nicely geographically.* The eastern margin of South America most clearly fits into the bight of western Africa, but in fact South America, Africa, North America, and Europe fit together very well, especially when one includes their continental shelves, which are just the submerged parts of the continents. This makes sense only if these continents were once together and subsequently moved apart.
- *Many geologic features of the continents only make sense if those continents were once together.* These include evidence of Late Paleozoic glaciation in the Southern Hemisphere, wherein glaciers seem to have emerged from the oceans onto tropical to temperate continental regions (if we only assume present geography). If we mentally move the continents backward to where they fit together, the tracks of the glaciers diverge from one near-polar region onto all the southern-hemisphere continents. The distribution of Late Paleozoic terrestrial fossils, especially the famous *Glossopteris,* are similarly only explicable if the continents have moved apart. Likewise, ages of Precambrian "basement" rocks have boundaries and trends that line up if one moves South America and Africa back together in a pre-Atlantic geography.
- *The lateral distribution of earthquakes supports the idea of rigid plates that bump against each other at the margins but have little deformation within.* The vast majority of earthquakes are at the mid-ocean ridges and trenches that we have identified as plate boundaries.
- *The vertical distribution of earthquakes supports the idea that oceanic plates move down into the Earth below trenches.* At trenches, earthquakes below the over-riding plate are progressively deeper farther from the trench, compatible with the notion that these earthquakes occur on the top side of a subducting slab that is diving under the over-riding plate.
- *Radiometrically-determined ages of dredged or drilled samples of seafloor show that the youngest seafloor is at the mid-ocean ridges, with progressively older seafloor away from the mid-ocean ridges.* This supports the idea that seafloor forms at the mid-ocean ridges and then moves away from them.
- *Great heat flow, hydrothermal vents, and volcanic eruptions at mid-ocean ridges all support the idea that hot earth material (asthenosphere) wells up underneath those ridges.* By contrast, low heat flow near trenches suggests that cold oceanic crust sinks into the Earth there.
- *The distribution of ages of at least some ocean volcanoes supports the idea that those volcanoes formed on an oceanic plate moving away from a mid-ocean ridge and toward a trench.* The classic example is the Hawaiian chain of volcanoes, which are progressively older to the northwest of the present active volcano(es) on the Big Island. They seemed to have formed from an underlying source of magma as the Pacific Plate moved away from the East Pacific Rise and toward the trenches of the northwestern Pacific.
- *GPS measurements show that regions of Earth's surface are moving relative to each other.* This modern observation, independent of any geologic theory, is the latest and most inescapable piece of evidence for moving plates of Earth's surface.

With that said, we can turn to how these motions of Earth's surface (i.e., the lithosphere) interact with Earth's interior.

Plate tectonics and mantle flow

If you stretch your hand out, with your fingers pointing down, and move your hand straight down into a tub of water, your hand displaces water. The water below your fingertips must flow away as your fingertips advance downward, that water must displace other water, which displaces other water, etc. Ultimately water rises to the surface, but we have no expectation that the rising water is the same water that was directly beneath your fingers - it's only water driven by the chain of displacement begun by your fingers. The same must happen at much larger scale when a subducted slab descends into the asthenosphere (left side of Fig. 9-4). There must be flow of the mantle (or, more correctly, of the asthenosphere).

To envision the effect of rollback (right side of Fig. 9-4), now put your outstretched hand into the tub of water and pull your hand toward you. Water must move. If the tub is shallow, water must move around the sides of your hands, from the palm side to the backhand side. If the tub is shallow and narrow, water will well up on the palm side and its level will drop on the back-hand side. Again, the water is the asthenosphere as we consider the dynamics of plate tectonics, and the hand is now a slab that is rolling back as it is subducted.

Current thinking about Earth's interior suggests that, like the hand descending into or moving through the water, movement of lithosphere from the surface is the mechanism driving flow in the mantle.[57] In the words of a recent paper, "Theoretical studies indicate that plate motion is primarily controlled (~90%) by convective [mantle] flow driven by density heterogeneities in the mantle, particularly those associated with sinking oceanic slabs".[58] In this view, movement of plates and mantle flow are a by-product of subduction and rollback. Likewise, the rising of hot asthenosphere at mid-ocean ridges results from flow that is driven by sinking plates, and upwelling of the asthenosphere is enabled by the opening of the passive crack pulled apart at the mid-ocean ridge (Fig. 9-8). Some earlier models suggested essentially the opposite – that movement of the plates was driven by flow of the mantle, which was in turn driven by convection of heat from Earth's core. However, layering of the mantle precludes such whole-mantle convective flow, and migration of mid-ocean ridges defies the notion of fixed cells of convection. Instead, sinking plates seemingly drive plate motions, or in some areas they set in motion asthenospheric flow that moves, or contributes to the motion of, overlying plates (Fig. 9-8). Thus "slab pull" of plates and "top-down tectonics" driving flow of the upper mantle seem to be our best explanation of plate tectonics and mantle flow.

Two views of rollback and mantle flow

Figures 9-8 and 9-9 present two different perspectives on rollback and mantle flow. Figure 9-8 is a vertical cross-section that assumes no variability in the third dimension into or out of the plane of the page. Thus, in Figure 9-8, rollback of the sinking slab of oceanic lithosphere forces (on the bottom side of the slab) asthenosphere to flow toward to the mid-ocean-ridge from which the oceanic lithosphere came. At the same, rollback of that same slab leaves behind (on the top side of the slab) volume into which the overriding plate moves and, at greater depth, volume into which asthenosphere must flow. Meanwhile, the subductive motion of the slab pushes a volume of subducted lithosphere into the asthenosphere below the overlying plate, acting as a piston to displace asthenosphere that must flow to fill the volumes vacated. The result is large-scale mantle flow that accounts for the upwelling of asthenosphere at mid-ocean ridges and, in part, for the motion of continents (Fig 9-8).

A second and more local view of rollback and mantle flow is that in Figure 9-9. Figure 9-9 assumes that the subducting slab has lateral edges around which asthenosphere can flow. Thus asthenospheric mantle can flow from behind (oceanward of) the sinking slab and flow around the edges of the slab to the region under the overriding plate (Fig. 9-9). (In the analogy of the outstretched hand pulled through water above, this is the case is which water moves in a sideways curve from the palm side to the backhand side of your hand.) The result of this flow is that the center of the subducting plate rolls back faster than the edges of the subducting slab, so that the trench becomes convex oceanward (Schellart et al. 2007). This is why the volcanoes on the overlying plate generally lie in a convex-oceanward arc, thus giving the expression "island arc volcanoes" (Fig. 9-9)

The other result of rollback and the oceanward migration of the edge of the overriding plate is extension of the overriding plate (Figs 9-7

[57] Anderson, D.L., 2001, Top-down tectonics?: *Science,* v. 293, p. 2016-2018.

[58] Eaton, D.W., and Frederiksen, A., 2007, Seismic evidence for convection-driven motion of the North American plate: *Nature,* c. 446, p. 428-431. See also Conrad, C.P., and Lithgow-Bertelloni, C., 2002, How mantle slabs drive plate tectonics: *Science,* v. 298, p. 207-209. The only bone of contention in the quotation would be the word "convection", which generally implies roughly circular cells of flow driven by heating from below.

and 9-9). In some cases, this leads to mid-ocean-ridge-like spreading zones and the generation of new oceanic lithosphere in the back-arc region. Rollback away from a continent can pull away a sliver of that continent and cause development of a back-arc basin between the continental sliver and the rest of continent. This seems to have happened with pulling of Japan away from the Asian mainland and opening of the Sea of Japan, all as a result of the great age of the oceanic lithosphere in the western margin of the Pacific Plate. Alternately, extension may take place on the oceanward side of the volcanic arc, giving a fore-arc basin (Fig. 9-7).

The visions of rollback and mantle flow in Figures 9-8 and 9-9 are not mutually exclusive, in that some asthenosphere may flow back under the subducting plate and some may flow around the edges of the descending slab.

Rollback and styles of deformation at plate boundaries

Rollback of oceanic lithosphere complicates an otherwise convenient correlation of plate tectonic settings with styles of structural deformation. Firstly, divergent plate boundaries are dominated by extensional structures like normal faults, because extension is the rule there. So far, so good. Secondly, transform plate boundaries are dominated by strike-slip faults because of the slippage of plates past each other. So far, still so good. Thirdly, convergent plate boundaries at which continental lithosphere collides against continental lithosphere are dominated by compressional features like thrust faults and major folds. *However,* deformation at convergent plate boundaries where oceanic lithosphere is subducted can range from compressional (where rollback is minimal) to at least partly extensional (where rollback happens). We will return to that thought with regard to mountain-building in Chapter 15, when we tour the Pacific Rim.

Subduction isn't so simple: oblique subduction

Subduction as discussed above is assumed to be orthogonal: both the subducting and over-riding plates are assumed to move perpendicular to the subduction zone. That need not be true, and it cannot be true wherever the edge of the overriding plate is not a straight line (Fig. 9-10). Instead, subduction is commonly oblique, or not perpendicular. Oblique subduction is significant because it generates wrenching stresses that lead to strike-slip faulting in at least the over-riding plate (Fig. 9-10). Cross-sections like those in Figure 9-5 that show compressional features in the plane of the page must thus additionally include strike-slip faulting that moves material in and out of the plane of the page, and maps of mountain belts generated by oblique subduction will show strike-slip faults, as well as normal faults.

Sources and readings

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Hamilton, W.B., 2007, Driving mechanism and 3-D circulation of plate tectonics, in Sears, J.W., Harms, T.A., and Evenchick, C.A., eds., *Whence the mountains? Inquires into the evolution of orogenic systems: A volume in honor of Raymond A. Price*: Geological Society of America Special Paper 433, p. 1-25.

Schellart, W.P., et al., 2007, Evolution and diversity of subduction zones controlled by slab width: *Nature,* v. 446, p. 308-311.

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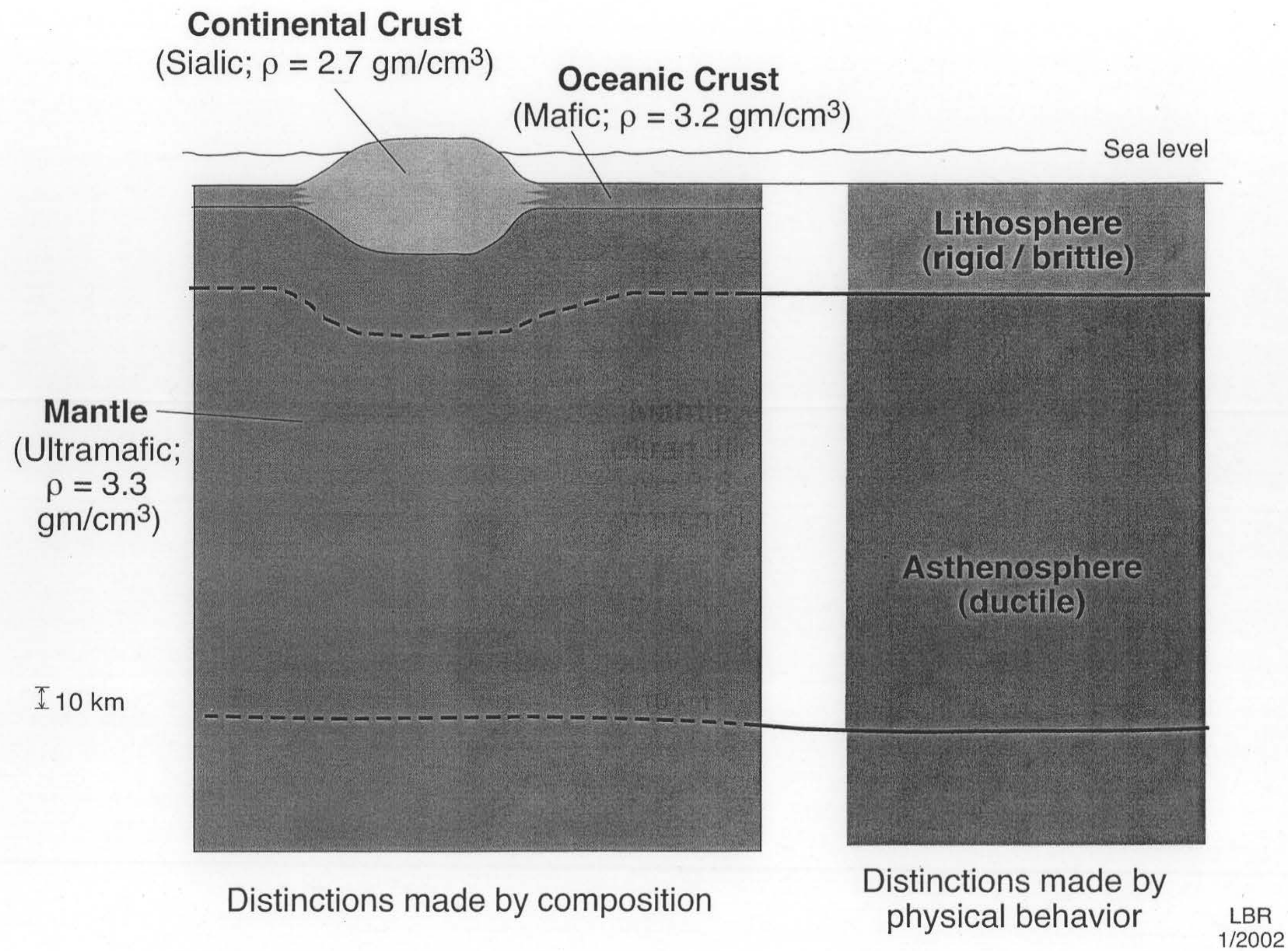

Figure 9-1: Terminology for Earth's interior.

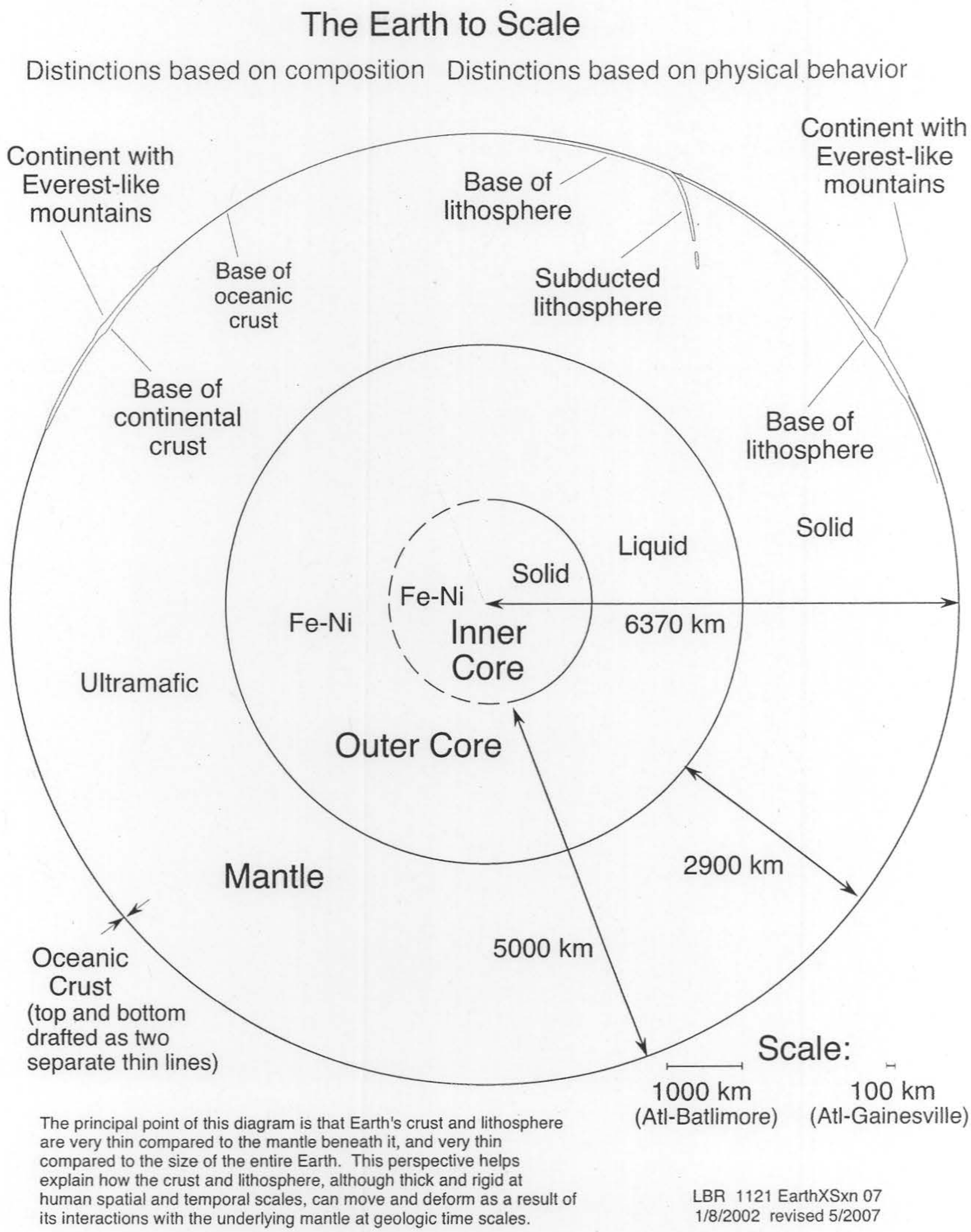

Figure 9-2: A cross-section of the Earth.

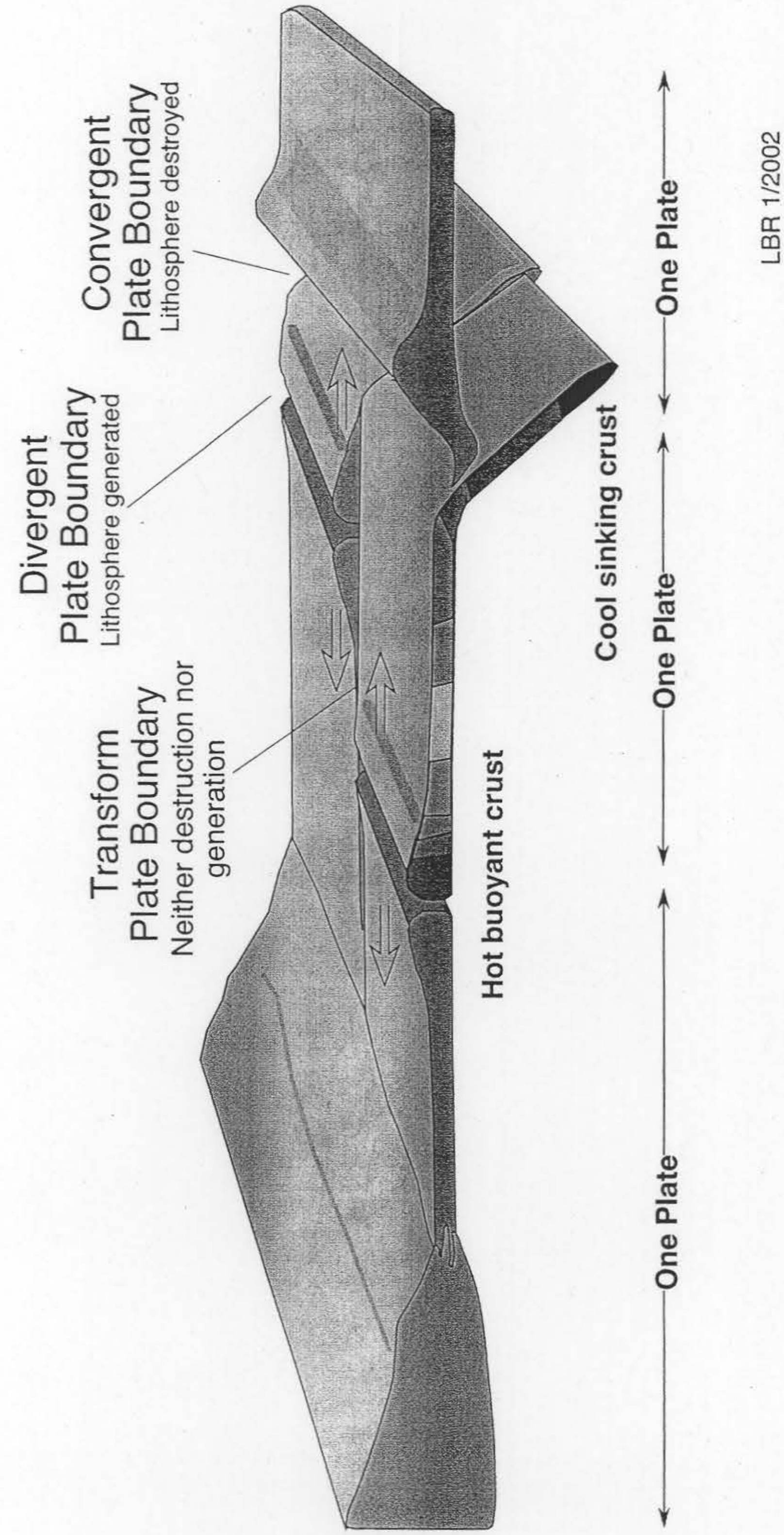

Figure 9-3: A schematic three-dimensional view of plate tectonics.

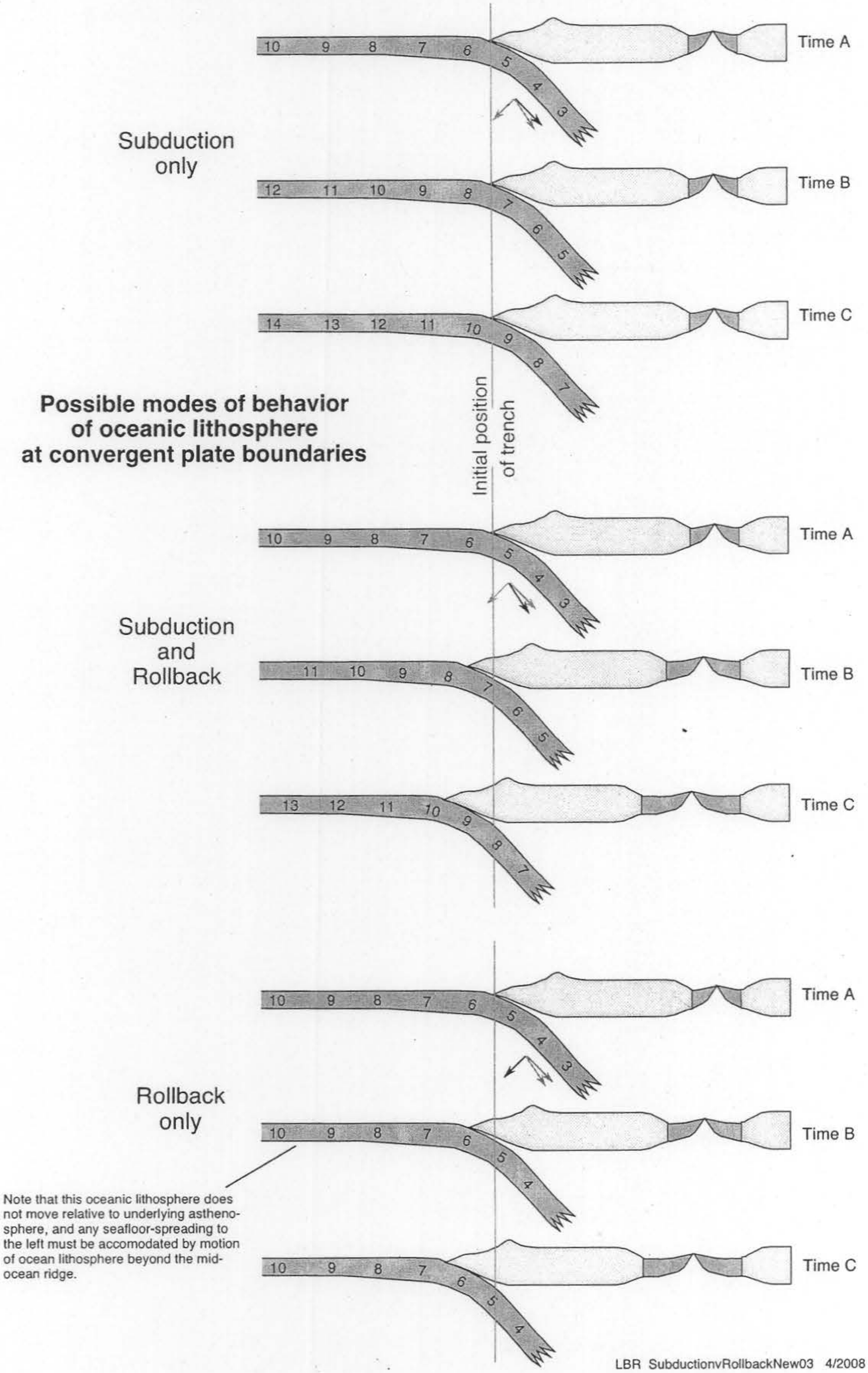

Figure 9-4: Sketches of subduction and roll-back. Rollback is also illustrated in Figures 9-8 and 9-9.

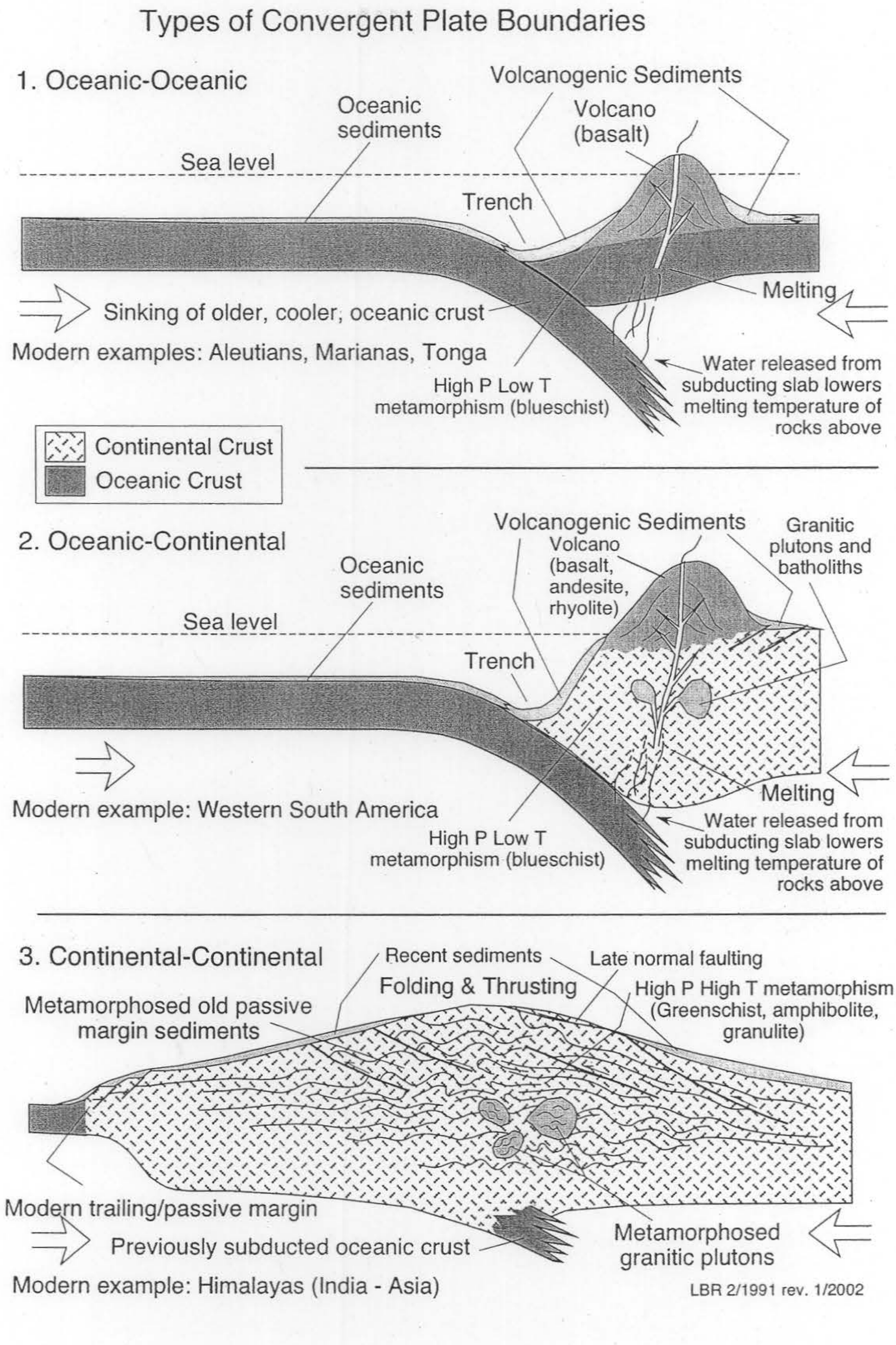

Figure 9-5: Cross-sections of the three kinds of convergent plate boundaries.

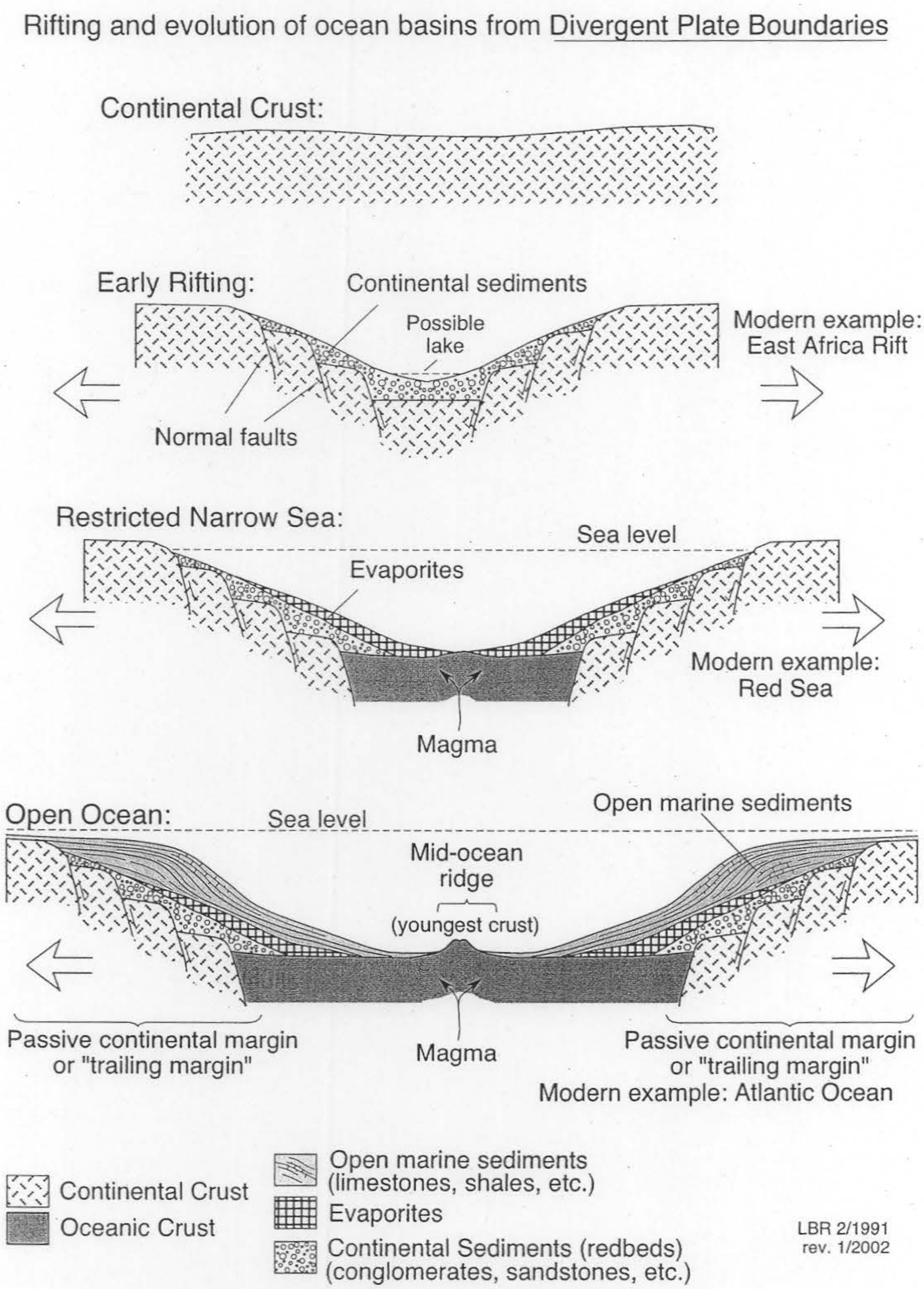

Figure 9-6: Cross-sections illustrating the development through time of divergent plate boundaries.

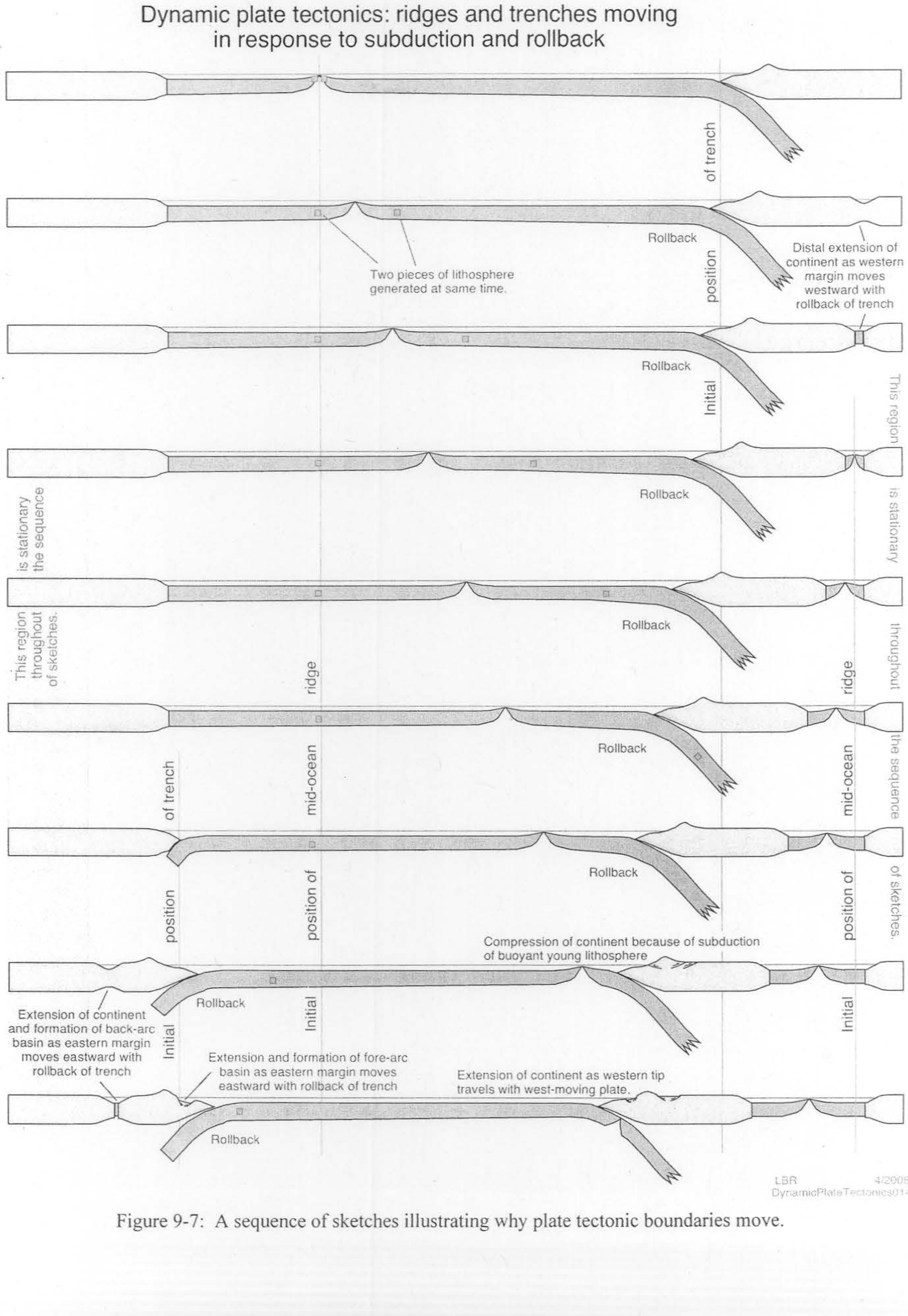

Figure 9-7: A sequence of sketches illustrating why plate tectonic boundaries move.

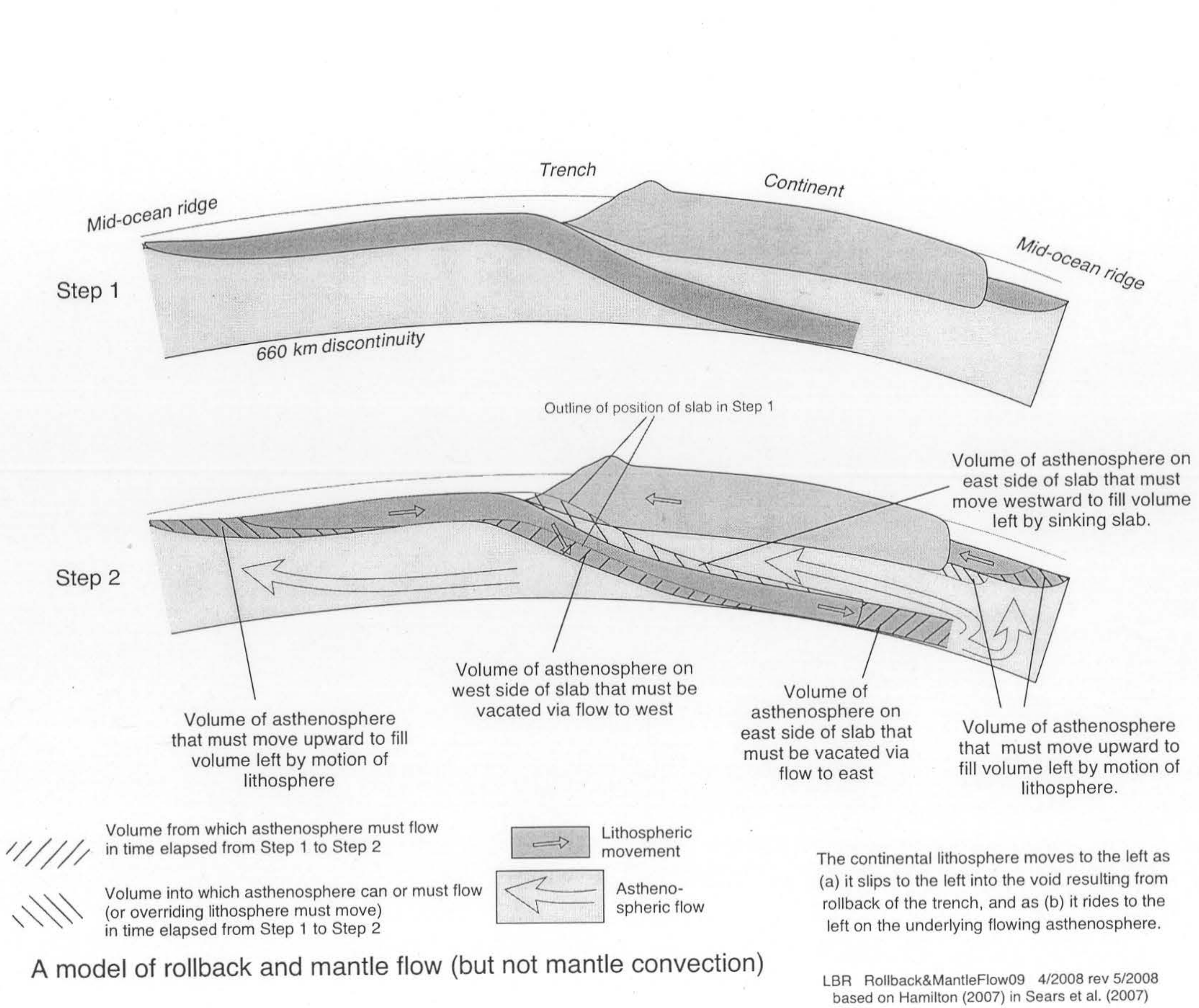

Figure 9-8: Sketches illustrating how rollback drives flow of the asthenosphere. These sketches assume that lithospheric slabs "pond" above the 660-kilometer discontinuity in the mantle, rather than sink through the entire mantle, but the latter would only change the vertical scale of the scheme shown.

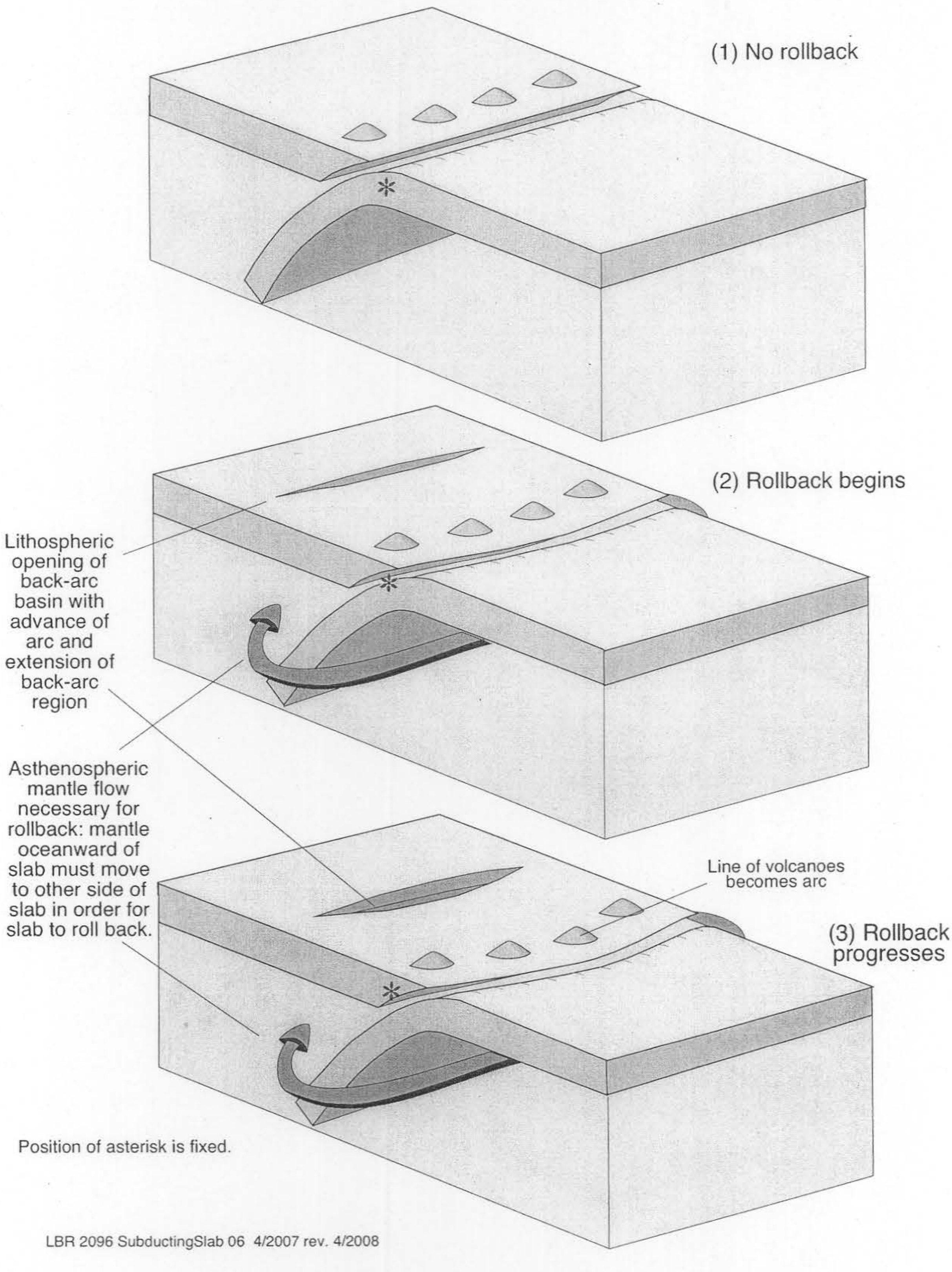

Figure 9-9: Sketches illustrating how rollback generates a curved trench and island arc.

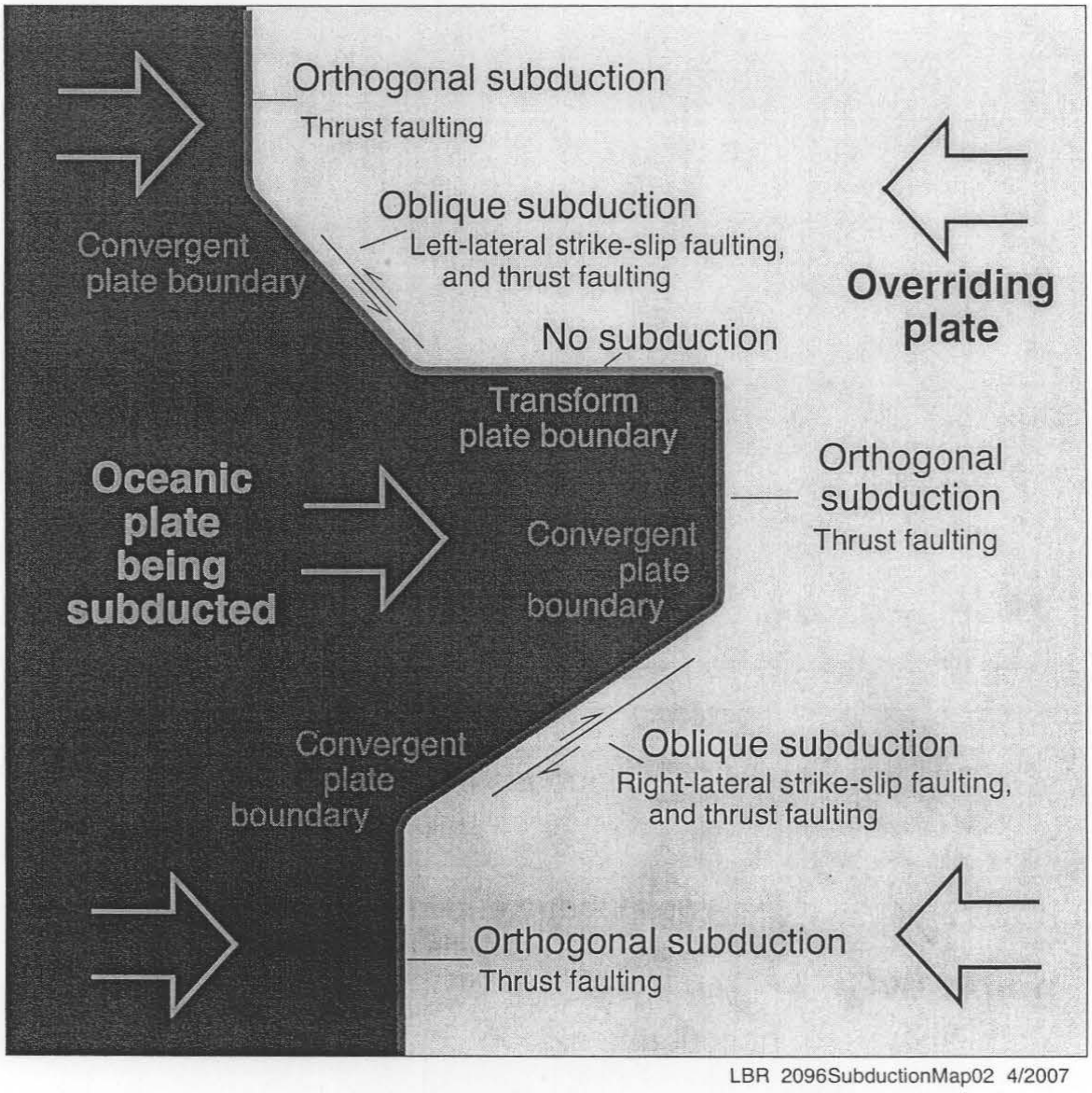

Figure 9-10: Schematic map illustrating some possible geometries of subduction.

Part II. Mountains and Orogeny

CHAPTER 10. OROGENIC PROCESSES

This chapter introduces some of the many ways in which mountains and mountain ranges form, and thus it covers orogenic[59] processes. The chapter generally works from small to large, and it begins with features both so small and so structurally simple that many professional geologists, and especially tectonicists, would not consider them mountains[60] (Fig. 10-1). Later chapters of Part II will focus on specific examples of the larger kinds of mountains and mountain ranges.

1. Erosion to produce isolated uplands

Differential erosion can work in a variety of ways to produce isolated uplands. These uplands are usually not as large as most mountains, but they are sufficiently steep-sided that at least some are called "mountains" and so are mentioned here.

Tower Karst[61] is a landscape developed on terrains of limestone that have been riddled with caves. Erosion cuts away large volumes of the honeycombed limestone to leave towers or pinnacles of limestone that commonly has many caves too. Such landscapes are best known from southern China, where they have been the subject of dramatic landscape paintings for centuries.

Development of tower karst requires a wet climate to promote dissolution of limestone to form caves. It also requires either uplift of the land or dramatic lowering of the water table to leave the honeycombed limestone above the water table and susceptible to erosion.

Mesas form in flat-lying sedimentary rocks when strata that are easily eroded underlie a bed resistant to erosion. Across much of the landscape, both kinds of rocks are removed, but isolated patches of the bed resistant to erosion cap uplands that remain as flat-topped hills or small mountains. **Cuestas** form by the same process in tilted, rather than flat-lying, strata and so are hills or small mountains with sloping planar tops.

Mesas are common in the southwestern U.S.; the steeped-sided flat-topped features of Monument Valley, Canyonlands National Park, and Mesa Verde are good examples. However, mesas can also form in more humid climates. Lookout Mountain above Chattanooga, Tennessee, is a good example of a mesa in a humid climate, in that it is a flat-topped mountain capped by flat-lying Pennsylvanian sandstones. The tree-covered Tepuis of Venezuela are likewise famous mesa-like mountains in a humid climate.

Remnants of plutons can form striking isolated uplands. These features result from preferential erosion of the metamorphic rocks into

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[59] "Orogeny" as a general term refers to building of mountains, especially where lithospheric plates converge. The word is also used to denote any one mountain building event, as in "the Himalayan Orogeny". "Orogen" is a genetic term for a region or body of rock formed in one orogeny, and thus an orogen is commonly, but not necessarily, a mountain range (which is a more strictly geographic term).

[60] In conventional usage, hills are small and mountains are large. To most geologists, hills have been shaped solely by erosion, whereas mountains and mountain ranges have been shaped by faulting, folding, and tectonic uplift. Thus the Black Hills are geologically mountains and Stone Mountain is geologically just a hill.

If a mountain is a large topographic feature originating from some geologic process beyond erosion, a mountain range is an elongate array of such features. A massif (to paraphrase the American Geological Institute's Glossary of Geology) is a huge topographic feature, often within an orogenic belt, that is commonly formed of rocks more rigid than those surrounding it. Massifs thus generally consist of igneous or metamorphic rocks.

The German language has two words for "mountains". "Berg" is the singular of one for which the plural "Berge" has the sense of multiple mountains, regardless of their relationship (hence the common sign "Halten Sie die Berge sauber": "Keep the mountains clean"). On the other hand, "Gebirge" means "mountains" in the sense of a related set or a mountain range, as in the "Wetterstein Gebirge" in northern Tirol.

[61] "Karst" is a term for any landscape in which chemical weathering has proceeded so far as to dissolve large volumes of rock to produce caves, sinkholes, or other dissolution features. Most karst develops on limestone bedrock.

which a pluton has been intruded, while the plutonic rock is more resistant to erosion and is left standing above the surrounding landscape. Stone Mountain east of Atlanta, Georgia, is a well-known example, as is the nearby Panola Mountain. Sugarloaf Mountain above Rio de Janeiro is another well-known example. Stone Mountain and Sugarloaf Mountain are examples of **monadnocks**, isolated hills or mountains of resistant rock rising above eroded lower land. Mount Monadnock in New Hampshire, the mountain from which the name comes, is likewise a granitic pluton. **Inselberg** is another term for a mountain originating as an isolated erosional remnant, and Mt. Mulanje in southeast Malawi is a prominent example associated with that term.

Other erosional remnants that are less easily explained or categorized can nonetheless make prominent hills or mountains. Perhaps the most famous monadnock is Ayers Rock or Uluru in the Northern Territory of Australia, an example in which sandstone with vertical bedding sits above an otherwise strikingly flat landscape, seemingly as just an isolated remnant of much more extensive exposures long ago.

2. Extension and normal faulting

Extension of the earth's crust across large areas can cause a series of normal faults to form. Most commonly, these are listric normal faults (normal faults that flatten at depth) that dip toward the center of the area of extension. On each fault block, the side toward the center of the area of extension is tilted upward to make a linear mountain range. The other side of each fault block, which is the side toward to the *edge* of the area of extension, is tilted downward to form a linear basin in which sediments accumulate.[62]

The foremost example of this system of mountain ranges is the Basin and Range province in Nevada and western Utah. Mountain ranges stretching north to south are separated by basins filled with Quaternary sediments, some of which are topped by salt flats in the present arid climate. Death Valley is the westernmost basin of the Basin and Range, and as the lowest point in North America it nicely illustrates the downdropped nature of the basins of the Basin and Range. The Basin and Range extends eastward to its last basin, the Great Basin of Utah, in which the modern Great Salt Lake sits.[63]

3. Hotspot volcanism, and other models for linear volcanic chains

The "Hotspot" model is commonly used to explain linear[64] chains of volcanic islands and seamounts[65], of which Hawaii and the Emperor Seamount Chain are the classic example. The model also accounts for large isolated volcanoes on land, like Mt. Kilimanjaro. In this model (Fig. 10-4), a tectonic plate moves across a plume of magma rising from the mantle. The magma breaks though the lithosphere to build a volcano, but movement of the plate carries that volcano off the plume and so the volcano becomes dormant and contracts as it cools. A new volcano forms over the plume or "hotspot" until it too is carried away, and yet another forms and so on until a chain of volcanoes has formed. Ages of islands in the Hawaiian chain support this model nicely, in that the islands are progressively older to the northwest, the direction in which the Pacific plate is thought to move. Most of the other linear chains of volcanoes and seamounts in the Pacific are

[62] Textbooks commonly describe these areas as a succession of faults dipping in opposite directions. The result is a series of elongate uplifted crustal blocks called "horsts" that alternate with downdropped blocks called "grabens". The horsts are mountain ranges that expose previously buried rocks, whereas the grabens are sites of accumulation of sediments and thus are covered with young sediments and sedimentary rocks. However, horst-and-graben systems may be less common in nature than such textbooks would suggest.

[63] The expression "the modern Great Salt Lake" is used because the present lake is much smaller than ancient Lake Bonneville, the Pleistocene-age lake that filled the Great Basin and ultimately spilled northward to the Snake River. Shorelines of Lake Bonneville are readily visible high up on the Wasatch Front, the slope that rises abruptly east of Salt Lake City.

[64] The word "linear" is used repeatedly here to emphasize that this explanation does not apply to island arcs of volcanoes associated with trenches and subduction.

[65] A seamount is a submarine volcano with a top that does not reach sea level. A guyot is a flat-topped submarine volcano, and its flat top is believed to have formed by subaerial and wave erosion when the volcano was younger and larger before it contracted to it present smaller size. Old seamounts are likewise thought to be smaller than their original hot expanded size.

believed to have likewise formed at hotspots over plumes rising from the mantle[66].

The Hawaiian chain is the classic example of hotspot volcanism, and its size is a testimony to the extent of such volcanism. The Big Island of Hawaii rises from the seafloor six kilometers below sea level to heights more the three kilometers above sea level. It thus dwarfs all individual mountains on land, such as Mt. Everest, and in fact in cross-section it is larger than the entire Himalayas (Fig. 10-1). Hot spot volcanism has thus produced Earth's largest mountain with Hawaii's Big Island, and even other hotspot outpourings, such as the Canary Islands and Mt. Kilimanjaro, are huge in comparison to other non-volcanic mountains (Fig. 10-1).

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[66] Having learned about the hotpot-plume model, one should nonetheless appreciate that some linear chains of volcanoes arise in other ways. One striking example is a linear array of volcanoes extending from the east-central Atlantic northeast into Cameroon. Casual comparison with the Hawaiian example would suggest that all these volcanoes formed over a hotspot in the east-central Atlantic. In fact, the ages of the volcanoes do not increase either to the northeast or the southwest and instead are a random series. The explanation of this line of volcanoes is that rifting of the Atlantic left a third failed arm of the rift that has served as a crack in the lithosphere through which magmas have sporadically leaked to the earth surface. Their linear arrangement thus has nothing to do with age or plate movement, and it formed very differently from the commonly accepted model for Hawaiian-style chains.

One should further appreciate that some geologists have come to question the hotspot-plume model used to explain Hawaiian-style chains. One alternative model for such chains is the propagating crack model in which a crack propagates across the crust, allowing sequential leakage of magma upwards to generate a series of volcanoes with ages decreasing the direction of the propagation of the crack. If one used such a model to explain the Hawaiian chain, the specific application would call for a crack propagating to the southeast with an orientation only coincidentally roughly parallel to the inferred direction of movement of the Pacific plate. For more, see www.gly.uga.edu/railsback/volcano-chains.html.

At this writing, students should still understand and tentatively accept the hotspot-plume model, but they should know that some linear volcanic chains have clearly formed by other means, and they should appreciate that the hot-spot plume model is not universally accepted and, like all theories, may ultimately be rejected if evidence is found to refute it.

4. Compression and high-angle faulting

Within continents, compression sometimes results in uplift of crustal blocks between high-angle faults that may be reverse faults (Fig. 10-5). Erosion soon removes relatively young sedimentary rocks from the top of the uplifted block, commonly exposing Precambrian igneous and metamorphic rocks. At the flanks of the uplift, surrounding sedimentary strata are dragged upwards. Among these strata, those more resistant to erosion form flatirons or hogbacks ringing the uplifted crystalline rocks.

Classic examples of mountain ranges formed in this way include the Black Hills of South Dakota and the Bighorn and Wind River mountains of Wyoming. The Black Hills, although small, have all the characteristics of such a mountain range: upturned younger sedimentary flanks, partly but not completely eroded sedimentary cover, and an exposed core of Precambrian igneous and metamorphic rocks. The igneous rock is in this case the granite from which the huge sculptures at Mount Rushmore and Crazy Horse have been blasted and carved.

These uplifts are commonly considered separately, but the parallelism of the northnorthwest-to-southsoutheast axes of these mountain ranges suggests a common origin. Those axes are perpendicularly to the Snake River Plain and thus to the apparent direction of motion of the North American plate, suggesting that the compression responsible for these mountains may be associated with plate movement. The compression has also been linked to subduction of a shallowly-dipping plate in the subduction zone off western North America, allowing compression further inland than that seen in most subductive settings.

5. Fold-and-thrust belts

Compression can also cause thrust-faulting that allows shortening of the crust at shallow depths. This style of thrusting occurs most commonly in sedimentary rocks, and a single more deformable stratum is commonly the surface (the "decollement") along which much horizontal slippage occurs. The result is shortening and stacking of the sedimentary cover, with seemingly little deformation of underlying crystalline rocks (Fig. 10-6).

Fold-and-thrust belts commonly form within or ahead of larger zones of deformation. Examples are the Absaroka Mountains of northwest Wyoming within the Rockies, the fold-and-thrust belt of the eastern Andes, especially in Bolivia, the Valley-and-Ridge of the Appalachians, and the Calcareous Alps, a subset of the entire Alps.

Fold-and thrust belts (this section) and the uplifts along reverse faults (discussed in the previous section) are two extremes in possible response of the crust to compression. In fold-and-thrust belts, shallowly-dipping thrusts allow shortening of the upper crust, whereas uplifts bring up deep crust along much steeper faults. These differences have led geologists to speak of the former as "thin-skinned tectonics" and the latter as "thick-skinned tectonics". Both kinds of deformation can occur, and sometimes they occur close together, as in the Absarokas and Bighorns of northern Wyoming.

6. Convergence of two plates of oceanic lithosphere

As was discussed in the previous chapter, subduction of oceanic lithosphere involves dehydration of the subducting slab. The water produced by dehydration rises into the overlying rocks and induces melting of relatively siliceous components of the lower crust. The resulting magmas rise to produce island arc volcanoes.[67] We use the term "island arc" because the separate volcanoes of an arc commonly form individual islands, but they also commonly merge to make elongate larger islands studded with volcanoes (Fig. 10-7).

Examples of island arcs include Tonga, the Marianas, and the Aleutians in the Pacific, Java and the much of rest of Indonesia in the eastern Indian Ocean, and the Windward Antilles in the eastern Caribbean. While none of the volcanoes as seen above water may seem like large mountains, one should remember that their bases are far below sea level, so that much of their area and most of their volume are covered by water (Figure 10-1).

In addition to their significance as volcanic mountains on their own, island arcs are also significant to the formation of larger mountain systems because they are commonly welded onto larger continental masses in mountain-building. As subduction closes the ocean between an island arc and a continent, the island arc eventually collides with the continent and ultimately is welded onto that continent. Thus accreted terranes, elongate regions of geologically exotic material welded onto the edges of continents, are commonly old island arcs. Examples in North America include the Taconic Mountains in the northern Appalachians and many parts of western Oregon, Washington, and British Columbia in the Cordillera.

7. Convergence of oceanic lithosphere and continental lithosphere

As with subduction of oceanic lithosphere under an oceanic plate, subduction under a continental plate leads to release of water that promotes melting of the mantle and lower crust to produce magmas that rise into the upper crust. However, when ocean lithosphere is subducted under a continent, the magmas are more sialic because they are derived from the continental crust. The results are emplacement of dioritic to granitic plutons and eruption of andesitic to rhyolitic volcanics in a continental (rather than oceanic) arc (Fig. 10-7).

The present outstanding example of this mountain-forming process is the Andes. In North America, the Cascade volcanoes are a good more-or-less modern example of the upper part of the process, with the 1980 eruption of Mt. St. Helens as a modest example of the explosive eruptions possible with sialic volcanism. The granitic batholith of the Sierra Nevadas in California provides a now-exhumed Jurassic to Cretaceous example of the plutonic part of the process.

8. Convergence of two plates of continental lithosphere

Subduction of oceanic lithosphere can ultimately lead to the closing of an ocean basin, and to the collision of an arriving continent with the continent that overlies the subduction zone (Fig. 10-8). As the last oceanic lithosphere is subducted, and as some of the newly arrived under-thrusting continent is subducted or at least wedged under the overlying continent, the margins of both continents are compressed to cause folding and thrust faults. Both the horizontal compression of the collision and stacking of crust vertically lead to high pressure that promotes extensive high-grade metamorphism. The result is the largest mountain ranges on land. Examples include the Himalayas,

[67] In other words, subduction leads to orogeny.

the Alps, and the Appalachians, each of which are discussed in subsequent chapters.

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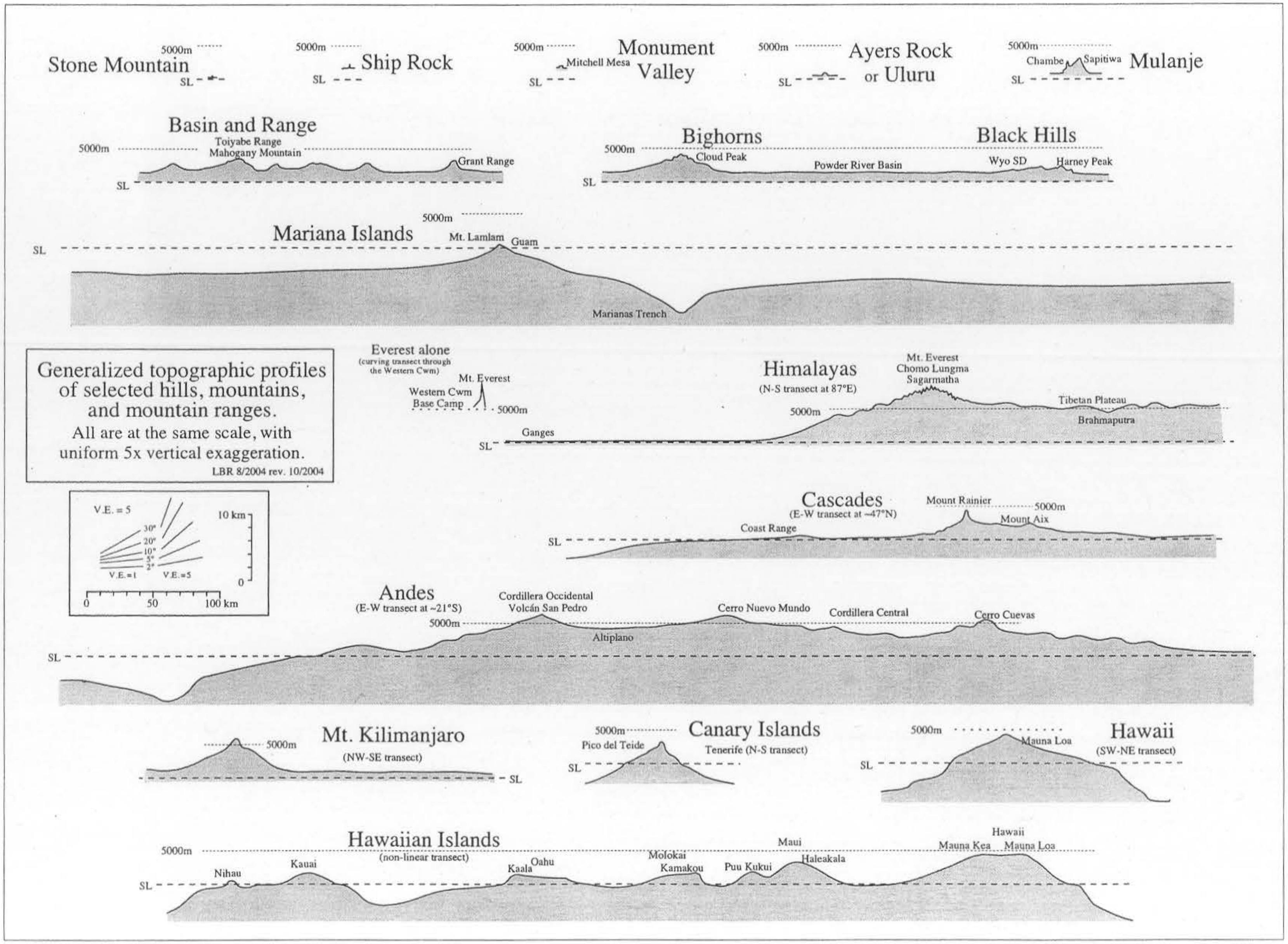


Figure 10-1: Topographic profiles of some hills, mountains, and mountain ranges.

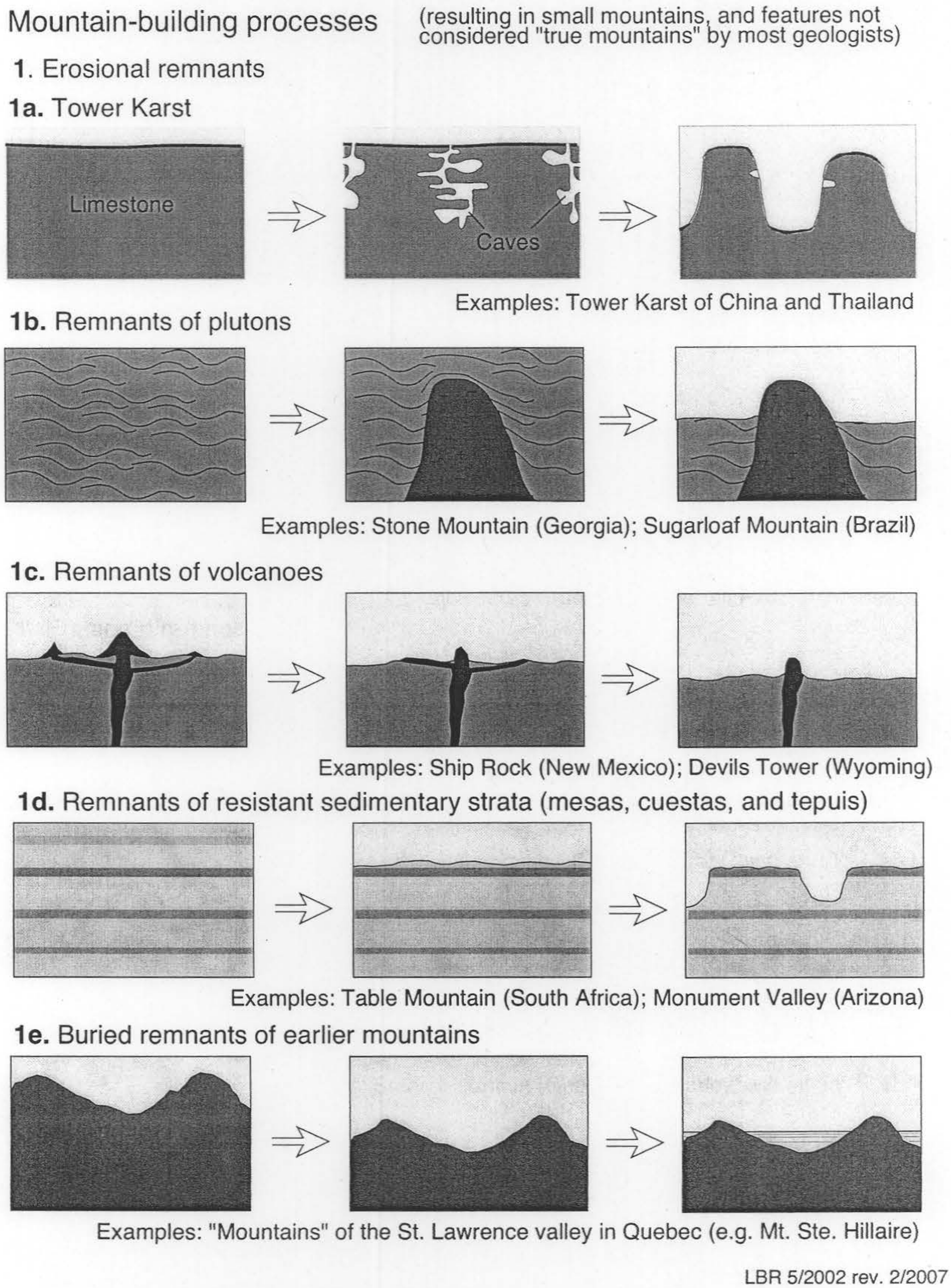
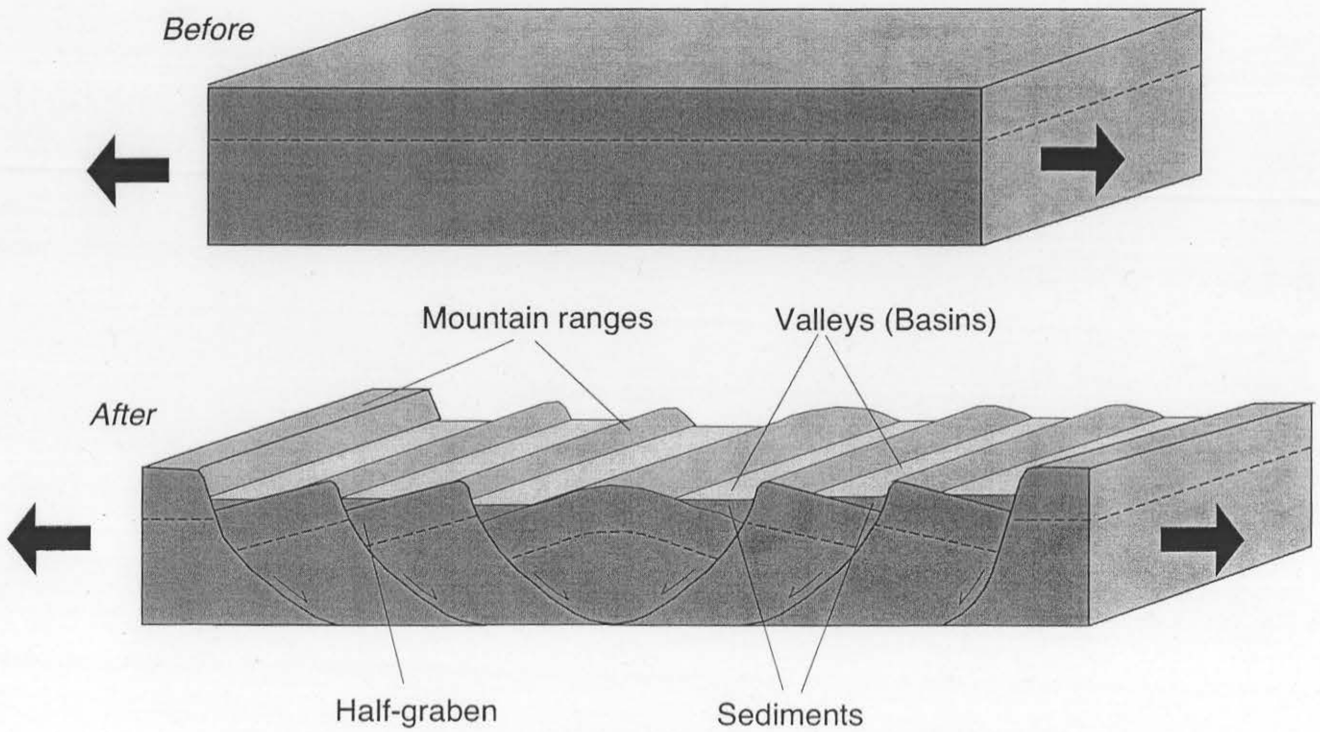


Figure 10-2: Origin of hills and small "mountains" via erosion.

Mountain-building processes

Extension and normal faulting: the half-graben model



Example: The Basin and Range province of the western United States

Based on Figure 5.8 of Moores and Twiss (1995).

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Figure 10-3: Formation of mountains via extension and normal faulting.

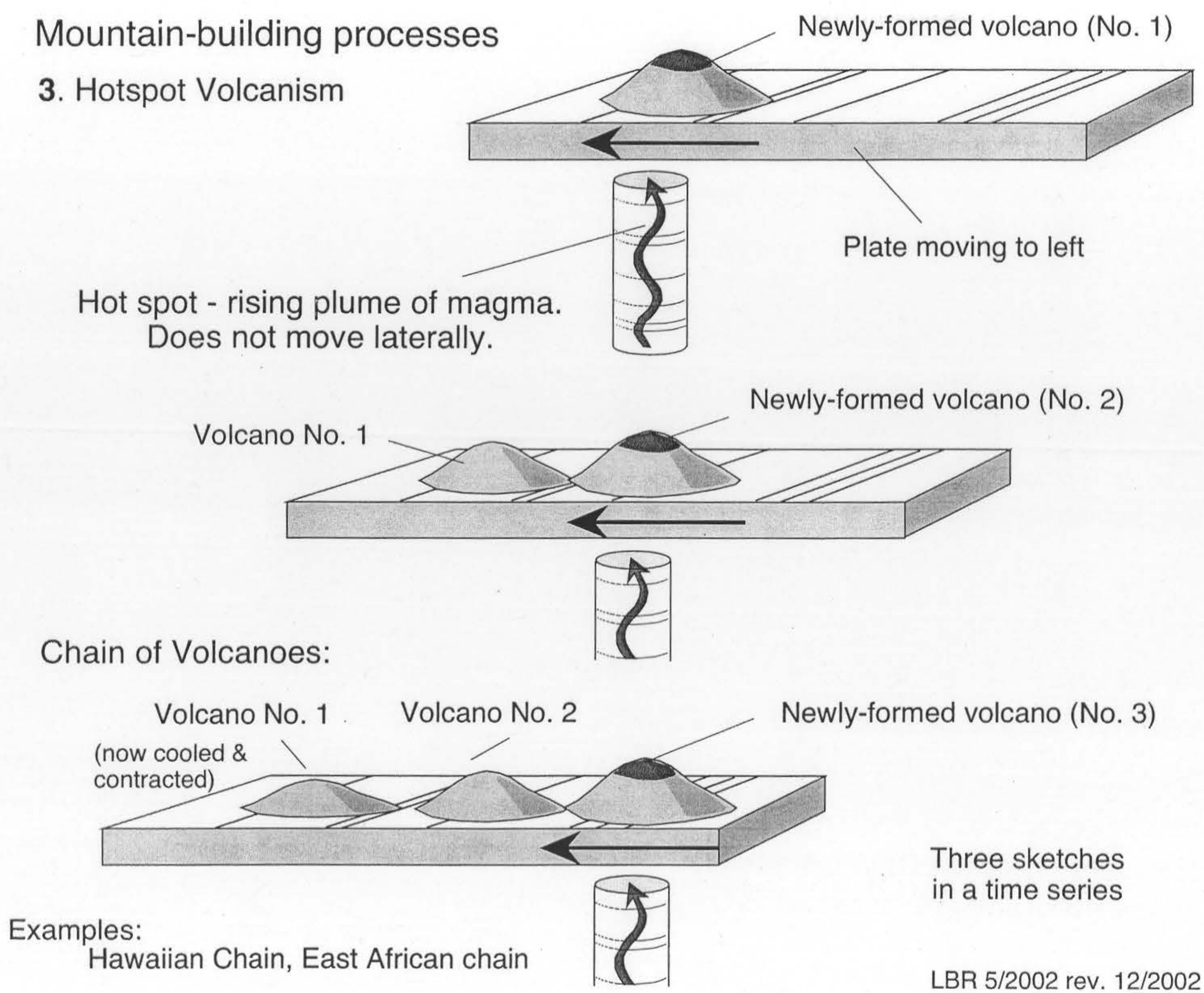
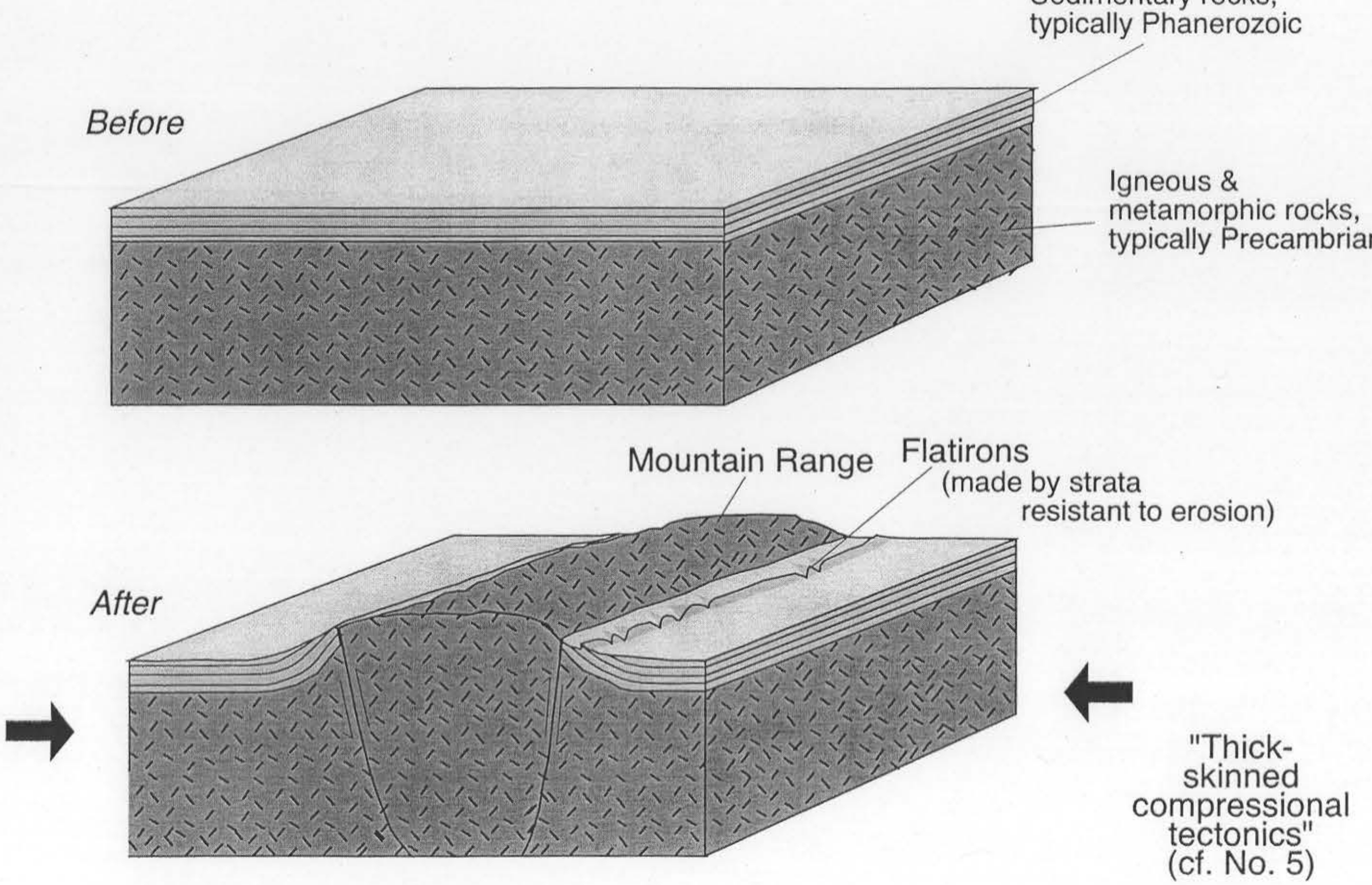


Figure 10.4: Formation of mountains by hotspot volcanism.

Mountain-building processes

4. Compression and high-angle (reverse?) faulting



Examples: Bighorns, Black Hills, Rockies Front Range?

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Figure 10.5: Formation of mountains by compression and thrusting.

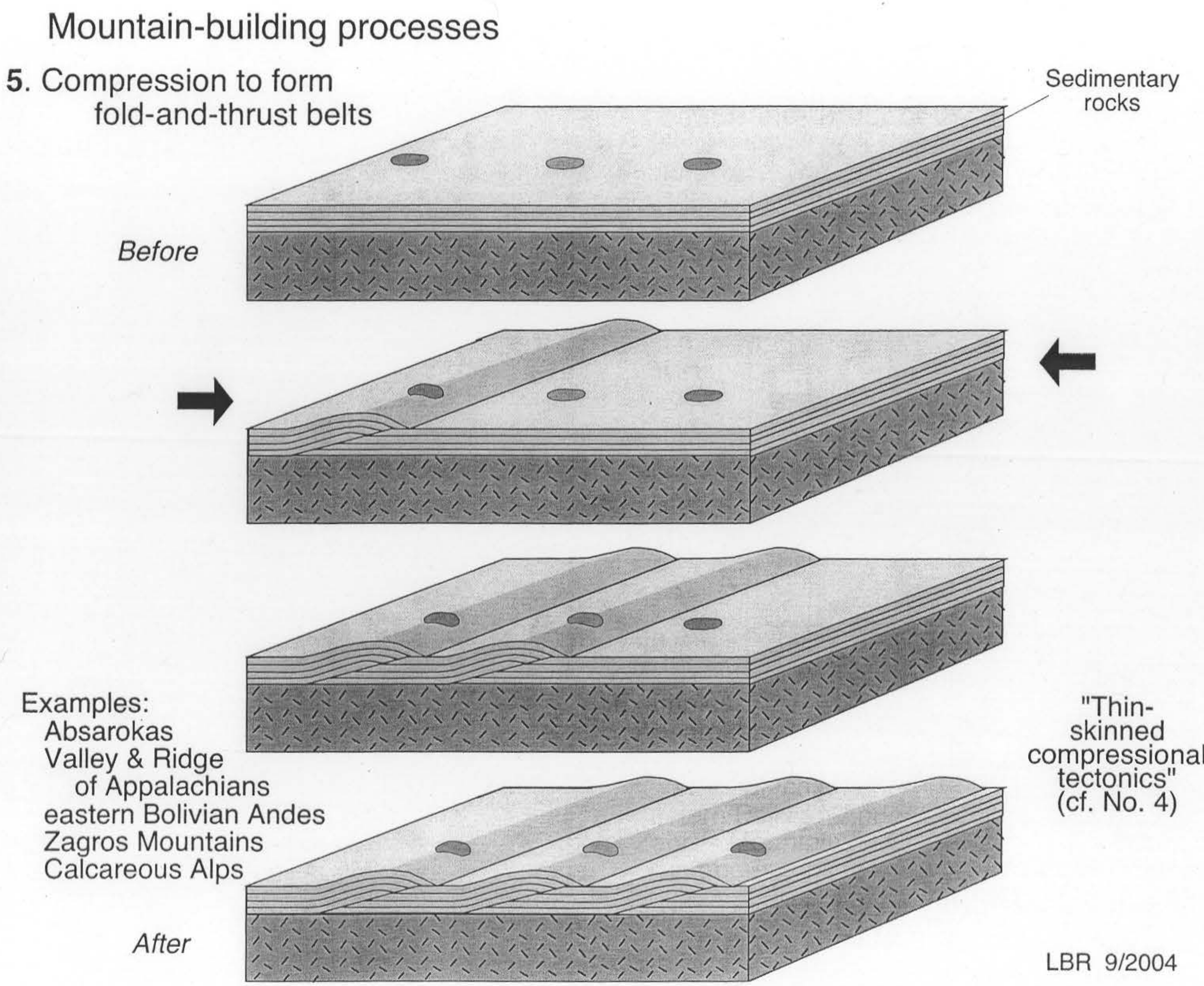


Figure 10.6: Formation of fold-and-thrust belts.

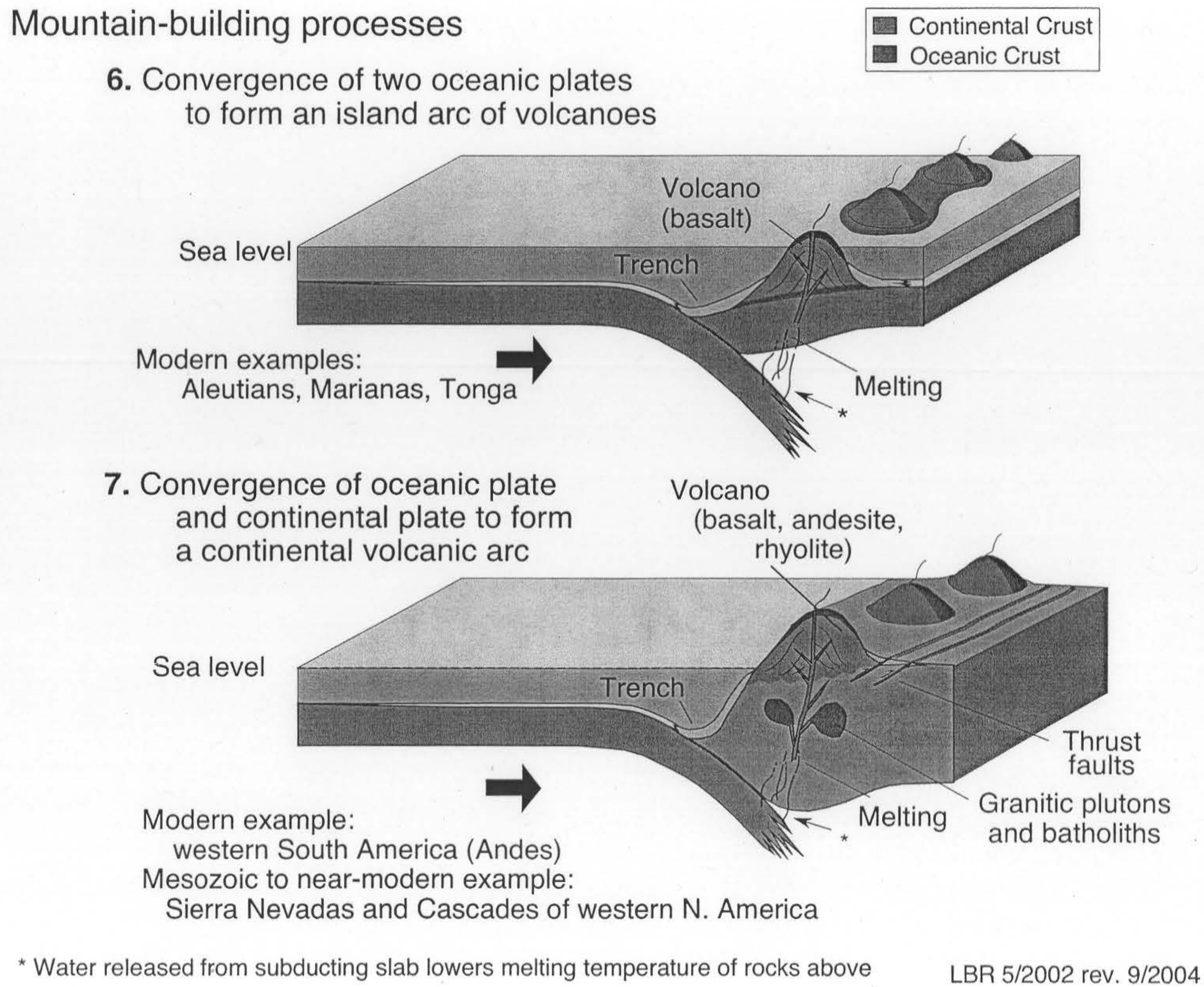
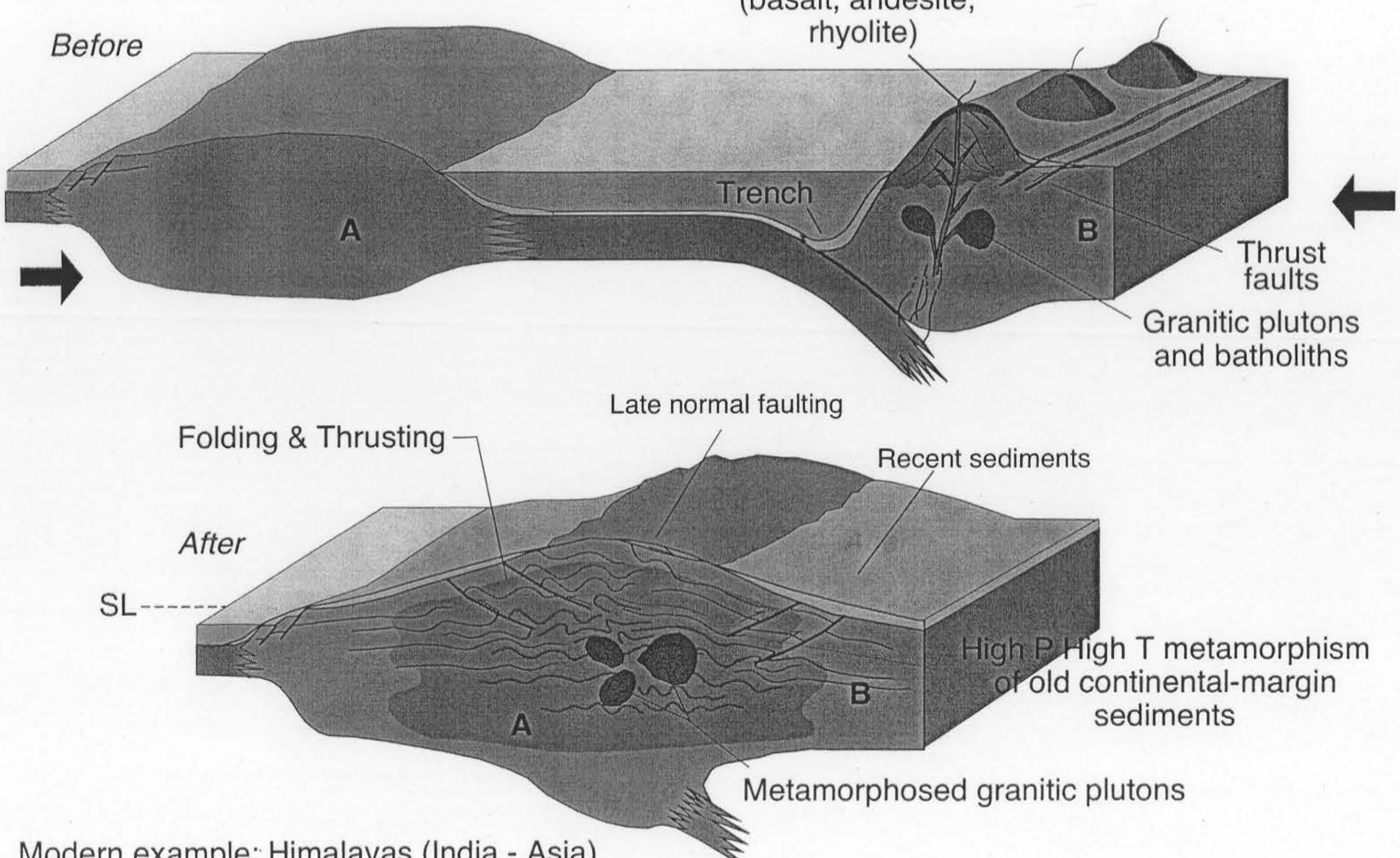


Figure 10.7: Formation of mountains associated with subduction of oceanic lithosphere.

Mountain-building processes

8. Convergence of two continental plates
(continent-continent collision)



Modern example: Himalayas (India - Asia)
Tertiary example: Alps (Europe - AfroMediterr'n)
Pennsylvanian example: Appalachians (Laurentia - Gondwana)

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Figure 10.8: Formation of mountains in continent-continent collision.

CHAPTER 11: THE TIROLEAN ALPS

The Alps[68] extend in an arc from southeastern France through Switzerland to eastern Austria[69]. Along their length, their geology varies considerably. This chapter therefore does not attempt to characterize the Alps along their entire length but instead focuses of the Alps of Tirol, both the present Austrian land of Tirol and South Tirol, which is today in northern Italy (Figs. 11-1 and 11-2). The chapter first describes the geology of the region and then attempts to provide a historical explanation.

Zonation

The Alps can be divided into three major zones, the Calcareous Alps, the Metamorphic Alps, and the Dolomites. They are framed to the north by the German Foreland Plain and to the south by the volcanic geology of central Italy (Fig. 11-3).

The **German Foreland Plain** is a flat expanse of sediments and sedimentary rocks draining northward toward the Danube. Munich is its most prominent city. The plain is so flat because sands and gravels shed off the front of the Alps and carried northward by rivers and glaciers have buried any pre-existing topography.

The sediments shed off the Alps are conventionally divided into two groups, flysch and molasse. Flysch is syn-orogenic sediment eroded off a growing mountain mass and deposited in front of the advancing orogen. It is thus typically deposited in the seaway in front of an orogen, and it subsequently undergoes deformation, mostly intense folding, as the orogen advances through that seaway. Molasse, on the other hand, is post-orogenic sediment eroded after the orogeny is finished, and it typically is deposited on a foreland plain and thus in non-marine conditions, and it does not undergo subsequent deformation. Most of the German foreland plain as seen today is thus molasse, and flysch can only be seen in outcrops near the northern margin of the Alps.[70]

The **Calcareous Alps**, or "Kalkalpen" in German, or Limestone Alps, make up the Alps north of the Inn Valley[71]. They consist of Triassic to Jurassic limestones (and minor shales) with marine fossils that document that these rocks were deposited on the seafloor. These rock layers have been so extensively thrust faulted that the same strata are commonly repeated three times vertically at any one location. Normal and strike-slip faults further disrupt these rocks, yielding a very rugged topography and a landscape across which few roads cut, even though the Calcareous Alps are not as high as the Metamorphic Alps to the south. Most of the strike-slip faults are left-lateral faults.

The Inn Valley is the boundary between the Calcareous Alps and the Metamorphic Alps at Innsbruck.[72] However, to the east the Inn Valley curves northward, and the Calcareous Alps lie on both the north and south sides of the Inn. Thus, although the Inn Valley is the path of an underlying fault, it is fault that cuts across the regional geology, and thus its separation of the Calcareous Alps from the Metamorphic Alps in Innsbruck is a coincidence. The fault in question is a left-lateral fault, like many of the strike-slip faults in the northern Alps.

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[68] The name "Alps" comes from the French word for a high mountain pasture (e.g., the Alp d'Huez of cycling fame). The corresponding term in Austrian German is "alm", likewise a pasture high above a town (e.g. Arzler Alm and Höttinger Alm above Arzl and Hötting near Innsbruck).

[69] The term "Alps" has the more general meaning of an impressive mountain range, thanks to epxort of the term to locations around the world. Thus there are the Lyngen Alps of northern Norway, the Arrocher Alps of western Scotland, the Japanese Alps on Japan's island of Honshu, the Australian Alps in southeastern Australia, the Trinity Alps in northern California, and most notably the Southern Alps of New Zealand. The size and signficance of the latter make it critical that one keep track of one's capitalization of "Southern Alps " (the range in New Zealand) and "southern Alps" (the southern part of the European original).

[70] "Flysch" and "molasses" are terms use in the Alps by European geologists. Attempts by other geologists elsewhere usually are rejected by European geologists as inappropriate applications of concepts perhaps unique to the Alps.

[71] The Calcareous Alps can be further divided into the Lechtaler Alpen and Allgäuer Alpen in western Tirol, the Wetterstein Gebirge (Wetterstein mountain range) west of Seefeld and Mittenwald, the Karwendel Gebirge north of Innsbruck, and the Rofan farther east. In Innsbruck one looks up to the Haflekarspitze, the peak of the Haflekar mountain, which is part the Nordkette ("north chain"), the ridge on the north side of the Inn Valley, and the Nordkette is the southernmost ridge of the Karwendel mountains, which are part of the Calcareous Alps.

[72] For more on the Inn Valley itself, see the next chapter.

The **Metamorphic Alps**[73] lie to the south of the Inn Valley. Although they rise more gradually than the Calcareous Alps, they ultimately rise to the greatest heights in the Alps. They consist of metamorphic rocks but differ on the two sides of the Sill (a difference that will become profound further along). East of the Sill, quartzphyllites and otherwise purely silicate metamorphic rocks dominate. West of the Sill, calcsilicates representing a mixture of metamorphosed limestones and silicate rocks are dominant. The calcareous nature of the rocks to the west leads to a steeper topography than that to the east, as a viewer looking south from Innsbruck can see by comparing the ruggedness of Serles with the rounded top of the Patscherkofel.

High in the Metamorphic Alps, and thus generally near the Austrian-Italian border, are areas of high-grade metamorphic rocks, such as granulite gneisses, amphibolites, and garnet-bearing and amphibole-bearing schists, as well as areas of granite and nearly granitic rocks. These have been called the Tauern Fenster or Tauern "windows" because they seemingly provide a window down through which one sees the core of the Alps. They have also been called "Altkristallin" areas because they were assumed to be of greater age to have achieved such intense metamorphism. The latter idea is probably not sound, but these areas certainly expose spectacular lithologies that set them apart from the more bland schists, phyllites, and meta-sedimentary rocks of the rest of the Metamorphic Alps.

To the south of the Metamorphic Alps are **the Dolomites**, rugged mountains in South Tirol. The Dolomites, as their name implies, consist of dolomitized limestones, and in this case the original limestones were Triassic to Jurassic in age, the same as those of the Calcareous Alps. The Dolomites are also like the Calcareous Alps in having a spectacularly rugged topography with dizzying vertical surfaces. Below those cliffs one commonly sees flat-bottomed valleys floored by volcanic rocks that mark the beginnings of Italian volcanic geology and the end of the Alps. Farther south is also the Po River plain, which consists largely of sediments derived from the Alps.

History, Part I: A local view

Almost all accounts of the origins of the Alps involve closing of the Tethys Seaway, an east-west seaway extending from the Pacific westward between Asia and India and then between north-central Europe to the north and Italy, Africa, and the Arabian Peninsula to the south. This seaway was closed as Africa and its associated lands moved northward and, about 50 million years ago, began the collision that would build the Alps.

Figures 11-4 to 11-6 provide one scenario for construction of the Tirolean Alps. Parts 1 to 4 show the closing of the Tethys Seaway under the northwards-advancing Apulian (or African-Italian) plate. Parts 5 to 7 show the development of the Alps with faulting along two thrusts; a more complete account would inevitably involve more such thrust faults. The result is nonetheless the necessary three-fold stacking of strata in the Calcareous Alps and uplift of deeply metamorphosed rocks in the heights of the metamorphic Alps, and exposure of the latter as windows with erosion of the entire system. The building of the Alps has long since ceased, but minor earthquakes still are felt as rebound occurs in response to erosion and as minor adjustments occur along old faults.

History, Part II: The bigger picture

The Australian geologist David Ford once wrote, "Fifty years ago, my introduction to Alpine geology was Collet (1927), a Swiss, on the western Alps, and Heritsch (1929), an Austrian, on the eastern Alps. The two accounts were so different that the two halves of the Alps could have been on different planets". That perplexing dichotomy persisted through the 20th century. However, recent seismic research seems to have explained part of the problem.

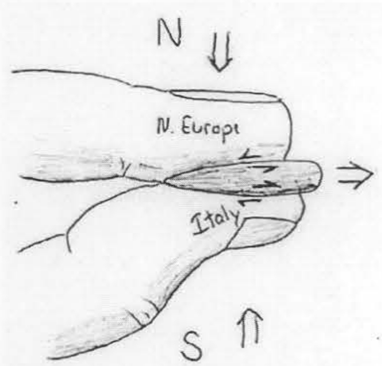
The conventional understanding of the plate tectonic history of the Alps was that the southern margin of the pre-Alps European continent was subducted under the plate

[73] "Calcareous Alps" (or "Kalkalpen") is a standard term used by all geologists and by lay persons. "Metamorphic Alps" is an LBR invention and could readily be criticized either because the area in question contains some rocks that are barely metamorphic (they're metasediments at most) or because it contains many different geologic areas that arguably should not be lumped into one region. "Zentralalpen" is a term used by some, but it runs into the confusion of whether "central" is in the sense of east to west, or north to south. A Tirolean would probably refer to the Metamorphic Alps as the Ötztaler Alpen, the Stubaier Alpen, die Tuxer Voralpen, die Zillertaller Alpen, usw.

approaching from the south (the Apulian Plate). Thus the subducted slab dipped to the south. Imaging of the crust and mantle confirms that view for Switzerland and far western Austria (the western Alps). However, seismic imaging in the eastern Alps indicates that most recently the subduction zone has dipped to the north, with the plate approaching from the south subducted under the European plate. This change through time in the eastern Alps seems to have happened as the subducting European plate foundered and rolled back to the Carpathians, and the already subducted part of the plate broke away into the mantle (Fig. 11-7). This break allowed the northern edge of the Apulian plate to sink, beginning its subduction northward. The boundary between the western region, in which the European plate maintained its integrity and direction of subduction, and the eastern region, in which the European plate foundered and subduction changed, coincides with Innsbruck and the Sill Valley.

If the previous paragraph is concerned with vertical movements, the foundering of the European plate north of the eastern Alps also set in motion horizontal movements. The new availability of surface space to the northeast of the Alps allowed much of the eastern Alps to be pushed east-northeast by the north-south horizontal compression that in general built the Alps. As an analogy, consider holding a watermelon seed between the thumb and forefinger of your left hand, as you face north. Your thumb and forefinger represent the two plates pressing north-south against each other to form the Alps. The vacuity beyond the tips of your thumb and forefinger is the region south of the Carpathians where the European plate foundered, and the Black Sea. As you press your thumb and forefinger, the watermelon seed squeezes out to the east. The interface between your forefinger and the watermelon seed is a left-lateral strike-slip fault (like the Inn Valley's Inntal Fault), and the interface between the seed and your thumb is a right lateral fault like the Pustertal Fault in the Periadriatic Line in southeastern Tirol (Fig. 11-8).

The geological result has been a major extrusion of crust eastward between the left-lateral faults in the northern part of the Eastern Alps and right-lateral faults in the southern part of the Eastern Alps.[74] This sense of motion along faults can be traced into eastern Switzerland, at least along the Engadine Line (a fault in the upper valley of the Inn). However, for the most part the extrusion of crust eastward is a feature of the Eastern Alps (the Alps east of Innsbruck) (Fig. 11-9).[75] Thus we arrive at two major differences between the Western and Eastern Alps: the deep-seated difference in the direction of subduction over the last twenty million years, and the nearer-surface development of the extrusion or escape tectonics along strike-slip faults in the Eastern Alps during more or less the same period. Innsbruck and the Sill Valley emerge as the boundary between these two regimes, and thus a good, if occasionally perplexing, base from which to examine the Alps.



The watermelon-seed model of the extrusion tectonics of the Eastern Alps.

[74] Thus the watermelon seed of the previous paragraph is in reality several slivers divided by several faults, rather than one entity bounded by just two faults. However, the northern faults all mimic the fore-finger-seed relative motion, and the southern faults all mimic the seed-thumb relative motion.

[75] Again, the watermelon-seed model is a simplification, in that there are two seeds squirting eastwards, the smaller being the core of the Eastern Alps and the larger being the region between the Adriatic and the Danube. See Figure 11-9.

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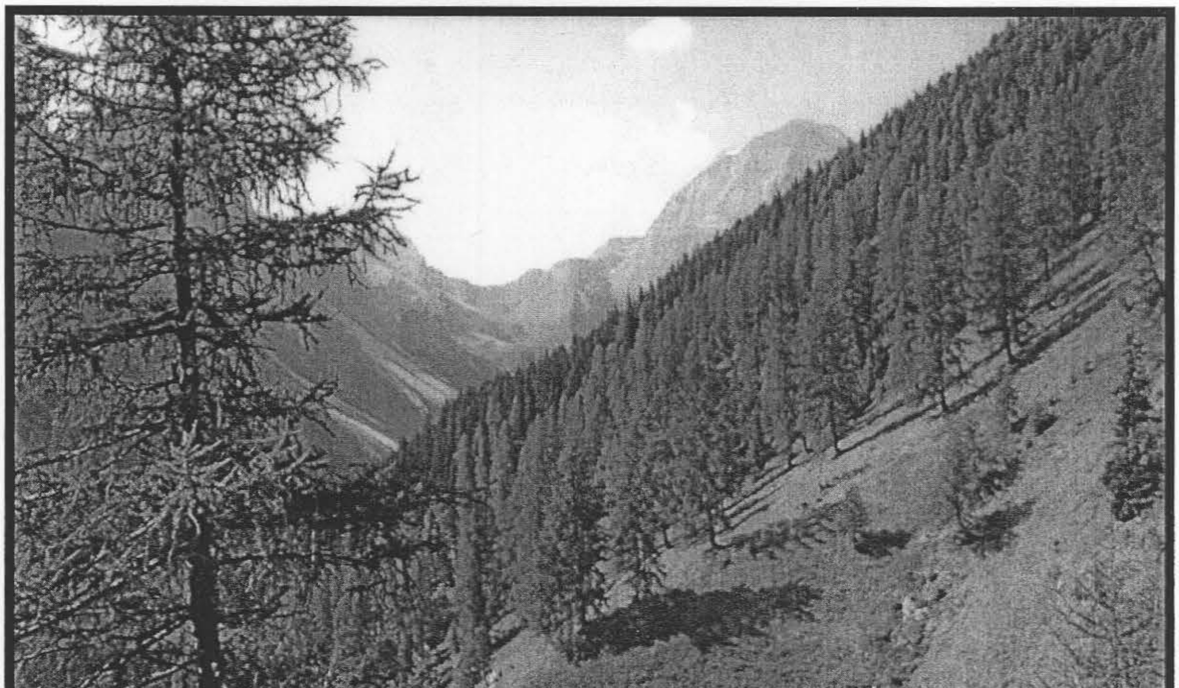


A view up the Pinnistal, a tributary of the Stubaital, in the metamorphic Alps southeast of Innsbruck

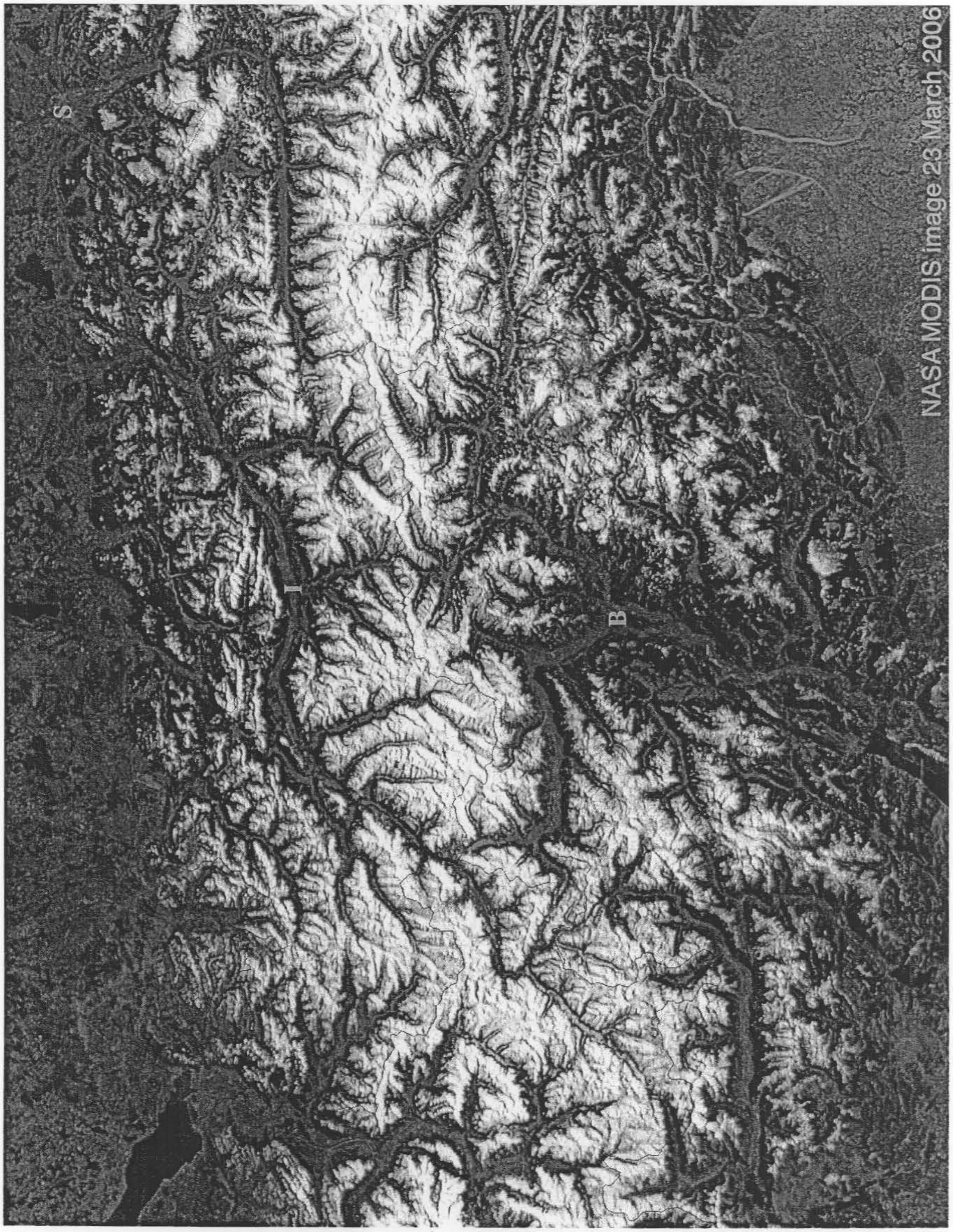


Figure 11-1: A satellite image of the Tirolean Alps. I = Innsbruck, S = Salzburg, B = Bolzano.

Next page:
Figure 11-2: A geologic map of the Tirolean Alps,
from München (Munich) to Bozen (Bolzano).
This is a small portion of Vetters, H., 1923,
Geologisiche Karte der Republik Österreich
(Geologischen Bundesanstalt, Vienna), reprinted 1968.

Highly simplified key, from north to south:

White:	Quaternary sediments
Yellow:	Cenozoic terrestrial sediments & sedimentary rocks
Red line:	north limit of end moraines
Green:	Tertiary to Cretaceous sediments (largely flysch)
Light and dark blue:	Jurassic limestones of the Calcareous Alps
Pinkish brown:	Triassic limestones of the Calcareous Alps
Light brown east of Sill:	Innsbrucker Quartzphyllite
Orange east (and west) of Sill (γg):	Gneisses
Dark blue rimming orange:	Limestones & metamorphic equivalents
Peach west of Sill:	Metasediments
Brown with vertical lines:	Calc-phylllites
Red-dotted brown northwest of Meran:	Micaceous schist and gneiss
White with blue curving lines:	Glacier
Red between Sterzing and Brixen:	Granite and tonalite
Light brown near Brixen (Bressanone):	Quartz phyllite
Bright orange near Bozen (Bolzano):	Andesites and trachytes
Purples east of Bozen:	Basalts and diabases
Pinkish brown east and west of Bozen:	Dolomitized Triassic limestones
Light & dark blue east and west of Bozen:	Dolomitized Jurassic limestones

A high-resolution (2316 x 3149 pixels) color jpeg file of Figure 11-2 is available on the World-Wide Web at www.gly.uga.edu/railsback/AG/AlpsMapFourPlexCropped2.jpg. When printed at 330 dpi, the file yields at 7.5 inch x 9.5 inch print that fits on the opposite page. Also available on the Web is an even larger (3448 x 5528) jpeg image of the same map at the same resolution, but covering a larger area, at www.gly.uga.edu/railsback/AG/AlpsMapFourPlex.jpg. Its northern extent includes Munich, its eastern extent includes Traunstein, KitzBühel, the Grossglockener, and Venice, its southern extent includes Venice, Padua, and Verona, Padua, and its western extent includes Brecia, the Tirol-Vorarlberg border, and Memmingen,


MÜNCHEN
Pasing
Planegg
Landsberg
Kaufbeuren
Weilheim
Schongau
Rosenheim
Kufstein
Walchensee
Garmisch
Partenkirchen
Mittenwald
Wetterstein Gebirge
Schwaz
Hall
Landeck
Matrei
Sterzing
Brixen
Bozen
Mt. Cristallo
Cortina d'Ampezzo

Division of the Alps through Tirol and surrounding regions:

1. German Foreland Plain:

   Area of sediments shed northward off the Alps

   Flysch: Folded synorogenic sediment
   Molasse: Undeformed post-orogenic sediment

2. Calcareous Alps (Limestone Alps) - north of Inn Valley*

   Triassic to Jurassic limestones and shales
   Marine sediments from basinal to reefal

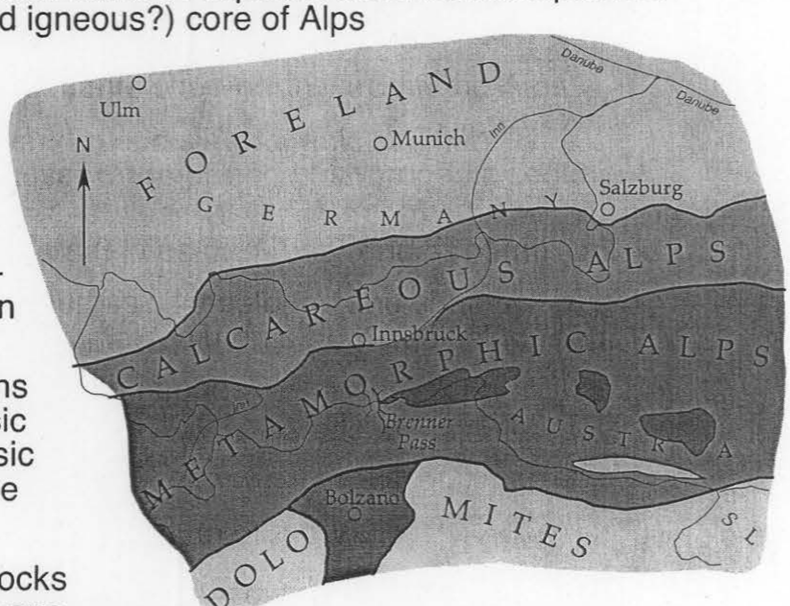
   Folded into nappes
   Thrust-faulted
   Faulted vertically

   * near Innsbruck

3. Metamorphic Alps - south of Inn Valley*

   East of Sill: Paleozoic quartzphyllite and graywacke
   West of Sill: Paleozoic calcareous metamorphics, and schists and gneisses

   Tauern window: exposes most metamorphosed (and igneous?) core of Alps



4. Dolomites - northern Italy

   Mountains of Triassic to Jurassic dolostone

(5). Volcanic rocks near Bolzano and southward

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Figure 11-3: Basics of the Tirolean Alps.

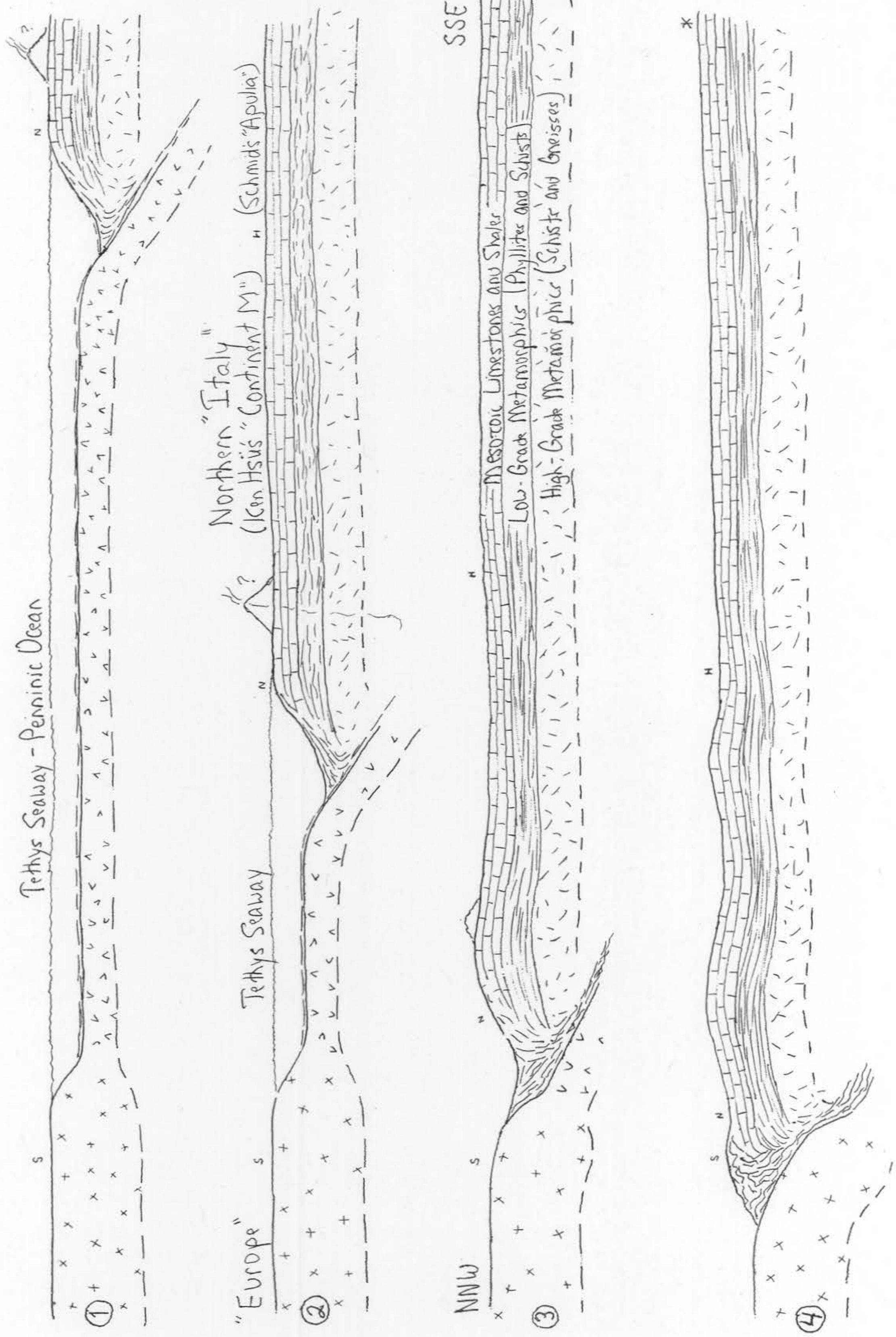

Figure 11-4: The Origin of the Tirolean Alps, Part 1.

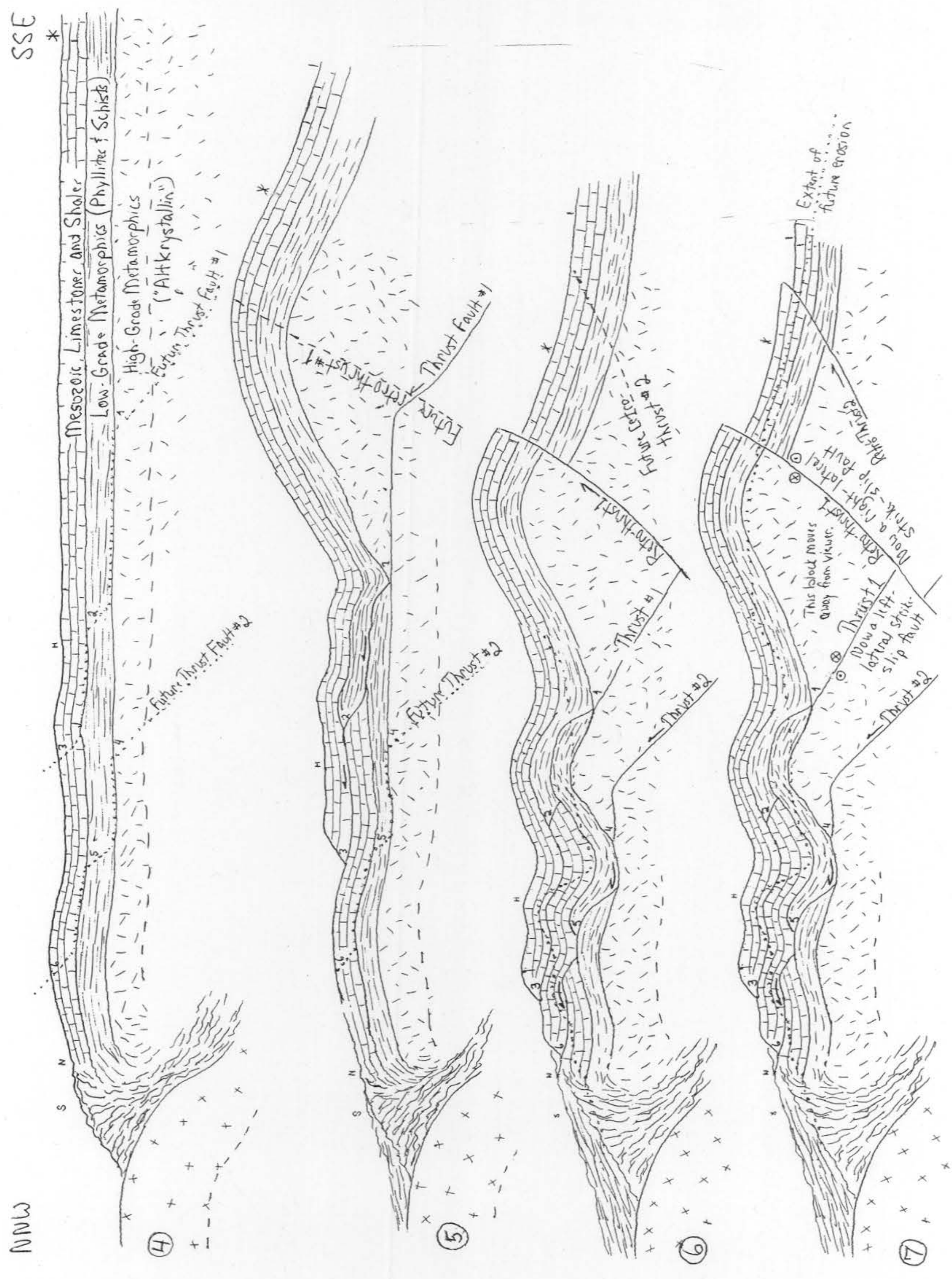

Figure 11-5: The Origin of the Tirolean Alps, Part 2.

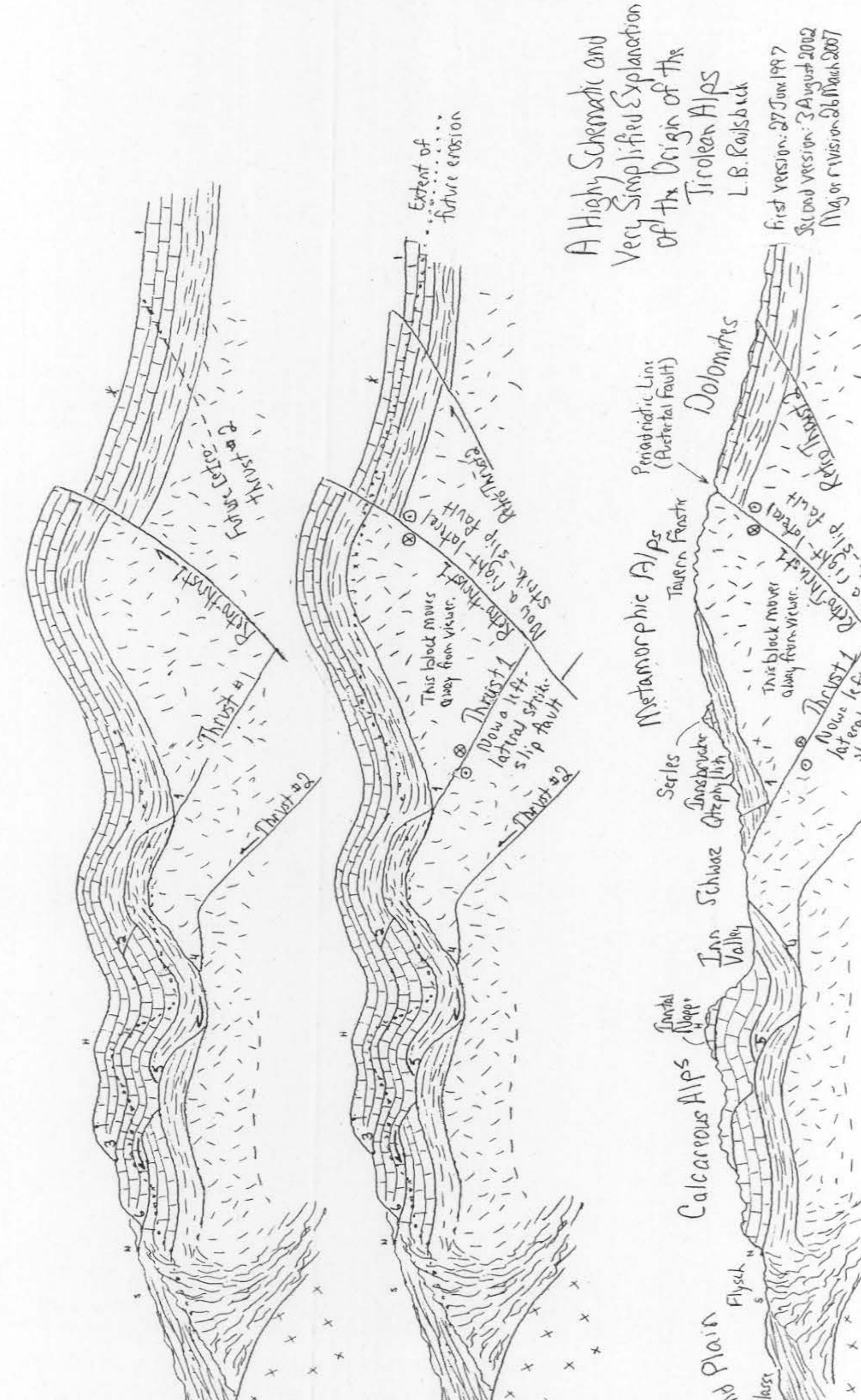

Figure 11-6: The Origin of the Tirolean Alps, Part 3.

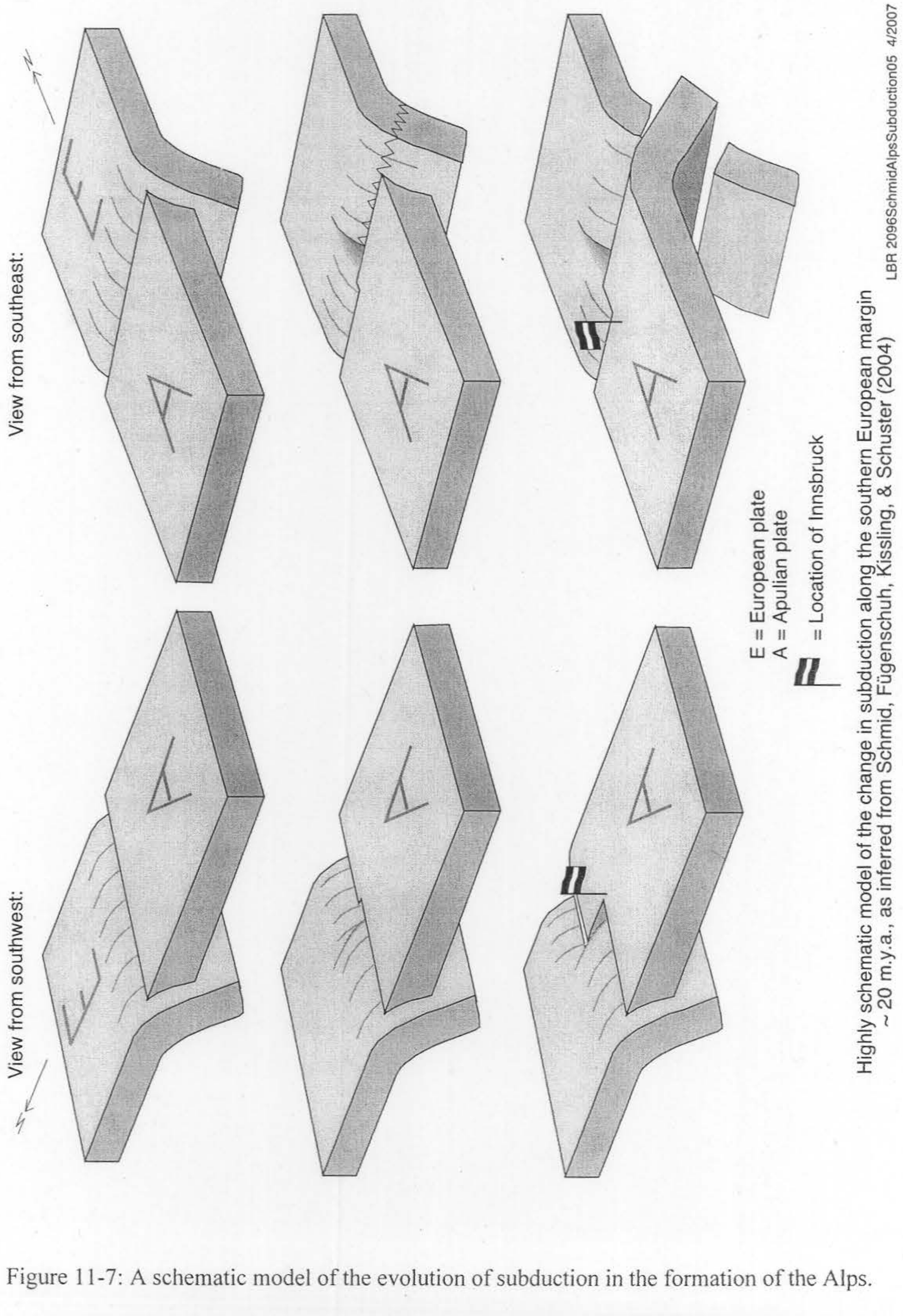

Figure 11-7: A schematic model of the evolution of subduction in the formation of the Alps.

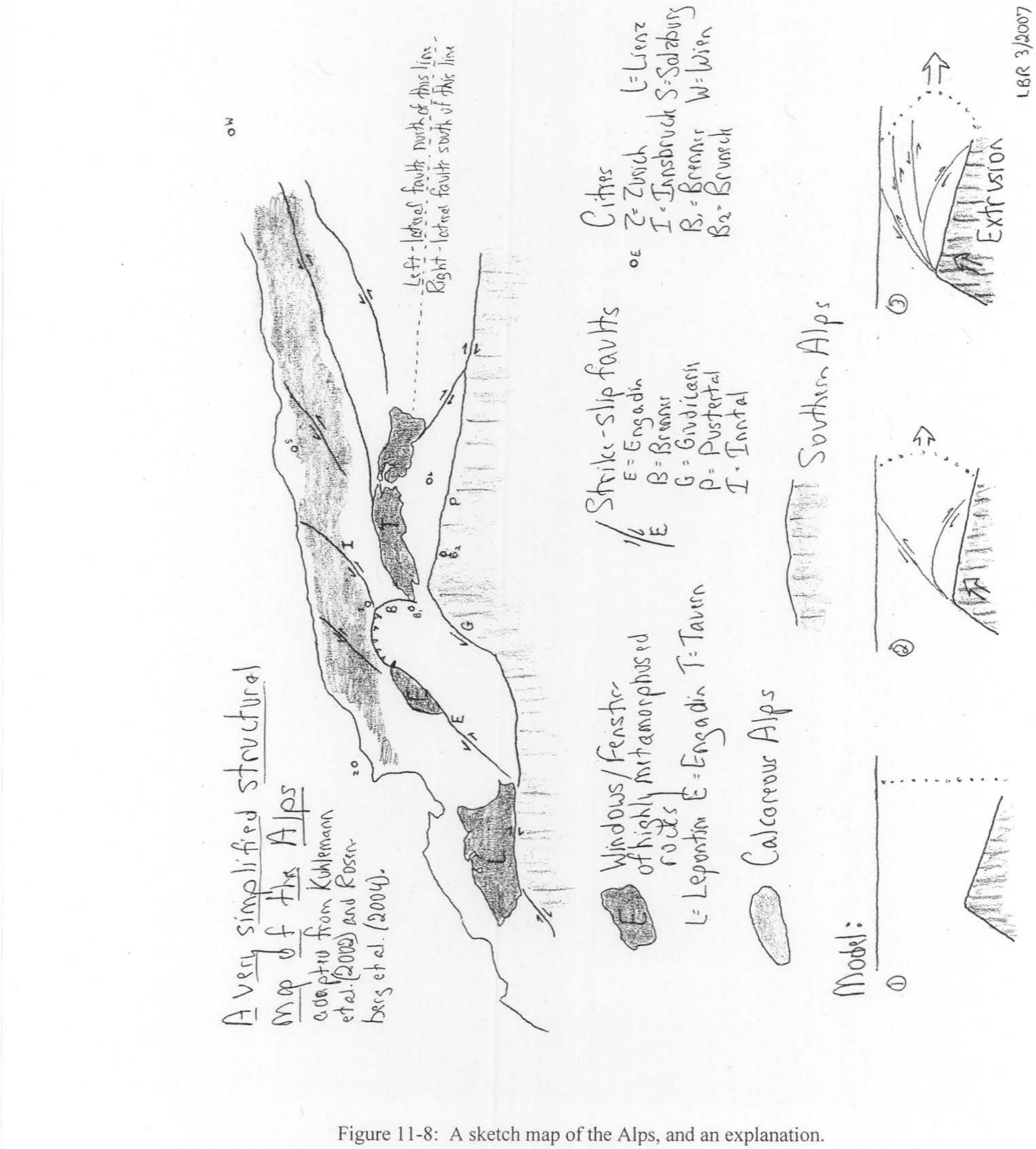

Figure 11-8: A sketch map of the Alps, and an explanation.

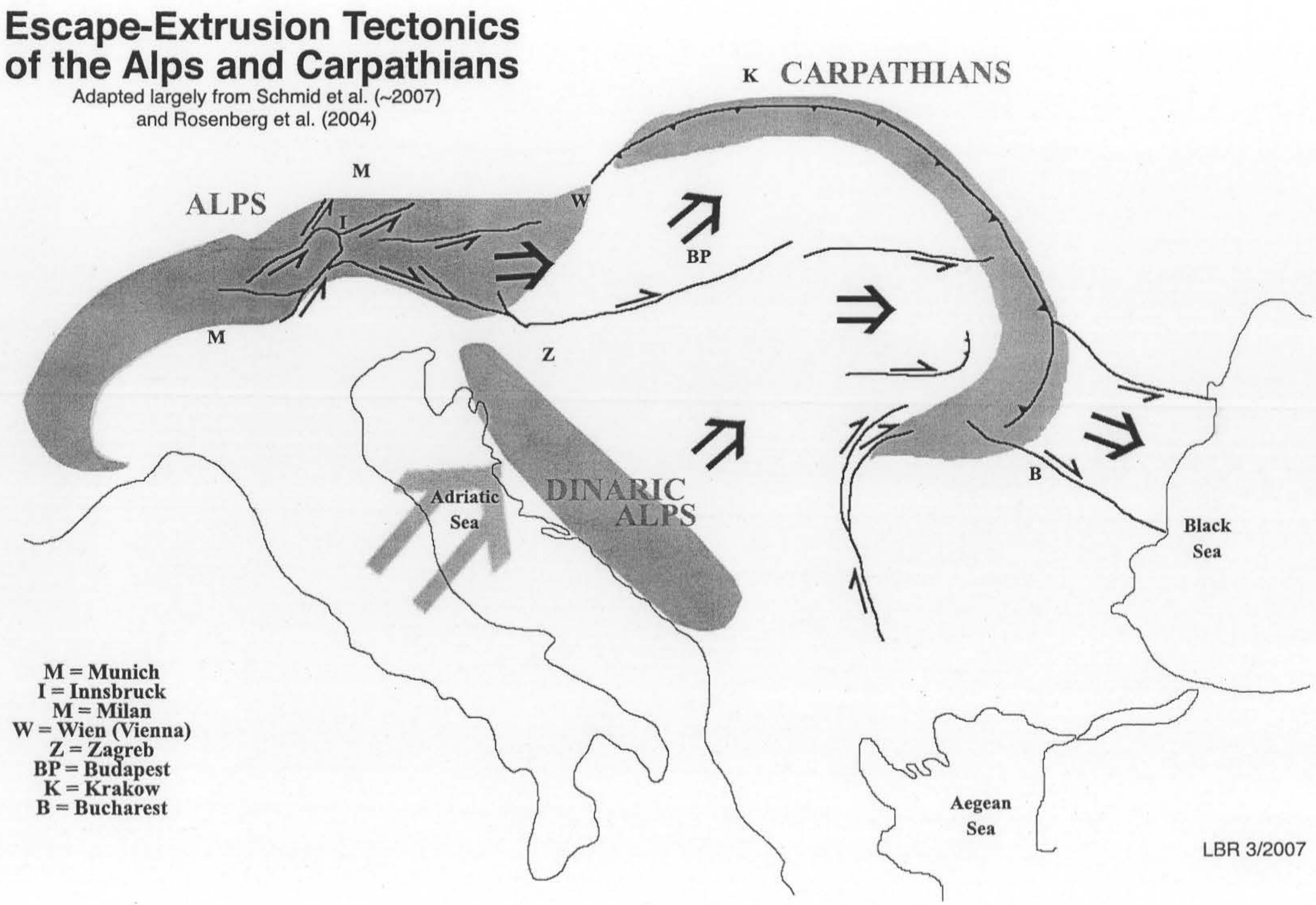

Fig 11-9: A highly generalized tectonic map of the Alps, Carpathians, and Hungarian Plain.

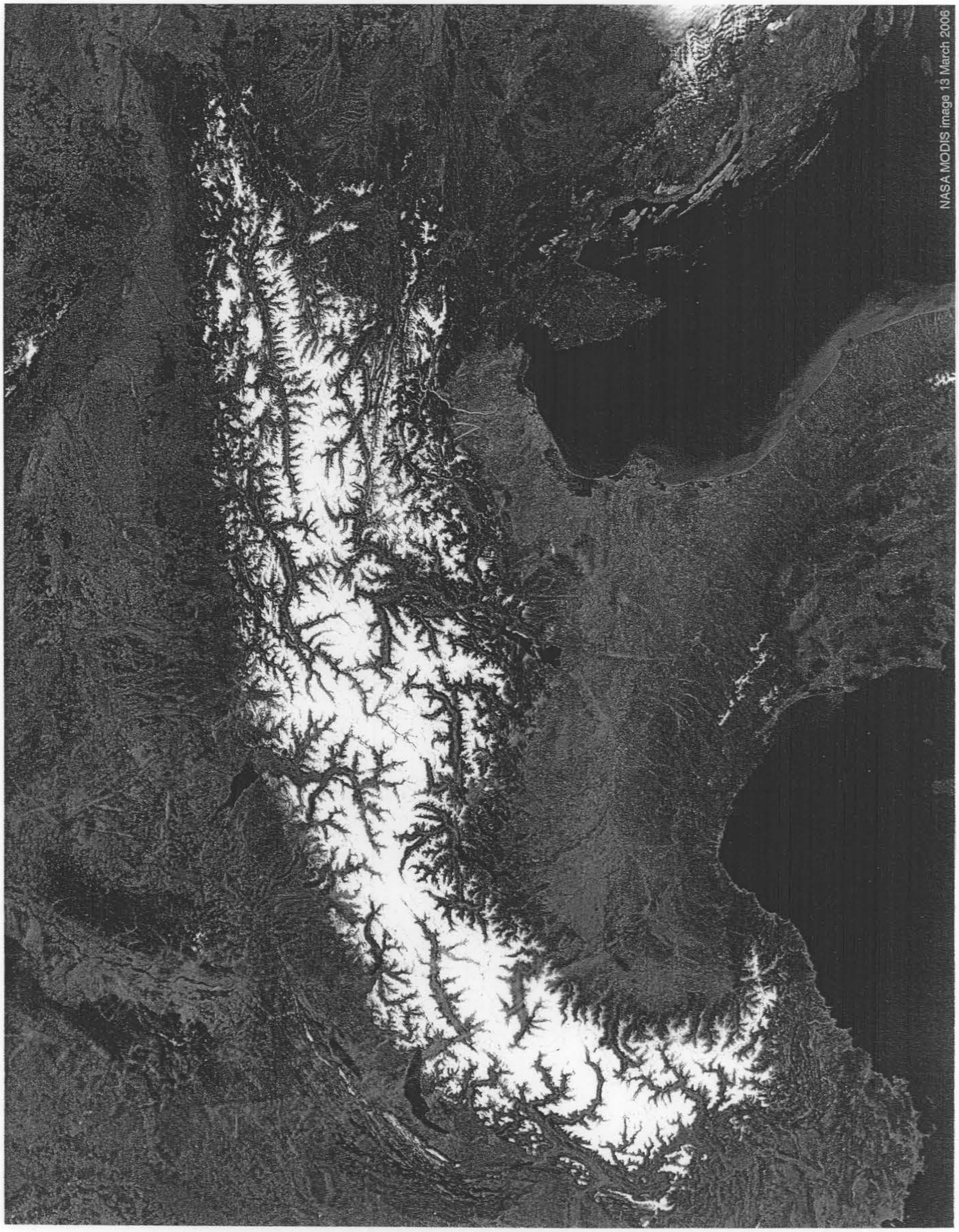

Fig 11-10: A satellite image of the Alps. See also Figure 11-1.

Chapter 12: The Inn Valley

The Inn Valley (das Inntal) is a major geologic boundary, a critical European thoroughfare, and the site of much of Tirol's economic life. This chapter briefly describes the Inn Valley from each of these perspectives, with its focus on the Inn Valley in Tirol and in the vicinity of Innsbruck.

The Inn Valley, geographically

The headwaters of the Inn River are below Maloja Pass in southeastern Switzerland, and near the famed resort of San Moritz (Fig. 12-1). There the river is called the En, and the valley is the Engadin, or the "Garden of the Inn". The river flows northeastward from there into Austria. At Landeck, in western Tirol, the Inn is joined by the Trisanna and Rosanna.[76] Thereafter, the Inn flows generally east-northeastward, passing by or through Imst, Telfs, Zirl, Innsbruck, Hall, Jenbach, Wörgl, and Kufstein before it turns north and leaves Tirol and Austria. Its waters flow into the Danube at Passau, east of Munich and well west of Vienna, and so they ultimately flow into the Black Sea.

The Inn River increases in volume of flow along this path as tributaries draining the Alps join it. The Inn Valley likewise widens, with hardly any valley floor in its upper reaches to a valley with a broad floodplain down-river from Innsbruck.

The Inn Valley, geologically

Near Innsbruck, the Inn Valley is a major geological boundary between the Calcareous Alps (die Kalkalpen) to the north and the metamorphic Alps to the south. However, its origin probably has more to do with east-west geologic trends, in that the Inn Valley near Innsbruck coincides with one of several left-lateral strike-slip faults. These faults collectively have acted as escape structures allowing slipping of blocks of the Alps eastward. Many mountain valleys, especially relatively straight ones, coincide with faults and commonly result from fracturing of rock to allow more extensive erosion. The Inn Valley seems to follow that generalization.

Flat-bottomed mountain valleys are almost always floored with sediments that have filled what would otherwise be a deeper V-shaped, or perhaps U-shaped, valley. Again, the Inn is no exception, in that drilling and seismic work show that hundreds of meters of sediments underlie the floodplain of the Inn (Figs. 12-2 and 12-3). These are sediments derived from erosion of the surrounding mountains and deposited by a river unable to transport all that sediment to its ultimate point of deposition in the Black Sea. These sediments contain a wealth of groundwater, and the uppermost sediments are the substrates of the soil on which almost all of Tirol's row crops are grown.

Above the valley floor, and flanking it, are terraces or benches of Quaternary-age sediments (Fig. 12-3). Near Innsbruck, the villages of Hungerburg and Arzl sit on these benches. These benches consist largely of two kinds of sediment. One is the breccias deposited by landslides carrying rock and soil off the mountainsides, and the breccia below Hungerburg is an excellent example.[77] The other is lake sediments deposited behind ice dams that periodically blocked the Inn Valley. These range in texture from clays like those excavated in pits at Arzl and Baumkirchen to sands and layers of gravel exposed above Arzl and near Ranggen. Groundwater moves through these terrace sediments, and it is the water tapped by the fountains found in almost every valley town.

Finally, the Inn Valley is the site of the alluvial fans of the streams flowing into the Inn. These alluvial fans provide slightly higher elevations than the Inn floodplain itself, and so they have been sites for towns avoiding flooding by the Inn. The most prominent example of such a town is Innsbruck, which sits on the alluvial fan of the Sill and thus above the Inn floodplain. This explains why the first settlement of the area that is now Innsbruck took place in Wilten, where the Romans built a fort named Veldidina and where the Wilten monastery was built, near the apex of the Sill fan. Only later, when the Inn Bridge giving Innsbruck its name was built, did development begin in the Altstadt, much nearer the Inn itself and at the lower margin of the alluvial fan.

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[76] More precisely, the Trisanna and Rosanna join west of Landeck to form the Sanna, which then flows into the Inn at Landeck. The Rosanna's valley is the one taken by the railway line and autobahn westward to Vorarlberg and Switzerland.

[77] This is the breccia seen in the foundations of many buildings in Innsbruck, and in the walls of the Jakobsdom (the cathedral of St. Jakob) in the Altstadt.

The Inn Valley as a Corridor

The Inn Valley is today a major corridor through Tirol and western Austria, with at least six continuous avenues passing through it. By definition, the Inn River passes through the valley, and although it is not used today for boat traffic, it was in the past.[78] It is the corridor for the east-west railway from Vienna and Salzburg to Zurich.[79] It is the corridor for the autobahn or superhighway A12, and also for the two-lane highway B171[80]. It is also the corridor for the Inntal Radweg or bikepath, a non-trivial feature on a continent that loves bicycling. Finally, it is the corridor for high-voltage power lines carrying the electricity generated by the hydroelectric generating stations of the Alps, and a picture of the Inn Valley almost always has an electrical tower in the background.

In the lower and broader parts of the valley, all of these avenues combine to take up a non-trivial part of the valley floor, which is valuable real estate for farming as well as for commercial or residential development. In the upper parts of the valley, these avenues can combine to occupy almost all of the valley, requiring painful compromises about space and noise, especially in towns and villages built long before railways, power lines, and superhighways.

The Inn Valley, ecologically

In its natural state before human modification, much of the Inn Valley floor would have been a wetland. In its upper reaches, the river's banks and surrounding slopes would have been tree-covered, and it would have had local small floodplains. The same would have been true of its tributaries. In its lower reaches, the Inn flood plain would have consisted of marshy swamps hosting a variety of waterfowl.

Today, the ecological situation is much different. In the upper reaches, much of the tree cover has been removed, and many of the tributaries have been straightened and given steep stone-lined walls. In the lower reaches of the Inn, the floodplain has been drained to allow farming, and only very isolated remnants of ponds and marsh like the Gaisau near Inzing remain. The banks of the Inn have been lined with stone to constrain its wandering, and floodwalls have been built, most noticeably in Innsbruck, where vertical walls trap the Inn to keep it from flooding the city.

There are two principal results of these changes. One is that the draining of the Inn flood plain to allow farming has eliminated the habitat of much wildlife, and thus eliminated much wildlife, in the Inn Valley. The second is that all of these changes increase the rate of runoff into the river, and so increase flooding down-river. Straightened channels of tributaries and stone walls allow faster run-off and prohibit storage of water for a few hours or days after heavy rains. Drainage of floodplains obviously moves water from those floodplains more quickly into the river, rather than storing water after rains in ponds and marshes. The floodwalls of towns like Innsbruck protect them from small to intermediate floods, but each of those events sends more water running even faster down-river, leading to greater flooding in the Inn downstream and ultimately in the Danube. Rains in Tirol mean floods in Passau and Linz. Thus what has been good for development of upland real estate over the past few centuries has not been so good both for wildlife and for flooding of cities downstream.

The Inn Valley, economically

All of the above considerations make the Inn Valley the economic corridor of Tirol. In the Middle Ages, silver was mined at Schwaz and copper mined at Brixlegg, both from ores that were probably precipitated from deep groundwaters moving along the Inntal Fault.[81] Today, rail and

[78] In fact, the citizens of Hall supported themselves in the Middle Ages by maintaining a boom over the river so that boat traffic had to stop there and be unloaded and reloaded to boats that would then proceed on the other side of the boom.

[79] The Inntal from Kufstein to Innsbruck is also the corridor for the north-south railway from Germany to Italy.

[80] Traffic in Europe follows the American-style pattern of driving on the right, rather than the British-style pattern of driving on the left. Thus traffic on the Autobahn in the Inn Valley follows the same pattern as movement of the mountains along the left-lateral fault in the Inn Valley: south side to the east, and north side to the west.

[81] The silver mined at Schwaz was minted at Hall, and much of the silver went into coins called "talers" after their origin in the Inn Valley or Inntal. The word "taler" survives in English as the "dollar", and so every English-speaker's use of that word harkens back to the Inn Valley.

auto traffic through the valley bring tourist dollars, commerce, and jobs to Tirol. The valley floor provides the farmland that, while not so picturesque as farmsteads on the slopes of the Alps, is a larger part of the Tirolean agricultural economy. Virtually all industry in Tirol is in the Inn Valley, for reasons of both space and transportation. Distribution of habitable space means that most of Tirol's population is in the Inn Valley. In short, the geological and geographical considerations discussed above dictate the great socio-economic significance of the Inn Valley to Tirol and western Austria.

Further browsing:

Railsback, L.B., A Trip Down the Inn:
www.gly.uga.edu/railsback/AG/TDImain.html

The floodplain of the Inn River, the village of Baumkirchen, and the Calcareous Alps

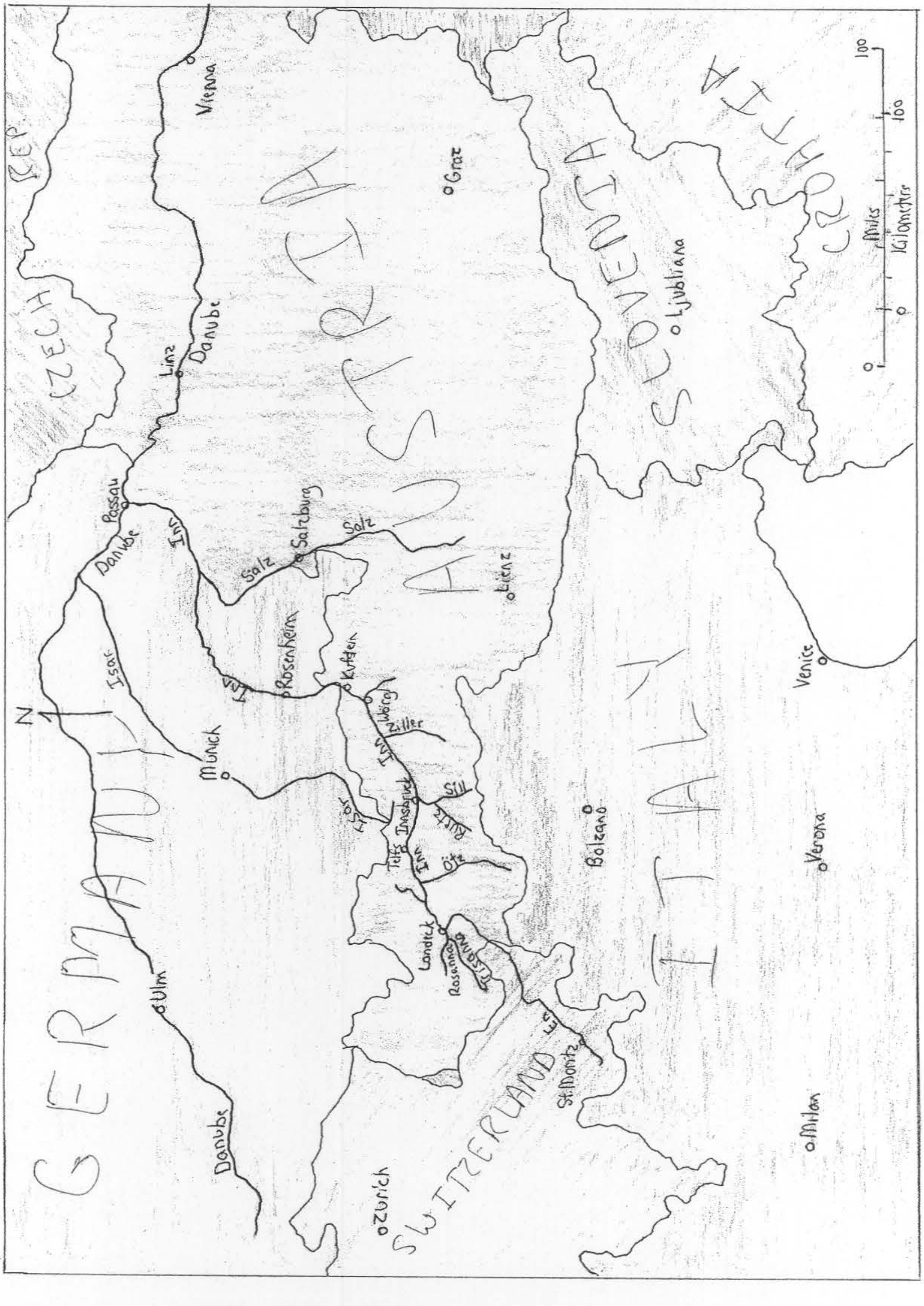

Figure 12-1: A map showing the Inn River, its major tributaries, and related rivers.

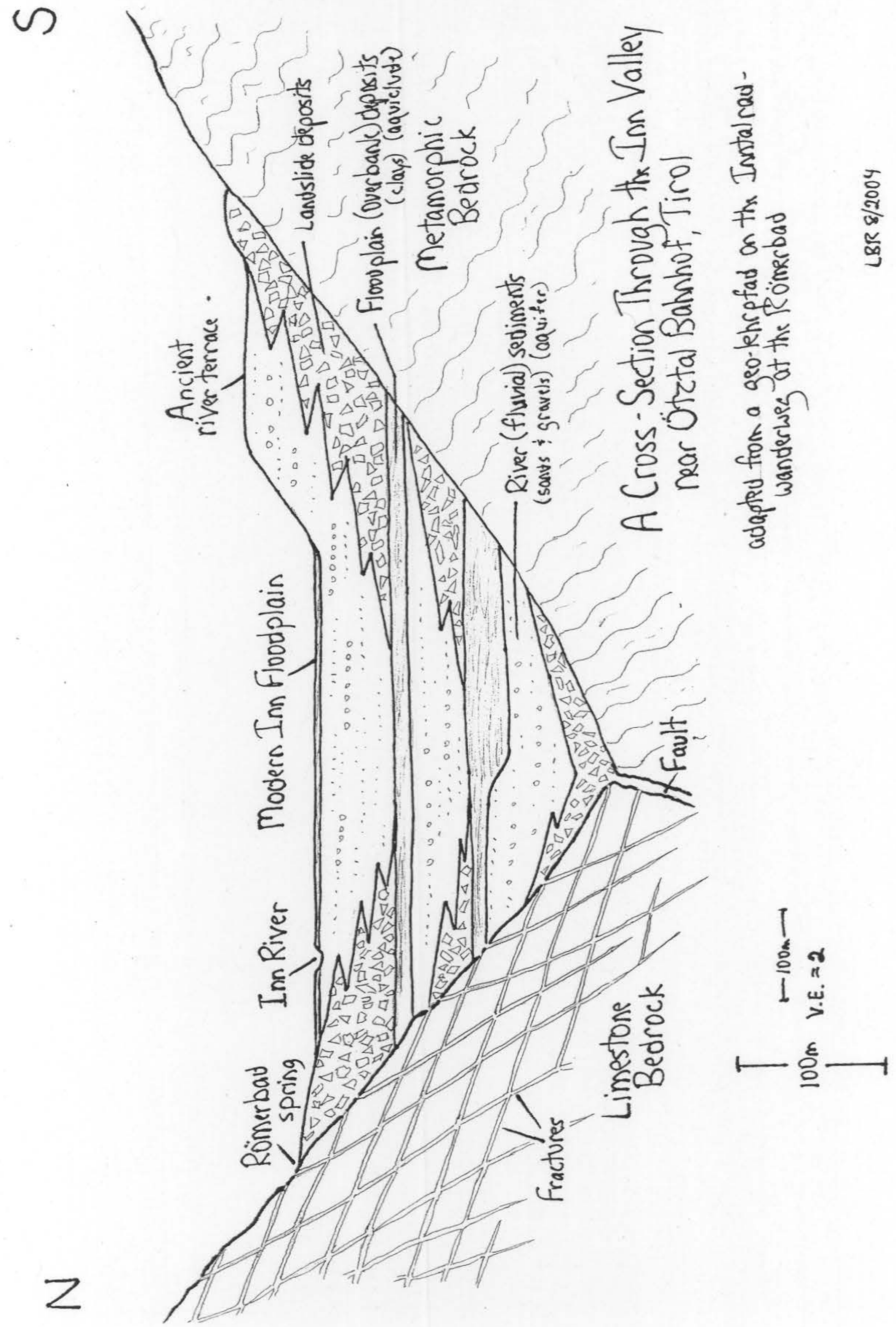

Figure 12-2: A cross-section of the upper Inn Valley.

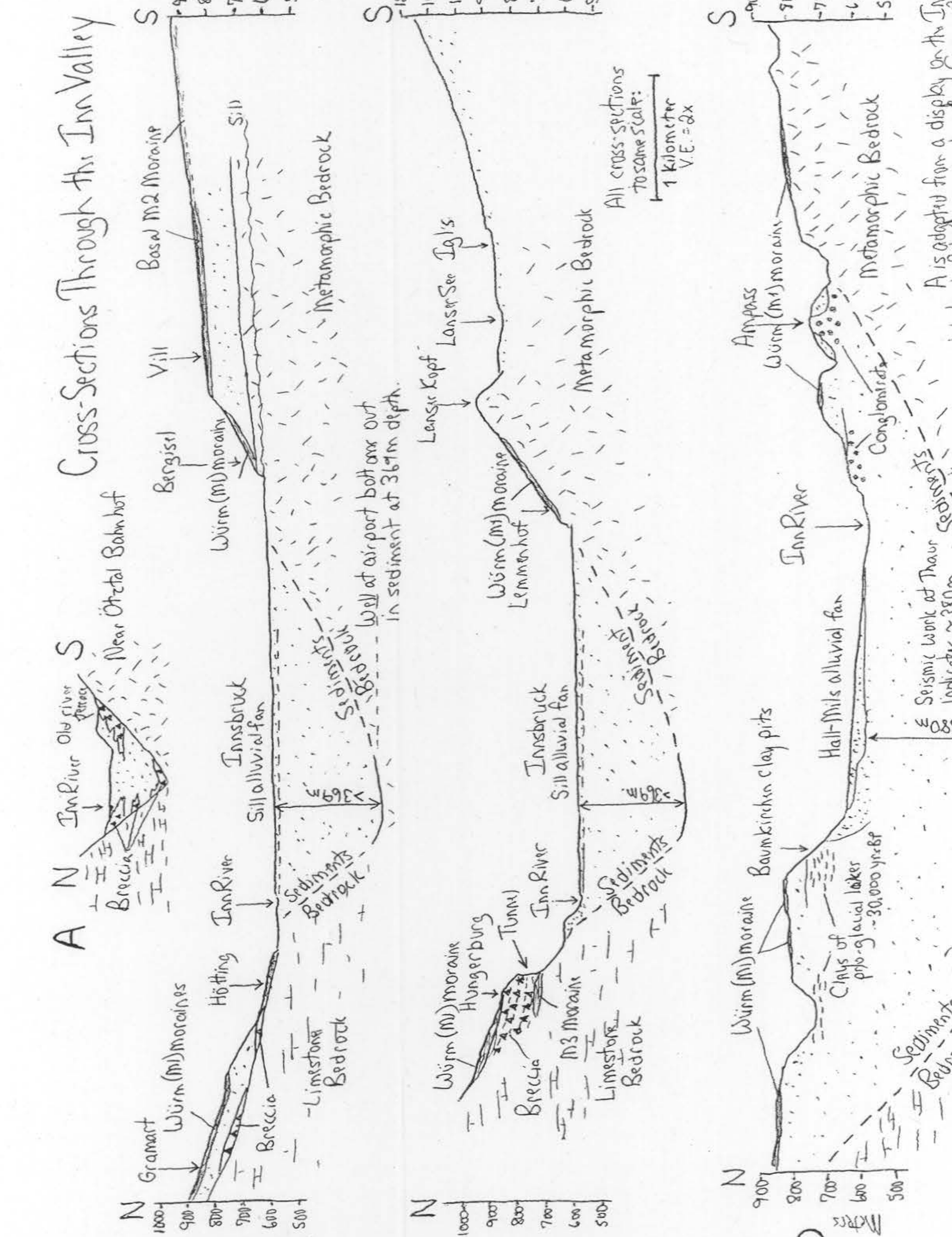

Figure 12-3: Cross-sections of the Inn Valley upstream and downstream from Innsbruck.

CHAPTER 13: THE SOUTHERN APPALACHIANS

Introduction

The Appalachian mountain range extends from eastern Canada to central Alabama and varies greatly both along its length and transversely from northeast to southwest. It has also changed greatly through time and is presently a rather faded remnant of its past glory. This chapter will describe the southern Appalachians and then attempt a brief history of their development.

Major zones

The southern Appalachians can be divided into three major zones framed by the Coastal Plain to the southeast and the Central Interior to the northwest. These three zones, the Piedmont, Blue Ridge, and Valley & Ridge, can be recognized at least from western Georgia to Virginia (Fig. 13-1).

The **Coastal Plain** is a region of undeformed Mesozoic to Cenozoic sediments and sedimentary rocks extending inland from the Atlantic coast to the Fall Line. These strata dip gently toward the coast and consist of marine and coastal deposits. They include sandstones and shales derived from the Appalachians and limestones produced by marine organisms. Triassic conglomerates can be found locally and are known from wells drilled below the Cretaceous to Tertiary cover.

The Coastal Plain is topographically very subdued, and outcrops of rock are scarce. Highways and railroads commonly extend along straight lines for miles to tens of miles because the landscape provides so few barriers to construction of thoroughfares.

The **Fall Line** is the boundary between the Coastal Plain and the Piedmont. It is so named because the Piedmont is more resistant to erosion, and thus streams and rivers have waterfalls, or at least rocky rapids, where they descend from the Piedmont onto the Coastal Plain[82]. These waterfalls are barriers to navigation of ships and barges coming up rivers from the coast, and towns developed on rivers where goods had to be unloaded onto wagons or onto other means of transport overland. As a result, the Fall Line is also a line of cities that include Tuscaloosa, Columbus, Macon, Milledgeville, Augusta, Columbia, Richmond, Washington, Baltimore, and Philadelphia.

The **Piedmont** is the southeastern-most of the three principal zones of the Appalachians from a geologic perspective, even though it topographically consists only of the foothills of the Appalachians (as its name of course implies). The Piedmont consists largely of high-grade gneisses that in places are migmatites, gneisses with gently flowing layers that suggest deformation very near the point of melting. Amphibolites are also common.

The Piedmont is dotted with granitic plutons, of which Stone Mountain and Panola Mountain near Atlanta are topographically prominent examples. Farther east in Georgia are granites near Elberton that have been the basis of a well-known quarrying and stone-cutting industry. These granites in Georgia are about 300 million years old, and their unmetamorphosed condition suggests that they postdate, if barely so, the major metamorphic events in the Piedmont. Farther northeast, the town of Granite Quarry, North Carolina, in the Piedmont near Salisbury attests to similar granites farther along in the southern Piedmont.

The presence of high-grade gneisses and amphibolites, as well as granitic intrusions, attests to the general deformation of the Piedmont at high temperatures and pressures. However, one should note that the Piedmont also contains belts of less deformed rocks, and Paleozoic fossils have been recognized in some of these rocks.

There are many ancient faults within the Piedmont, but one major zone of deformation is the **Brevard Zone**. This feature is named for Brevard, North Carolina, but it is readily recognized through North Carolina and Georgia to Atlanta. The Brevard Zone is a zone of intense deformation that has been interpreted at various locations by various workers as almost every kind of fault or shear zone. Like many faults, the Brevard Zone is

[82] The morphology of most rivers and streams also changes across the Fall Line. In the Piedmont, rivers and stream have narrow floodplains and follow irregular, sometimes straight and sometimes sharply angular, paths. Once these rivers and streams drop over the Fall Line, they commonly have broad floodplains across which they make large swooping meanders. The change in the character of the Savannah River at Augusta is good example.

topographically expressed as a linear depression, and Lake Lanier and the Chattahoochee Valley north of Atlanta follow the Brevard Zone to provide a strikingly linear pattern across north Georgia. The Brevard Zone is considered to be either a major boundary within the Piedmont or the boundary between the Piedmont and the Blue Ridge.

The **Blue Ridge** is a long and rather narrow belt that is the topographically highest part of the Appalachians. Great Smoky Mountains National Park, the Blue Ridge Parkway, and Shenandoah National Park all lie in the Blue Ridge. Blue Ridge rocks range from metamorphosed sedimentary rocks (metasediments) to schists and gneisses, with scattered igneous rocks. The metamorphic grade is thus generally less than that of the Piedmont, but faulting and folding have been intense.

The **Valley and Ridge** is a belt of folded and thrusted sedimentary rocks. These rocks range in age from the Cambrian to the Pennsylvanian, and Ordovician limestones and shales are the most abundant strata. These rocks have been pushed northwestward on thrust faults whose outcrop trends southwest to northeast. The available evidence suggests that the successive thrusts developed from southeast to northwest, the same direction as the thrusting itself.

The southwest-to-northeast trend of faults and folds is reflected in the topography of the region, which consists of valleys and ridges running southwest to northeast. The effect of this topography can be seen in highway maps, which show that most roads run along the valleys and thus southwest to northeast. Chattanooga and Knoxville are cities in the Valley and Ridge.

Beyond the Valley and Ridge is a region of flat-lying sedimentary rocks ultimately extending to the Mississippi River. In this sense the region is the undeformed Continental Interior, and in Tennessee and Kentucky it would be called the Cumberland Plateau. From the Appalachian perspective, it is most strikingly a region in which the youngest strata are Pennsylvanian sandstones. The coincidence of the age of these sandstones with the uplift of the Appalachians, and paleocurrent directions that indicate that the sands were transported from the southeast, combine to suggest that these sands were derived from the newly uplifted Appalachians and deposited in an **Appalachian Foreland Basin**. The sands were deposited in river systems, estuaries, deltas, and shallow seas that would have extended from the Appalachians to the Illinois Basin. Interbedded with these sandstones are the coals that have been mined so extensively in Tennessee, Kentucky, West Virginia, and Pennsylvania.

Geologic history

Most accounts of the history of the Appalachians involve three major orogenic events, the last of which is most evident in the southern Appalachians (Fig. 13-2). The first, in the late Ordovician, was probably the collision of an island arc or arcs with eastern Laurentia (ancestral North America). In the southern Appalachians, the **Blountian Orogeny** seems to have involved abortive subduction of eastern Laurentia into an encroaching subduction zone. In New England, a similar event at roughly the same time (perhaps 10 million years later) is called the Taconic Orogeny after the Taconic Mountains.

The second conventionally accepted orogeny in the Appalachians took place in the Devonian as the western European continent collided with eastern Laurentia in the Devonian. This event, the **Acadian Orogeny**, closed the ocean basin between Laurentia and Eurasia and so generated a large continent called Laurasia. Vast masses of sediment were shed west into the so-called Catskill Delta, actually a mass of non-marine, estuarine, deltaic, and shallow marine sands and shales in New York and Pennsylvania. A mirror-image mass of siliciclastic sediments was shed to the east and is known as the Old Red Sandstone in Britain.

In the Pennsylvanian, the southern margin of this new Laurasia collided with the margin of Gondwanaland (northwestern Africa in modern geography) in the closing of the remaining ancestral Atlantic and in the unification of Laurasia and Gondwanaland to form the Pangaean supercontinent. This collision may have involved as much lateral motion as head-on collision, and considerable debate exists to which side, Laurasia or Gondwana, overlay the subduction zone (and in fact there may have been multiple subduction zones dipping in opposite directions along the edges of the closing ocean basin). The result of the final Appalachian orogeny, known specifically as the **Alleghenian Orogeny**, was to weld Gondwanaland and Laurasia together with a mountain range analogous in style, but probably not in elevation, to the modern Himalayas.[83]

The Pangaean supercontinent was only short-lived and began to rift apart in the Triassic. In the southeastern U.S., it split not along the Gondwanan-Laurasian suture at roughly the present North Carolina-Tennessee boundary but instead to the east, from New Jersey to east-central North Carolina to central Georgia[84]. Parts of Gondwanaland were thus left attached as part of modern North America; much of the Piedmont probably falls in this category, and all of Florida certainly is old Gondwanan terrain now included in North America. As rifting opened basins on the east side of the newly defined North America, sediments began building out the modern Coastal Plain.

The Appalachians are thus now 300 million years old, an age at which most mountain ranges have undergone sufficient erosion that they are not major topographic features. Geologic evidence suggests that the Appalachians had probably declined to near flatness by the Cretaceous or early Tertiary. For example, the meandering paths of major rivers like the Susquehanna directly across the structural grain of the Valley and Ridge suggest that such rivers flowed across a flat plain in the late Mesozoic. The Appalachians seem to have been rejuvenated in the Tertiary and Quaternary, so that the mountains we see today enjoy a second life as a less prominent but still notable expression of an ancient collision of continents.

[83] Evidence that the Appalachians were not as high as the Himalayas comes in part from the lack of an increase in ^{87}Sr/^{86}Sr ratio of seawater in the Pennsylvanian (see the Himalayan chapter for more on ^{87}Sr/^{86}Sr ratios) and in part from the presence of abundant coal in the Appalachian Basin. The Appalachians sat roughly on the equator and thus in the Trade Winds belt of winds from east to west; had the Appalachians been as high as the Himalayas, they would have created a rain shadow that would have precluded the development of coal swamps on the west side of the Appalachians.

[84] From the broader perspective, the rift that split Pangaea was not parallel to the Appalachians. Farther north in the Appalachians, the rift encroached more on the mountain range itself, so that parts of the orogen were split off to the east side of the Atlantic and can be discerned in Scotland and Norway.

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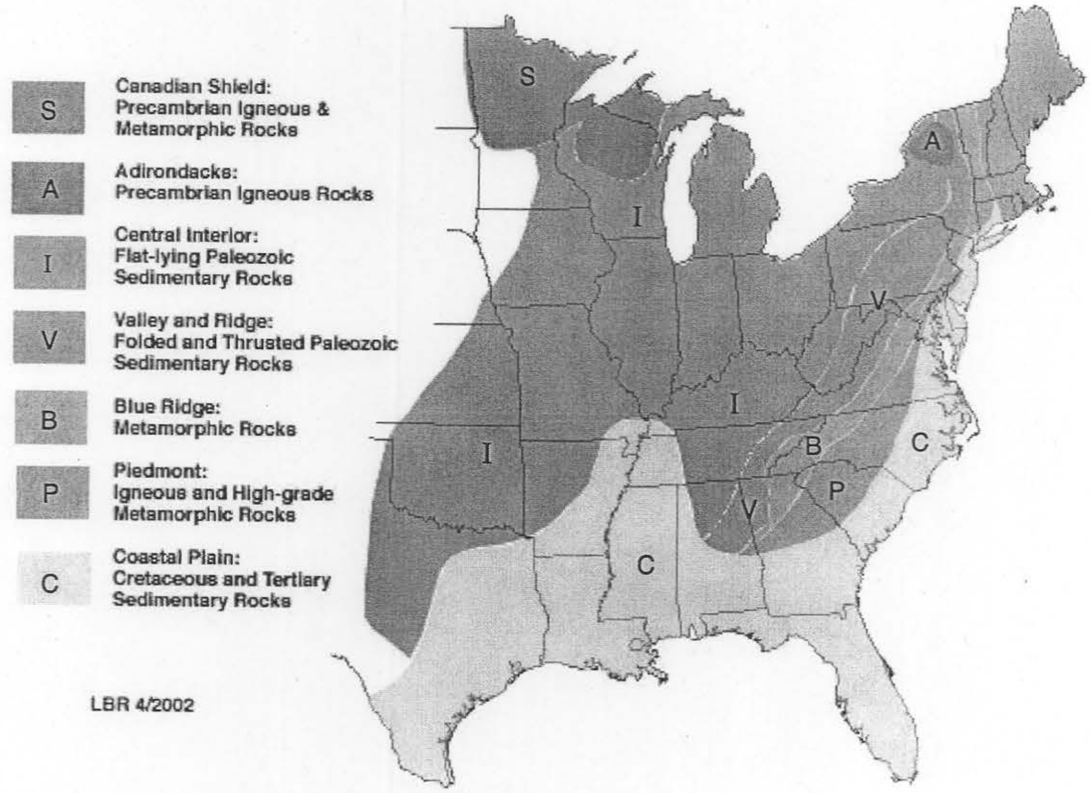

Figure 13-1: A generalized map of the geologic provinces of the eastern United States.

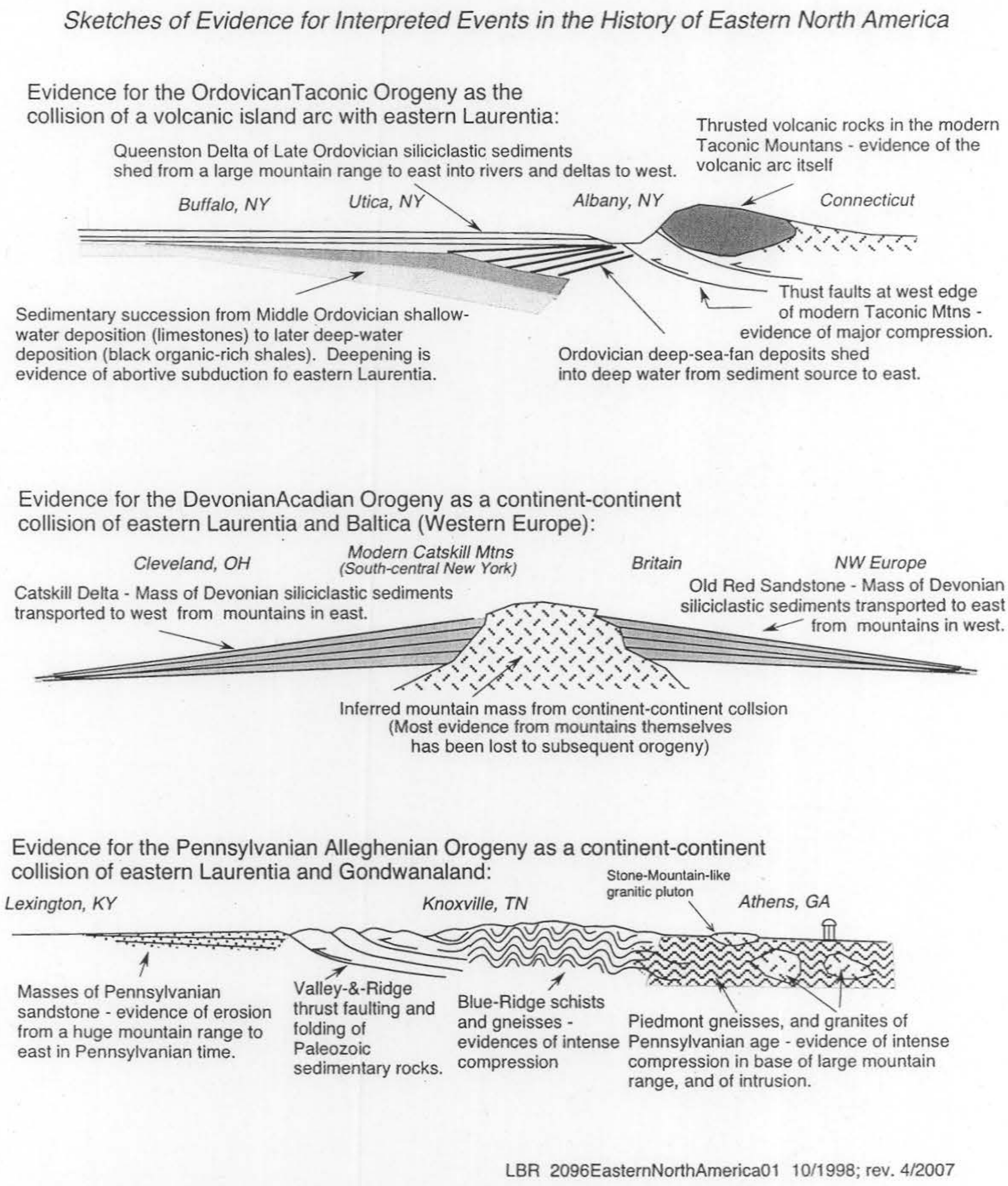

Figure 13-2: Schematic cross-sections illustrating some of the evidence for orogenies in eastern North America.

CHAPTER 14: THE HIMALAYAS

General observations

The Himalayan Mountains are the range of mountains defining the northern boundary of the Indian Peninsula. To the north is the Tibetan Plateau, and to the south are the Ganges and Indus plains and all of India. Adjoining this largely east-west mountain range on each end are ranges running north and south. These ranges adjoin the Himalayas, are part of the general Himalayan system, and originated in the same general tectonic event (the collision of the northward-drifting Indian continent with the southern margin of Asia). However, they are not part of the Himalayas in the strict sense and are considered "syntaxes" ("accompanying growths") of the Himalayas.

The Himalayas are presently Earth's most spectacular mountains, and they may well be among Earth's most spectacular mountains across geologic time. The first claim is supported simply by the elevations of the Himalayas. Chomolungma, or Mount Everest, is familiar to all as Earth's highest mountain at 8848 meters. Persons familiar with mountains and mountain-climbing will recognize K2, or Godwin-Austen, as the second highest at 8611 m. However, most of us fail to appreciate that at least 50 of Earth's highest peaks are in the Himalayas. The Himalayas' fifty-third highest peak is at 7756 m above sea level, whereas the highest peak outside southern Asia is the Andes' Aconcagua at just 6962 meters. Denali, or Mt. McKinley, is the highest peak outside southern Asia and the Andes, and it rises to only 6194 meters. For comparison, the highest peak of the Alps is Mt. Blanc at just 4807 meters, 4000 meters below the summit of Everest.

The claim that the Himalayas are exceptional mountains not only relative to the present but relative to much of geologic history is supported by geochemical evidence of global change, especially as evidenced by strontium isotopes. The $^{87}Sr/^{86}Sr$ ratio of strontium isotopes in seawater has varied through geologic history and has varied at least in part as a function of weathering of deep crustal rocks. Those rocks are rich in K-feldspar and thus contain a heavy isotope of rubidium, ^{87}Rb, that substitutes for potassium (K) in K-feldspars. ^{87}Rb decays to produce ^{87}Sr. Weathering of rocks rich in K-feldspar thus yields ^{87}Sr, and weathering of large masses of deep crustal rocks should thus lead to an increase in the $^{87}Sr/^{86}Sr$ in seawater. The $^{87}Sr/^{86}Sr$ record of seawater, known colloquially as the Burke curve[85], has its largest increase in the entire Phanerozoic during and since the uplift of the Himalayas, and that increase has been attributed to weathering of the Himalayas. For comparison, the building of the Appalachians in the late Paleozoic seems to have had no discernable effect on the $^{87}Sr/^{86}Sr$ ratio of seawater at all.

Other possible evidence of the long-term significance of the magnitude of the Himalayas includes the overall decrease in the CO_2 content of the atmosphere over the last 50 million years, which has been attributed to enhanced weathering of newly-exposed deep-crustal rocks. These arguments for the long-term significance of the Himalayas, as well as simply the presently observable remarkable elevations of the Himalayan peaks, certainly support further examination of the Himalayas as a great mountain chain.

Zonation of the Himalayas

The Himalayas are best understood as consisting of six zones, from north to south (Fig. 14-1). These are framed by the Tibetan Plateau to the north and the Ganges Plain to the south.

The **Tibetan Plateau** is, as its name implies, the high region of Tibet and thus north of the Himalayas. A first glance at a geologic map suggests that the Plateau is little different than the Himalayas that define its southern margin – both consist of east-west bands of rock types. The Plateau itself, however, is lower than the Himalayas, although still spectacularly high: for example Lhasa, the capital of Tibet, is at an elevation of about 3600 meters and thus sits higher than most of the peaks of the Alps.

The **Trans-Himalayan Batholiths** are the northernmost of the six zones of the Himalayas. They are large masses of Cretaceous to early Tertiary plutonic rock that formed on the southern margin of Asia over a subduction zone into which northward-moving Tethyan oceanic crust descended. To an American, they might thus be considered an analogue of the Sierra Nevada batholith that formed over North America's

[85] Burke, W. H., Denison, R. E., Hetherington, E. A., Koepnick, R. B., Nelson, H. F. and Otto, J. B., 1982, Variations of seawater $^{87}Sr/^{86}Sr$ throughout Phanerozoic time: *Geology,* v. 10, p. 516-519.

western Cordilleran subduction zone in the Mesozoic and that is now exposed after erosion of overlying volcanics. The Trans-Himalayan Batholiths thus not only are evidence of past subduction but also of an Andean-style volcanic chain now lost to erosion.

The **Indus-Tsangpo Suture Zone (ITSZ)** is thus named because the upper reaches of the westward-flowing Indus River and eastward-flowing Tsangpo River run through the valleys that it forms. It is called a suture zone because it appears to be the boundary between Asian rocks to the north and Indian rocks to the south and thus is where the two continents are sutured together. That observation is supported by the presence of intensely deformed deep-sea sediments from the bottom of the Tethys Seaway and of ophiolites (basalts and gabbros from the oceanic crust) from the Tethyan seafloor. The originally horizontal layering of these rocks is now nearly vertical after their intense compression in the suture zone.

The **Tethyan or Tibetan Himalaya** is a zone of Phanerozoic sedimentary rocks whose fossils confirm their origin as marine sediments. It is "Tethyan" in the geologic sense that the rocks originated in the Tethys Seaway, probably on the old northern passive margin of India. It is "Tibetan" in the sense that it includes the highest parts of the Himalayas that are in Tibet, and in fact it does not extend all the way to the west end of the Himalayas. Chomolungma, or Mount Everest, is in the southern Tethyan Himalaya, and its sedimentary origins are evident from its obvious layered structure.

The sediments of the Tethyan Himalaya have been uplifted along thrust faults, most of which rise southward on the south side of the belt but some of which rise northward along the north side of the belt. Figures 5 and 6 of Searles et al. (1987) suggest that there are at least twenty-four thrust faults, seventeen rising to the south and seven to the north, in a transect across the western Tethyan Himalaya. About 250 kilometers of horizontal strata have been shortened to about 98 present kilometers by the displacements along these faults.

The **Higher or Greater Himalaya** is a zone of high-grade metamorphic rocks, as well as some igneous rocks, that extends along the entire length of the Himalayas. It consists of deep crustal Indian rocks that have been uplifted along thrust faults that rise to the south. The southernmost of these thrust faults, and thus the southern boundary of the Higher Himalaya, is the Main Central Thrust (MCT). Figure 12 of Searles et al. (1987) suggests that all of these thrust faults formed after thrusting of the Tethyan Himalaya, so that the Higher Himalaya is a relatively late-uplifted mass but with a deeper origin than its neighbor to the north.

The Higher Himalaya includes most of the highest peaks of the Himalayas, so that its name is well deserved. However, one should note the irony that Chomolungma, or Mount Everest, is not in the Higher Himalaya but instead sits slightly to the north in the Tethyan Himalaya.

The **Lower or Lesser Himalaya** include Proterozoic and Lower Paleozoic strata from the old northern Indian passive margin. Many of these rocks have undergone metamorphism to at least lower metamorphic grades. The Lower Himalaya also include some klippen, outlying remnants of thrust sheets, from the High Himalaya. The rocks of the Lower Himalaya have been pushed up along thrusts that rise to the south, and the southernmost of these, and thus the southern boundary of the Lesser Himalaya, is the Main Boundary Thrust (MBT).

The **Sub-Himalaya** is the southernmost of the six zones of the Himalayas. It consists of Cenozoic sedimentary rocks, largely designated as the Muree and Siawalik Groups, of siliciclastic sediments derived from the Himalayas. They are thus generally considered mollasic sediments but have undergone some thrusting and folding. The Sub-Himalaya can be thought of loosely as the foothills of the Himalayas.

The **Ganges Plain** is the undeformed alluvial plain of the Ganges River, which flows southeastward south of, or in front of, the Himalayas. The Ganges carries a phenomenal amount of sediment eroded off the mountains, as do the Indus in the western Himalayas and the Tsangpo-Brahmaputra north and then east of the Himalayas. Rivers draining the Himalayas and Tibet drain about 4% of Earth's land surface but carry 25% of Earth's fluvial sediment load, and the Indus and Ganges rivers have built huge submarine fans of sediment in the northern Arabian Sea and Bay of Bengal, respectively.

History of the Himalayas

Both the present structure of the Himalayas and paleomagnetic data recording the motion of India northward from the break-up of

Gondwanaland support the widely accepted interpretation that the Himalayas arose in the collision of the north passive margin of India with the southern active margin of Asia. The syntaxes to the east (e.g., in Burma) and west (e.g., in Pakistan) are zones of deformation resulting from the insertion of northern India into southern Asia. The present structure of the Tibetan Plateau appears to have arisen largely as the result of its being squeezed eastward between the Himalayas to the south and the Altyn Tagh fault to the north (Fig. 14-2). In fact, the protrusion of Indochina and China into the western Pacific appear to be the result extrusion tectonics driven by the collision of India with Asia.

The timing of the Himalayan orogeny can be estimated from the ages of rocks deformed and from the ages of sediments generated as the mountains arose. Collision probably began in the Eocene, about 40 to 50 million years ago, and seemingly continues today. In general, deformation progressed from north to south, with earliest faulting and folding of the ITSZ, then faulting of the Tethyan Himalaya, and then uplift of the High Himalaya (Fig. 14-3). Thrust faulting within the late Tertiary and Quaternary sediments of the Sub-Himalaya exemplifies both the progression of thrusting southward and its continuation until recent times.

This simple history of the Himalayas overlooks many pre-Himalayan events. For example, recent studies of strata derived from the northern margin of India show that thrusting deformed those strata in the early Paleozoic (Gehrels et al., 2003). Those strata, thrusted in the early Paleozoic, are now nested in the Cenozoic-thrusted Lesser Himalaya. Further examples of earlier deformations included within the Himalayas are likely to become apparent as study continues.

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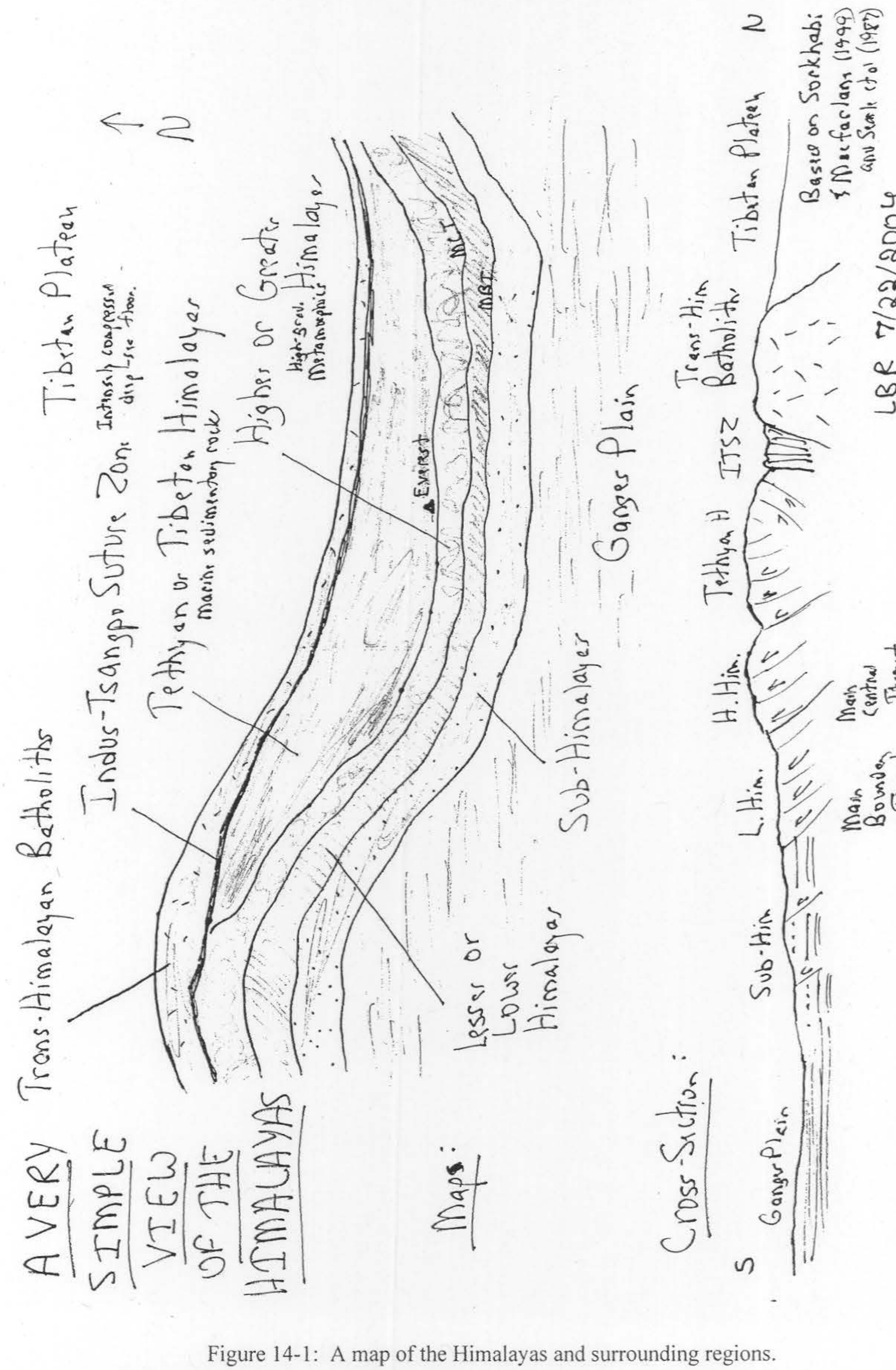

Figure 14-1: A map of the Himalayas and surrounding regions.

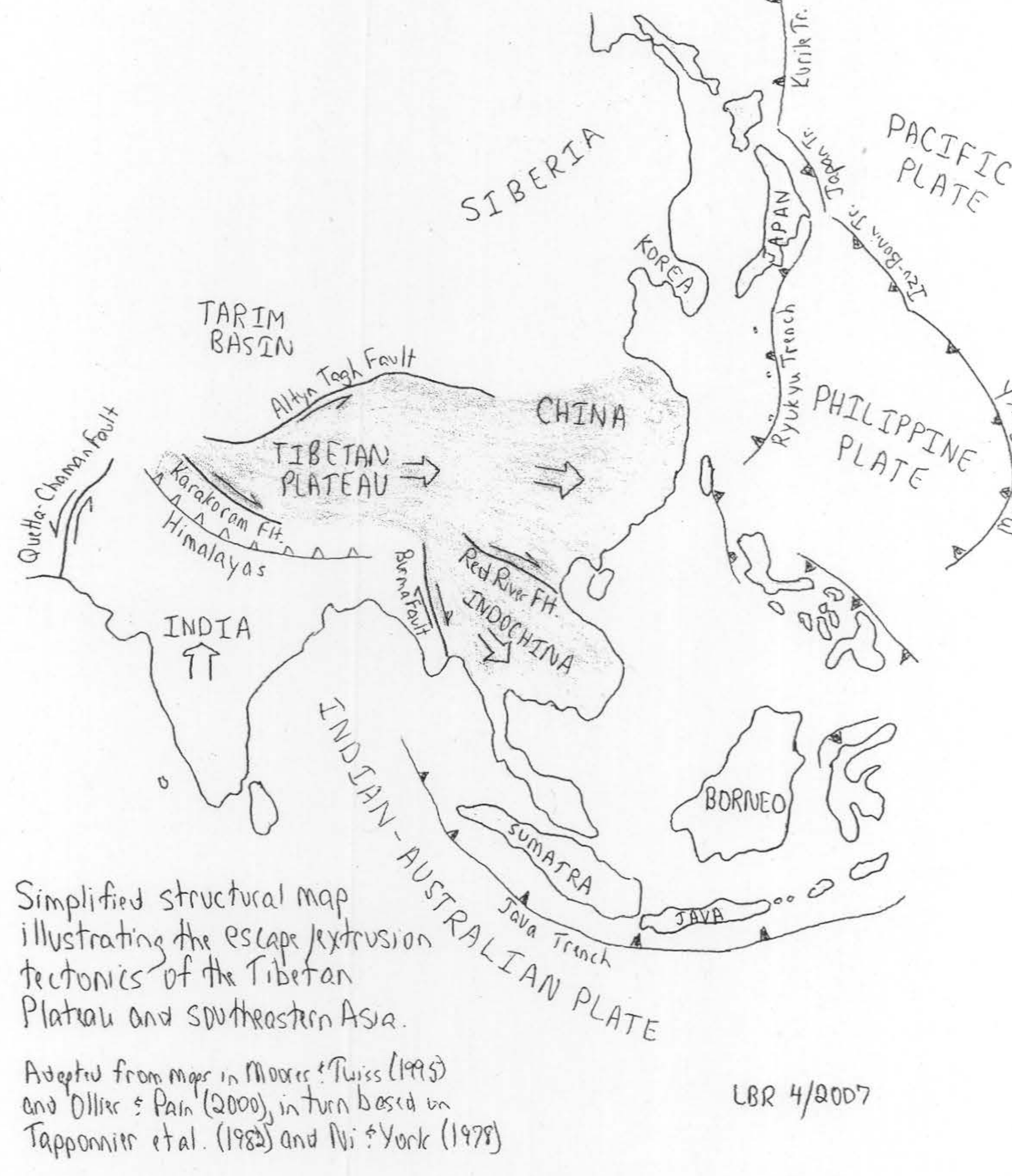

Figure 14-2: A map showing the origin of the Himalayas and the Tibetan Plateau.

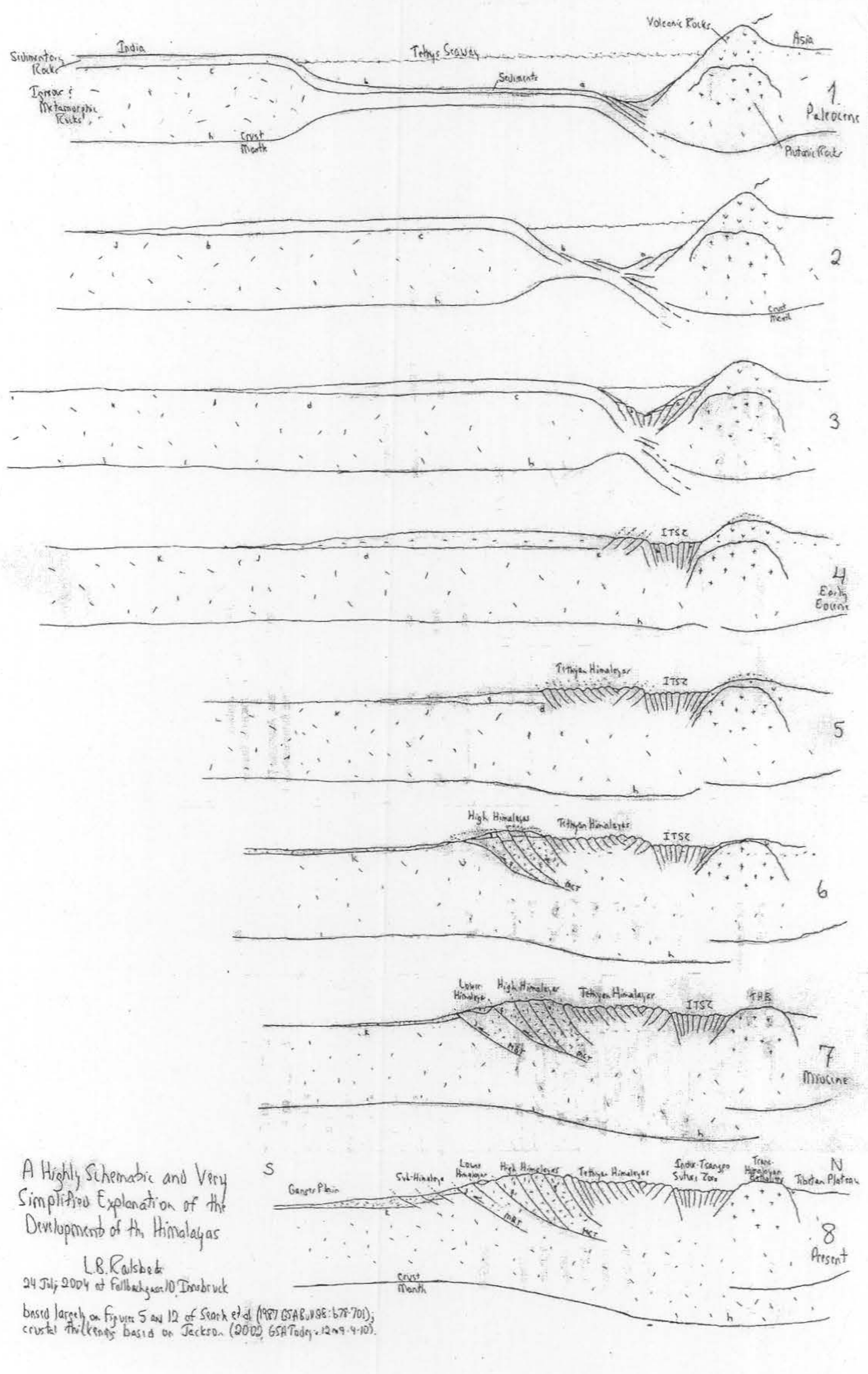

Figure 14-3: Cross-sections showing one interpretation of the origin of the Himalayas.

CHAPTER 15: FOUR CIRCUM-PACIFIC OROGENS

Most major orogenies take place at convergent plate boundaries. Other chapters have already dealt with Earth's recent orogenies at convergences of continental crust with continental crust, the orogenies that formed the Alps and Himalayas. This chapter deals with orogenies at convergences of oceanic crust with either continental crust or other oceanic crust. They are located around the Pacific, and in fact the circum-Pacific belt provides a zoo of possible results of subduction of oceanic crust (Fig. 15-1). Our trip around the Pacific will move in order of the age of the subducted oceanic crust, and not coincidentally in order of complexity of the resulting orogens. After examining (1) subduction of old oceanic crust under younger oceanic crust in the western Pacific, we'll look at a range of kinds of subduction of oceanic crust under continental crust, ranging from (2) subduction of old oceanic crust off Japan to (3) subduction of relatively young oceanic crust off western South America to (4) subduction, or lack thereof, of very young oceanic crust off western North America.[86] In all four cases, subduction and resulting dehydration of the downgoing slab have released water that has induced melting in the overlying lithosphere, and thus led to volcanism on the over-riding plate. However, in other respects, these four cases are very different, in part because of the differing age of the subducting oceanic crust.

The Marianas Trench and Tonga Trench

In the western Pacific, oceanic crust of the Pacific Plate has traveled far from its formation along the East Pacific Rise. It is thus relatively old oceanic crust, ranging in age to as much as 180 million years, the greatest age of any oceanic crust, just east of the Marianas. The age, and thus relatively low temperature, and thus relatively great density, of this crust doom it to subduction beneath the oceanic crust it encounters, which includes the Australian Plate west of the Tonga Trench and the Philippines plate west of the Marianas Trench. The subduction of Earth's oldest crest in the Marianas Trench leads to the oceans' greatest depth at about 11,000 meters in the Challenger Deep.[87]

The orogenic result of subduction of the Pacific Plate in these trenches has been the development of island arcs of volcanoes that include the islands of the Tongan chain and the Marianas. The islands themselves are small; for example, the highest peak in Tonga is only 1033 meters above sea level, and similarly Mt. Lamlam on Guam in the Marianas Islands only reaches 406 meters above sea level. However, these islands are only the peaks of volcanoes that are, in the case of the Marianas, thousands of meters high relative to the seafloor of the overriding plate on which they sit (Fig 15-2).

Behind (west) each of these island arcs is a back-arc basin, an elongate basin with a small mid-ocean-ridge-like spreading center parallel to the subduction. A back-arc basin results from extension in the over-riding plate. That extension results from advancement of the volcanic arc and subduction zone toward the subducting plate as it undergoes both subduction and rollback. That is why the great age of the western parts of the Pacific Plate is so important: the plate is sufficiently dense that it founders and thus undergoes rollback, pulling the arc forward and ultimately generating the back-arc basin. For comparison, the Aleutian island arc trench and Antilles trench subduct younger crust, and any rollback there has seemingly not been sufficient to generate a back-arc basin. The back-arc basin behind the Tonga Trench is the Lau Basin, and the one behind the Mariana Trench is the Mariana Trough.

Japan

North of the Marianas Trench and the associated Bonin Trench, the Pacific plate

[86] For an oceanic perspective, remember that the East Pacific Rise, the divergent plate boundary of the Pacific, is not in the center of the Pacific but to its southeast. Thus the Pacific Plate moving westward has a great distance to travel to the western trenches, and so it reaches great age. The plates going east don't travel nearly so far and so are younger when they reach the trenches on the east side of the Pacific.

[87] The Challenger Deep is the deepest part of the Marianas Trench, and the deepest point in the oceans. Its depth is estimated at 10,900 to 11,034 meters. Its name is ultimately derived from that of the British research vessel *Challenger* that circled the globe in the 1870s. The US diving vessel *Trieste* reached the bottom of the Challenger Deep in 1960. No diving vessel today is capable of descending into the Challenger Deep.

descends into the Japan Trench and thus under the Eurasian plate. The most obvious orogenic results are volcanoes, of which Mt. Fuji is the most famous. Fujisan reaches an elevation of 3776 meters, the highest point in Japan. Other Japanese volcanoes include Akan, Aso, Kirishima, Kowagatake, Kuttyaro, Mashu, Rausu, Sakurajima, Shiretoko-Iwo-Zan, and Unzen. Most or all are stratovolcanoes that are prone to erupt explosively, rather than as lavas.

Volcanoes are the most striking geological features of Japan, but they account for only a small part of the Japanese islands. Most of Japan consists of non-volcanic rocks, and Japan as a whole is continental crust. The topographic backbone of Japan consists of three mountain ranges, the Hida, Kiso, and Akaishi ranges, of which the highest peak reaches 3192 meters above sea level. These mountains, which from north to south combine to form the so-called Japanese Alps, consist mostly of granitic, sedimentary, and metamorphic rocks, rather than volcanic rocks. Fuji sits well off the trend of these mountains on a plain facing the Pacific.

Behind, or westward of, Japan is the Sea of Japan. The Sea of Japan is a back-arc basin, with ocean seafloor that is a few tens of millions of years old. It is generally accepted that the continental crust of which Japan consists has split from the Eurasian mainland as the subduction zone moved eastward with rollback of the foundering Pacific plate. The back-arc basin of the Sea of Japan is the feature that has allowed the extension of the Eurasian plate eastward. As in the case of the Marianas and Tonga, the age and density of the western Pacific Plate seemingly accounts for the rollback that has in turn led to this extension and seaward migration of the Japan Trench.[88]

The Andes

Along the western coast of South America, the Nazca Plate is subducted beneath the South American plate. The subducting oceanic crust is 20 to at most 50 million years old, much younger than the oceanic crust subducted in the western Pacific. Because the western margin of South America is not linear, subduction must have been in the past, and must be today, oblique (rather than orthogonal) along at least part of the South American western continental margin.

The orogenic result of this subduction is the Andes mountains, a chain more than 7000 kilometers long and containing Earth's highest peaks outside southern Asia (Fig 15-3). Those peaks include non-volcanic mountains (e.g., Aconcagua[89] at 6960 meters, Bonete at 6872 meters, and Tupungato at 6800 meters) and volcanic peaks (e.g., Ojos de Salado at 6908 meters).

One question that a newcomer might ask is why such young crust is subducted at all. For example, oceanic crust more than 100 million years old survives without subduction in the South Atlantic, and oceanic crust more than 150 million years old survives without subduction in the North Atlantic. The answer is probably that subduction of the relatively young crust of the Nazca Plate is at least in part the result of history. The Atlantic Ocean has clearly opened over the last 150 million years, meaning that South America would have been farther east in the past, and thus farther from the East Pacific Rise.[90] The Nazca Plate of the past would thus have had farther to travel eastward, and thus more time to cool and become dense. Thus subduction off the western margin of South America over the past 100+ million years would have been subduction of an older eastern margin of the Nazca Plate. Continued sinking of that older now-subducted part of the Nazca Plate may now be dragging behind it the present eastern margin of the Nazca Plate into the subduction zone, and/or continued sinking of that older part of the plate may maintain mantle flow that drives the South American plate west.

The Andes are a complex mountain range, and detailed description requires dividing them into several zones from north to south. However, one can generalize about at least the central Andes from Colombia to central Chile and Argentina (Fig. 15-4). In this region, the westernmost Andes

[88] Much farther westward of Japan is Lake Biakal, another extensional feature (specifically, a rift basin). Lake Baikal's long direction parallels Japan and its back-arc basin, so that Lake Baikal's direction of extension is compatible with distal back-arc extension. See Schellart, W.P., and Lister, G.S., 2005, The role of the East Asian active margin in widespread extensional and strike-slip deformation in East Asia: Geophysical Research Abstracts, v. 7.

[89] Aconcagua: ah-kone-KAH-hwa

[90] The use of "east" here is in reference to modern position of the plates. "East" is thus used in a relative sense, rather than with regard to absolute plate motions.

commonly consist of a coastal range of Permian to Mesozoic-age batholiths, clearly suggestive of past magmatism there. To the east are thrust faults on which overlying rocks have been moved to the west. East of these faults is the Western Cordillera or Cordillera Principal, a high range including Aconcagua[91]. The western Cordillera is the site of most later Tertiary to modern volcanism. Eastward of this zone is the Altiplano, a high plateau resting atop thrust faults. East of the Altiplano is a region of thrust faults, generally with motion of the overlying sheets to the east, which have generated ridges that make up the Eastern Cordillera and/or the Subandean ranges (the region of the most recent crustal shortening, at least in the central Andes[92]). In central Argentina, deformation extends farther eastward to include the upthrusted Pampean Ranges. In addition to all these compressional features, strike-slip faults cut through the western zones, presumably as the result of oblique subduction (Fig. 15-4).

One striking variation along the length of the Andes is the presence and absence of volcanoes in segments of the chain. Volcanoes are present where the depths of earthquake foci indicate that the subducting plate dips more steeply, and they are absent where such evidence indicates that the subducting plates descends with a more shallow dip.

The evidence above suggests that Andean subduction of the Nazca Plate has changed greatly over the last 200 million years. The dip of the subducting plate is thought to have decreased, as one might expect as progressively younger and more buoyant oceanic lithosphere has been subducted. Additionally, and presumably as a result, the zone of volcanism has moved eastward, so that the present major volcanoes are east of the Permian to Mesozoic batholiths of the coastal ranges. Thrusting in the east is also a relatively late (i.e., Tertiary) development for the most part (Ramos and Aleman 2000).

The Andes presently have no back-arc basin. However, there was a back-arc basin in the northern to central Andes in the Triassic to Jurassic. The back-arc basin underwent closure or inversion in the Late Jurassic to Cretaceous. Thus deformation along the Andean margin had an extensional component early in its history – an extensional component analogous to that of the modern western Pacific subduction zones, at a time when Andean convergence subducted oceanic lithosphere nearer in age to that of the present Pacific Plate in the western Pacific. On the other hand, more recent subduction of younger crust less prone to rollback has seemingly caused only compressional deformation in the Andes. That purely compressional mode of deformation may account for the extreme elevation and broad zone of tectonism that characterize the modern Andes.

Western North America
(the western contiguous U.S., and Canada).

In the Andes and the adjacent Pacific, we saw oceanic crust tens of millions of years old come from a mid-ocean ridge to the west and be subducted under the continent. Off western North America, the mid-ocean ridge is much closer, and in places has already been lost under western North America. Thus the oldest oceanic crust subducted along the western coast of North America is only a few million years old, in contrast to the older crust subducted at the margins discussed above. However, along much of the western coast of North America, no crust is subducted at all, because the mid-ocean ridge of the eastern Pacific has already passed beneath North America, leaving regions like western California on the Pacific Plate rather than the North American plate (Fig. 15-5). However, it seems reasonable to assume that, before the mid-ocean-ridge passed under the western margin of North America, subduction of east-moving oceanic crust would have taken place along the entire margin.

The orogenic result of this peculiar plate-tectonic arrangement is a complex array of mountain ranges and geologic provinces (Fig. 15-5). To describe them, we'll move through them geographically from west to east and also historically from older to younger, coming back from the east to look at more recent structures superposed on older. As always, this will be a

[91] "Cordillera" is derived from the Spanish word for "rope" and refers to a strikingly long mountain range, and thus almost inevitably to a mountain range at an active continental margin. The term is also used in North America, where it is instructive to note that the Geological Society of America includes a Cordilleran Section distinct from the Rocky Mountain Section. However, some geologists use "cordillera" for all the mountains of western North America from the Rockies to the west coast.

[92] Oncken, Hindle, et al. (2006), in Oncken, Chong et al. (2006).

highly over-generalized simplification attempting to bring together the most basic aspects of the entire system.

First, to almost literally set the stage for later events, one must begin with the observation that much of the western margin of North America consists of "suspect terranes" or "exotic terranes" that were welded onto the North American craton in the later Paleozoic and Mesozoic. Various sorts of subduction zones west of North America, or on the western margin of North America, allowed relatively small blocks of various origins to accrete onto North America, somewhat like an ice scraper (North America) moving across a windshield (ancient oceanic crust west of North America) and accumulating blocks of frost and ice (the exotic terranes) that adhere to the front of the ice scraper. Much of what will follow will happen on or in these accreted terranes.

With that stage set, we can begin on the western shore and move inland (Fig. 15-6). West of the Cascades and Sierras, west-verging thrust faults pass through wedges of sediment that either were shed west off the continent or scraped off the underlying oceanic plate. East of this is the principal zone of magmatism, with the Jurassic-to-Cretaceous-age granitic batholith of the Sierra Nevadas and the younger volcanoes of the Cascades (and volcanoes of western Mexico as well). These regions are collectively the Cordillera of North America. They are cut by major long right-lateral strike-slip faults that collectively extend from California to Alaska and include the Tintina and Straight Creek faults in Canada and the Walker Lane and Mojave-Sonora faults in the U.S. The San Andreas fault is also a spectacular modern example, serving as the transform plate boundary today between the North American plate and the Pacific Plate.

East of the Cordillera is a region of east-verging thrust faults that extend at least to the Absaroka Mountains east of Yellowstone National Park in Wyoming, and to the east side of the Canadian Rockies. These thrusts formed in Cretaceous and Tertiary. Farther east are isolated mountain ranges bounded by reverse faults. These are the Tertiary-age Laramide mountain ranges and include the Wind River Mountains, Bighorn Mountains, Laramie Range, Front Range of the Rockies, and ultimately the Black Hills, the axes of which all trend north-northwest to south-southeast.[93] Farther east still is the foreland basin of Cretaceous and Tertiary sediments extending across Kansas, Nebraska, the Dakotas, and the adjacent regions to north and south (Fig. 15-5).

Finally, we can return westward to look at two major more recent features superposed on this scheme. First, the Basin and Range is a region of normal faulting extending from the east side of the Sierras, at the Walker Lane fault, eastward across Nevada to central Utah. These normal faults cut apart much of the thrust belt that remains comparatively intact in northwest Wyoming, Montana, and Canada. The result is series of north-south mountain ranges separated by basins of sediments shed off those ranges. Secondly, cutting east to west from Oregon to Yellowstone is a belt of Quaternary-age volcanic rocks commonly attributed to passage of the North American plate over a hot spot presently located under Yellowstone.[94] The culmination of this cross-cutting sequence can be seen in northern Nevada and southern Idaho, where Paleozoic rocks thrusted in the Mesozoic have been cut into the north-south strips of the Basin and Range, and then the Snake River Plain hotspot volcanics cut west-southwest to east-northeast across those north-south strips (Fig. 15-6).

To compare western North America with the Andes, let's ignore the Basin-and-Range and Snake River-Yellowstone-hotspot complications and consider from west to east the belts of west-verging thrusts, commonly extinct magmatic arc, east-verging thrusts, and Laramide uplift (Fig. 15-7). The pattern is similar to the west-verging thrusts, volcanic arc and altiplano, east-verging

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[93] Even farther east is the Cedar Creek Anticline, a subtle arch in eastern Nebraska and western North Dakota, again with its axis from north-northwest to south-southeast. Although not an orogenic feature (nor even a topographic feature), it is a geologic structure compatible with compression parallel to that of the Laramide uplifts.

[94] However, the volcanic rocks of the Snake River - Yellowstone system do not progress steadily from older in the west to younger in the east. For example, the most recent volcanism is not at Yellowstone but at Craters of the Moon in south-central Idaho. For more skepticism about a plume as the source of these volcanics, see Christiansen, R.L., Foulger, G.R., and Evans, J.R., 2002, Upper-mantle origin of the Yellowstone hotspot: Geological Society of America Bulletin, v. 114, p. 1245-1256.

thrust, and Pampean uplift of the Andes, but only more so: the volcanic arc has more broadly become extinct, and the eastern uplifts are not one isolated Pampean range but instead the several Laramide ranges. The plate-tectonic setting is similar to that of the Andes, but again more so: if the Andes result from subduction of relatively young oceanic crust, subduction off western North America involves even younger oceanic crust, and has even progressed to subduction of the mid-ocean-ridge itself. As in the Andes, the dip of the subducting plate is believed to have decreased as progressively younger crust was subducted, accounting for the compressional uplifts farther and farther east.

The Basin and Range and the Snake River - Yellowstone-hotspot trail are two later complications superposed on the "Andes-gone-wild" view of western North America (Fig. 15-6). Development of the Basin and Range may have been an inevitable result of subduction of very young oceanic crust and then the mid-ocean-ridge itself, in that the latter event required that some of westernmost North America join the northwest-moving Pacific Plate, resulting in extension of North America to the west. On the other hand, a conventional understanding of the Snake River - Yellowstone system as the result of hotspot volcanism suggests that this final disruption of the continental interior was a matter of random chance, in that the location of the mantle plume had little to do with prior tectonic events in the North American crust.[95]

Summary

In our progression from subduction of older oceanic crust to younger, we've observed abundant volcanism and back-arc extension over steeply-dipping subducting plates give way to lesser volcanism and more extensive back-arc compression over less steeply-dipping subducting plates (Fig. 15-8). The most convenient explanation of this relationship is that the greater density of older oceanic crust allows steep subduction, and it allows roll-back that extends the back-arc region. One the other hand, the lesser density (or relatively greater buoyancy) of younger oceanic crust lets the plate sink less to give less steep subduction, to allow less roll-back for back-arc extension, and, in dipping less steeply, to cause back-arc compression and Laramide/Pampean-style uplifts.

The ultimate state of this progression is subduction of such young ocean crust that the mid-ocean ridge is subducted, leaving at least part of the cordillera on the plate moving away from the continent. This reverses the tectonic progression, in that increasing compression gives way to extension as the oceanward-moving plate pulls the cordillera apart (Part 4 of Fig 15-8).

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[95] Or maybe not. LBR hypothesizes that the Snake River - Yellowstone system is the result of tearing of the North American plate in reponse to the northwestward pull of the Pacific Plate, and thus a later and rotated manifestation of the same extension that generated the Basin and Range. This tear might extend all the way to the remarkably straight valley of the Yellowstone River in eastern Montana: that valley lines up well with the trend of the Snake River Plain. However, this hypothesis remains untested geophysically or otherwise.

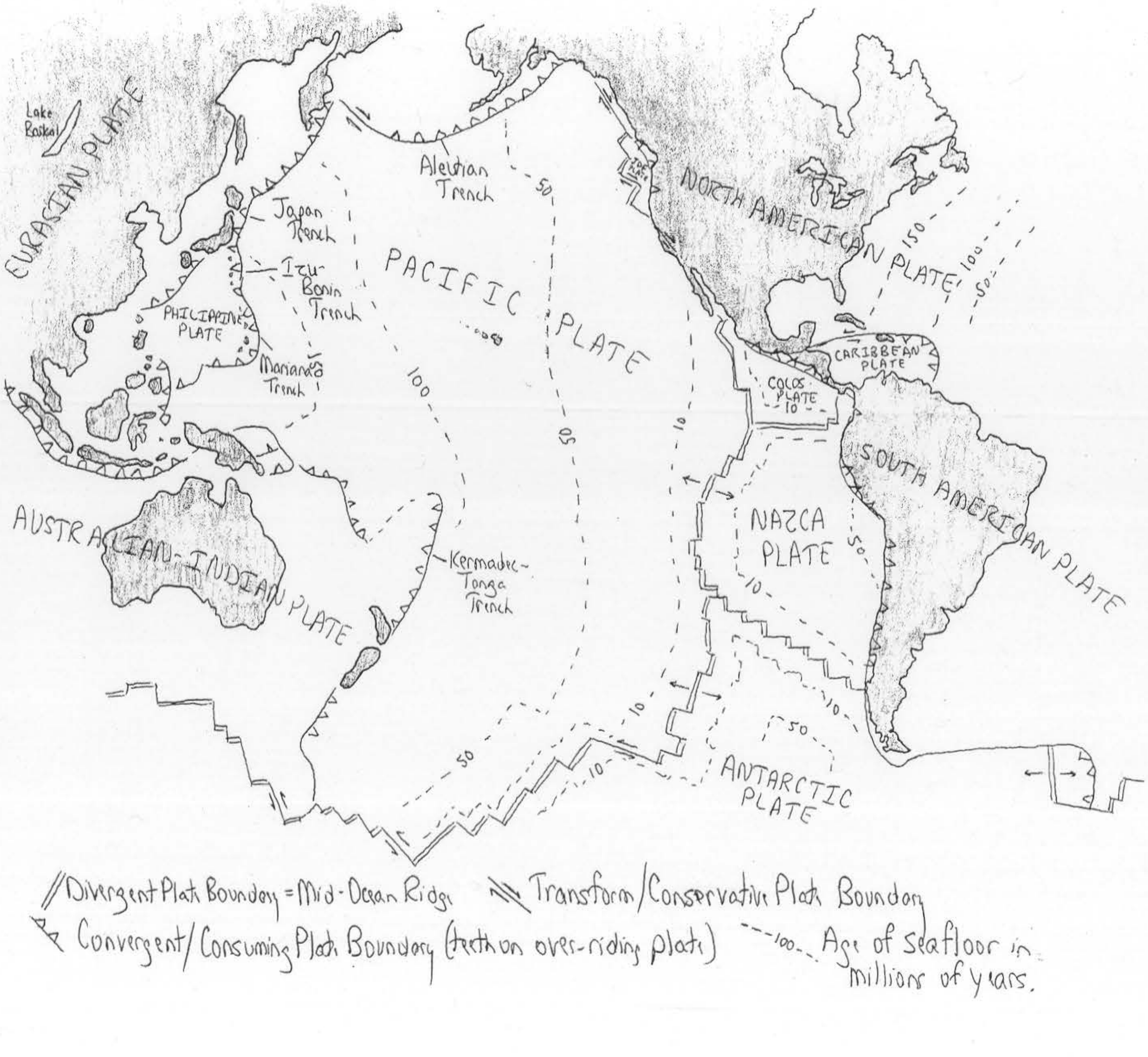

Figure 15-1: A tectonic map of the Pacific Ocean.

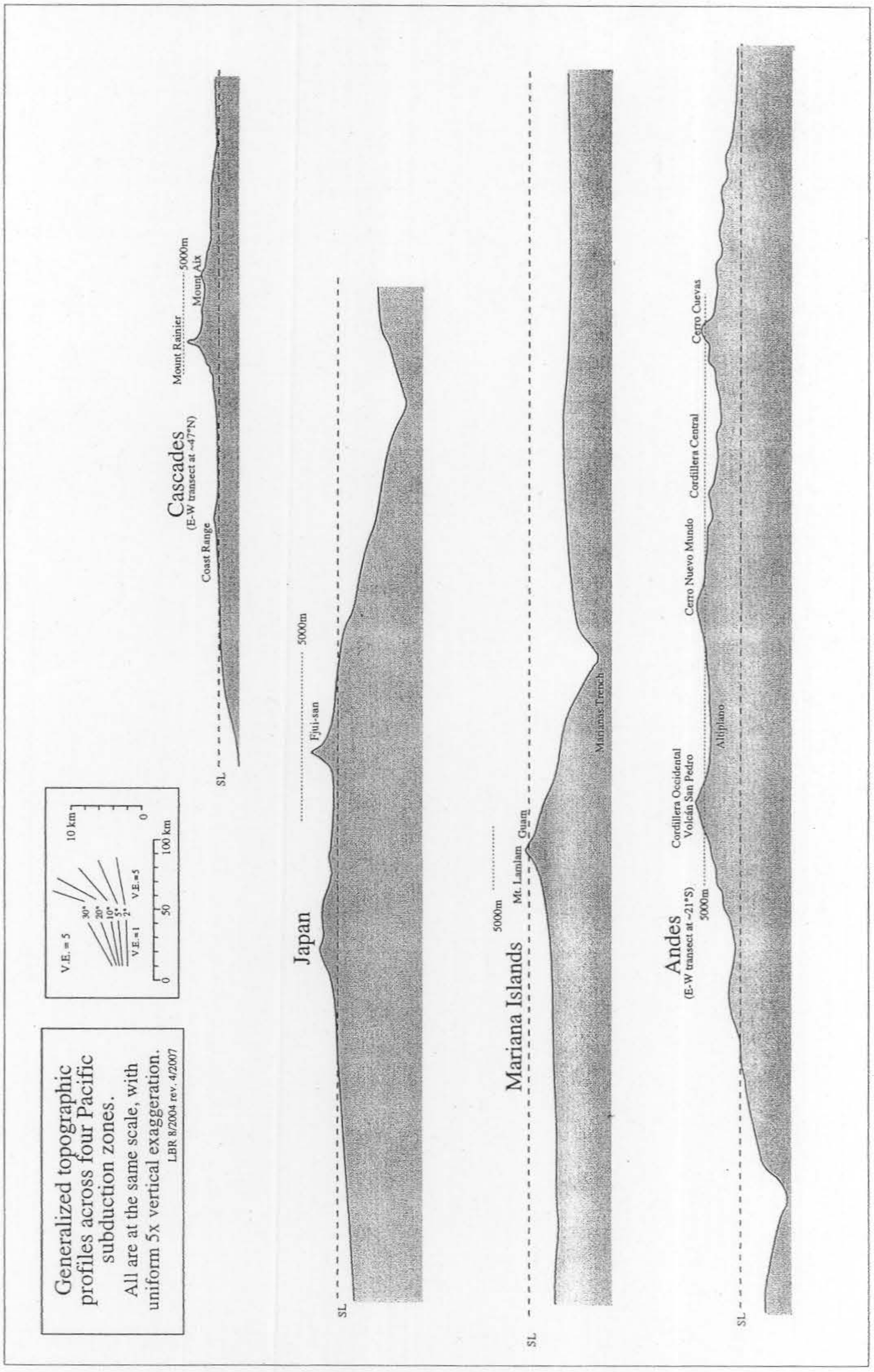

Figure 15-2: Generalized topographic profiles across four Pacific subduction zones.

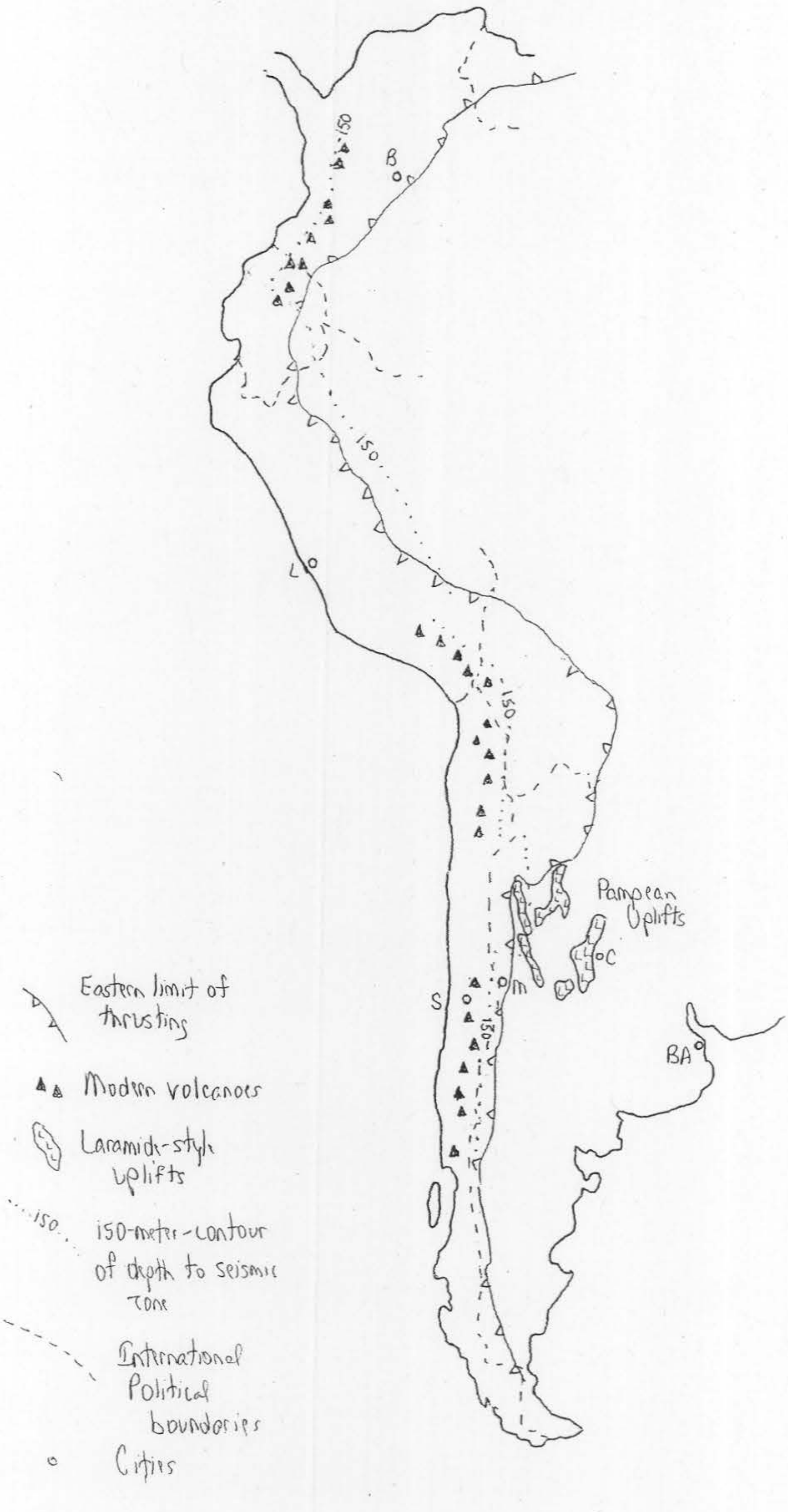

Figure 15-3: A map of geological features in the Andes.

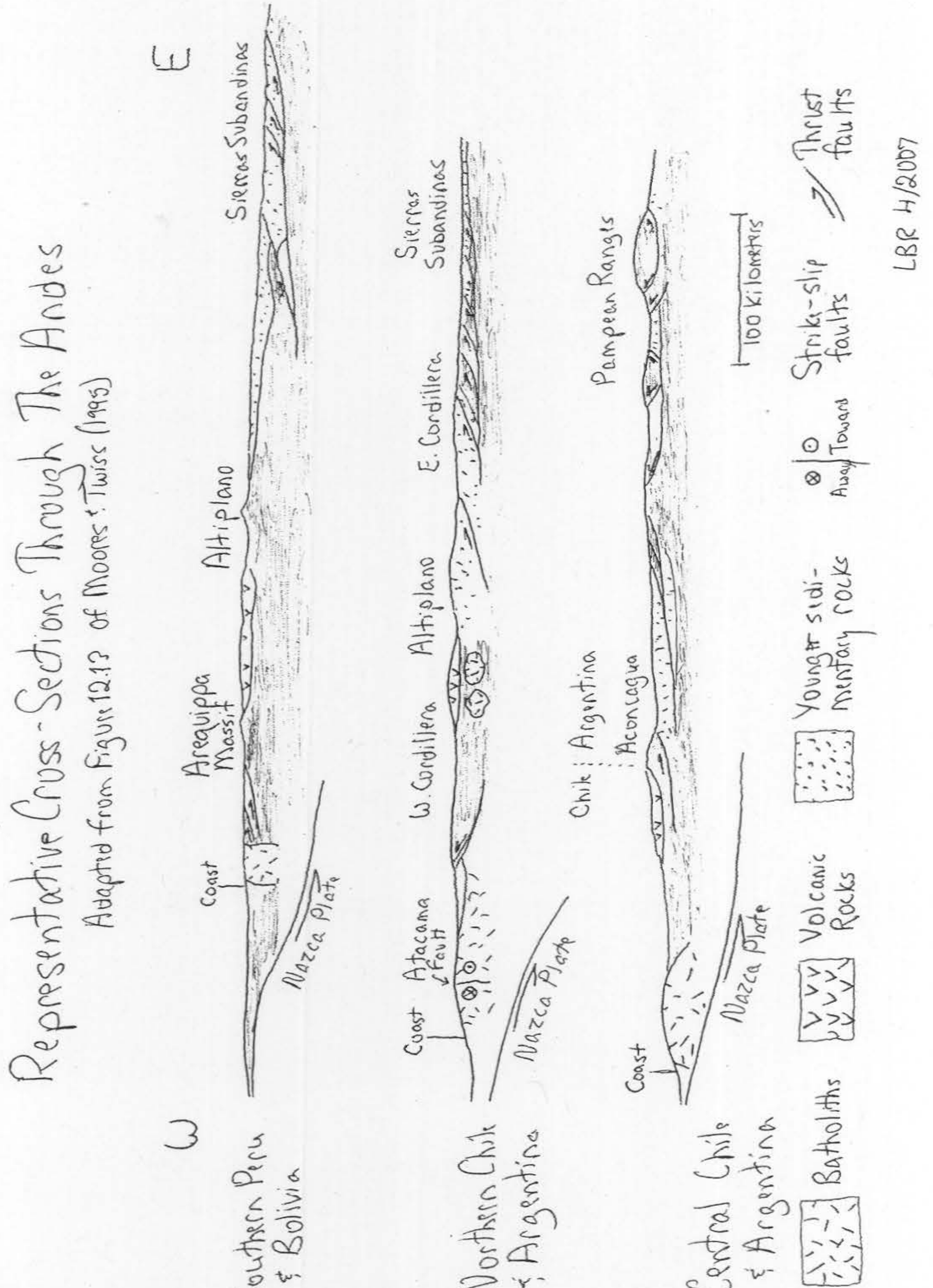

Figure 15-4: Three representative cross-sections through the Andes.

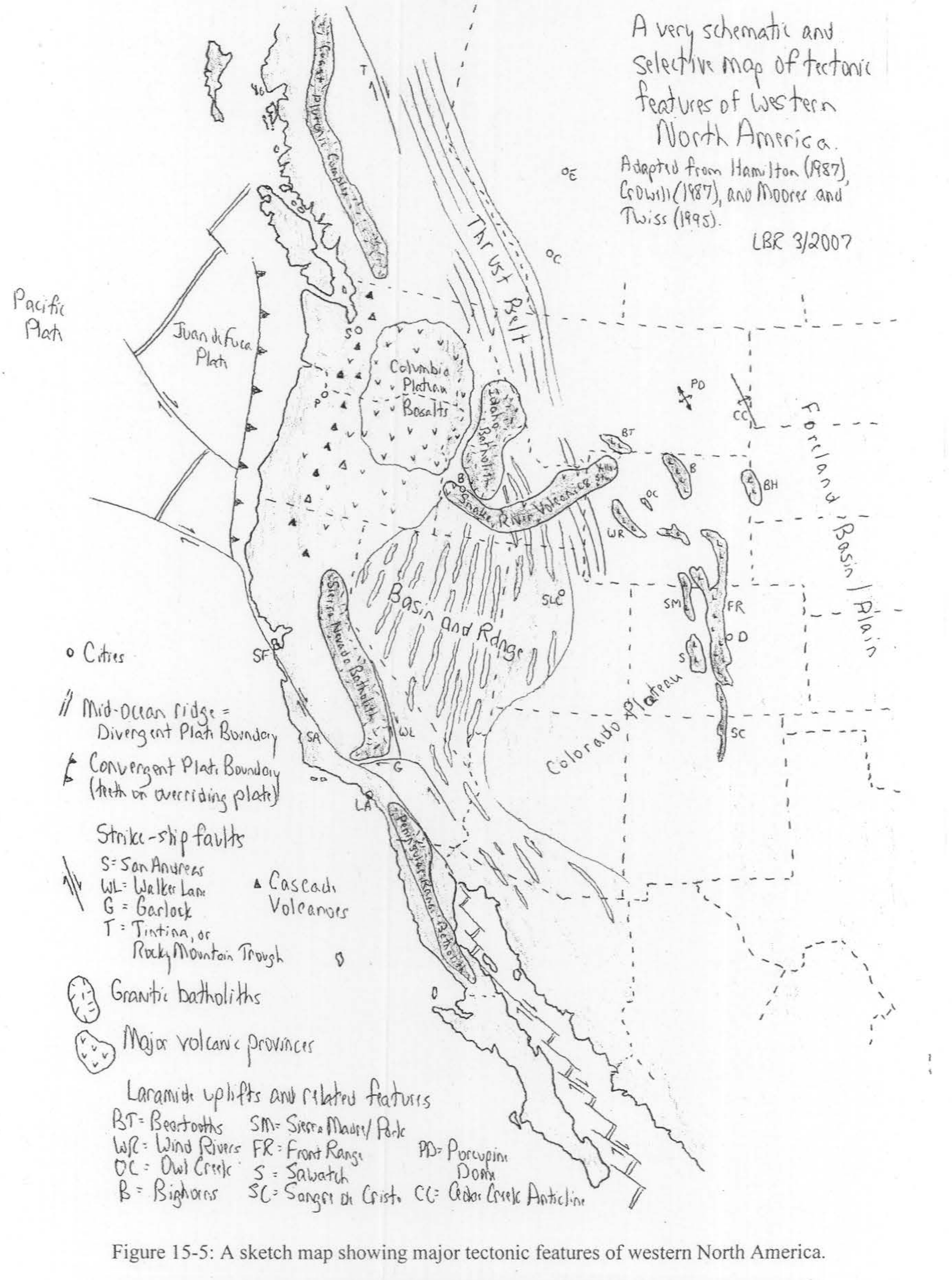

Figure 15-5: A sketch map showing major tectonic features of western North America.

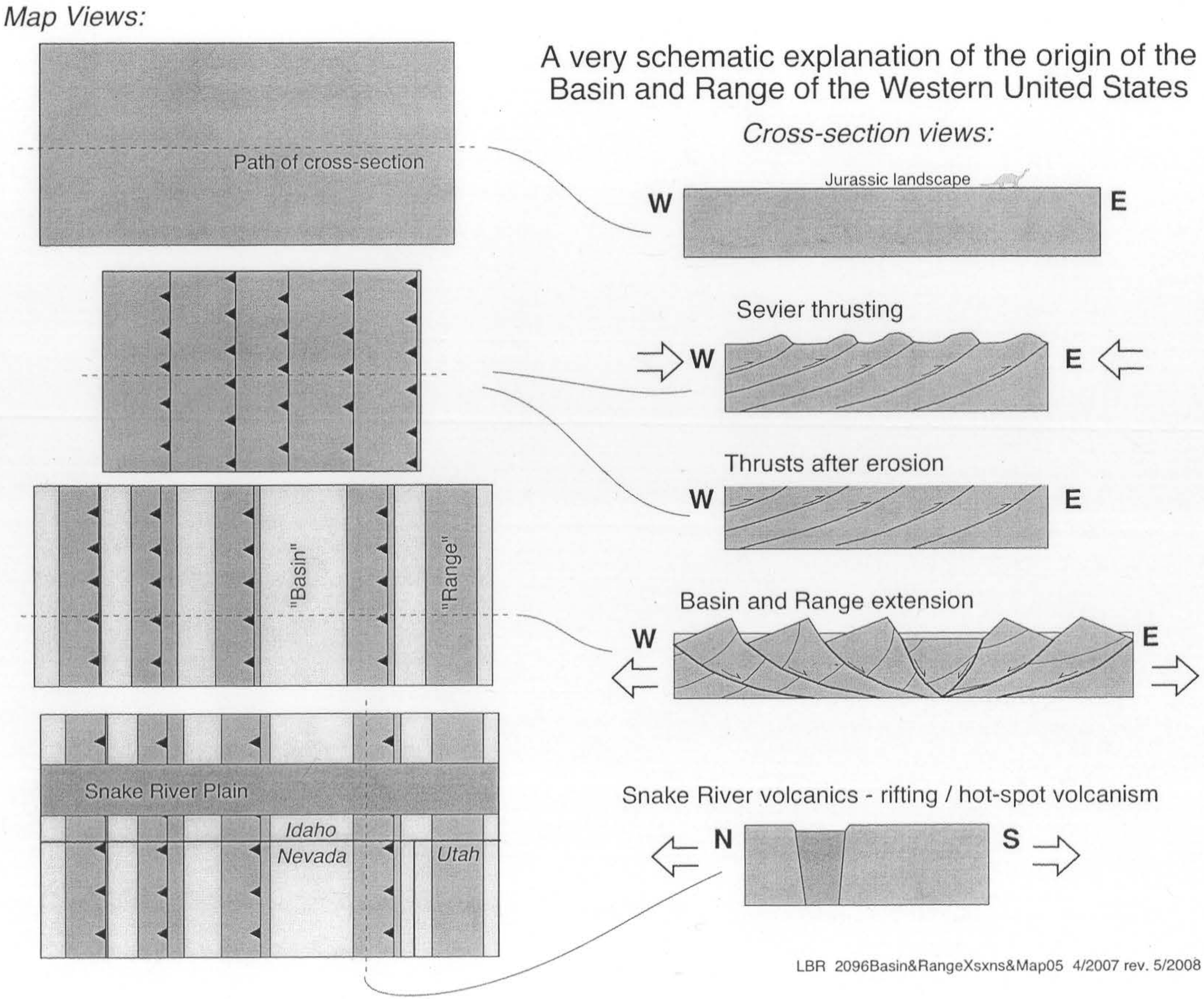

Figure 15-6: Maps and cross sections illustrating the Cenozoic history of the Basin and Range.

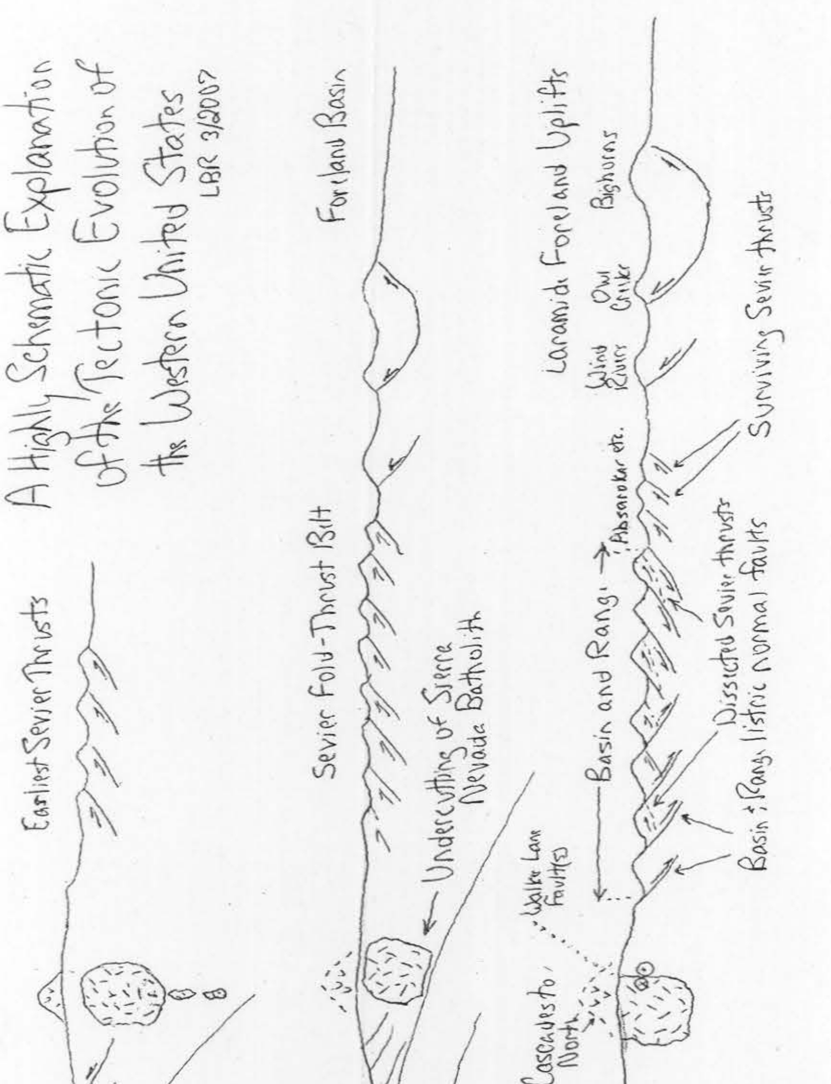

Figure 15-7: Three cross-sections illustrating the tectonic history of the western United States..

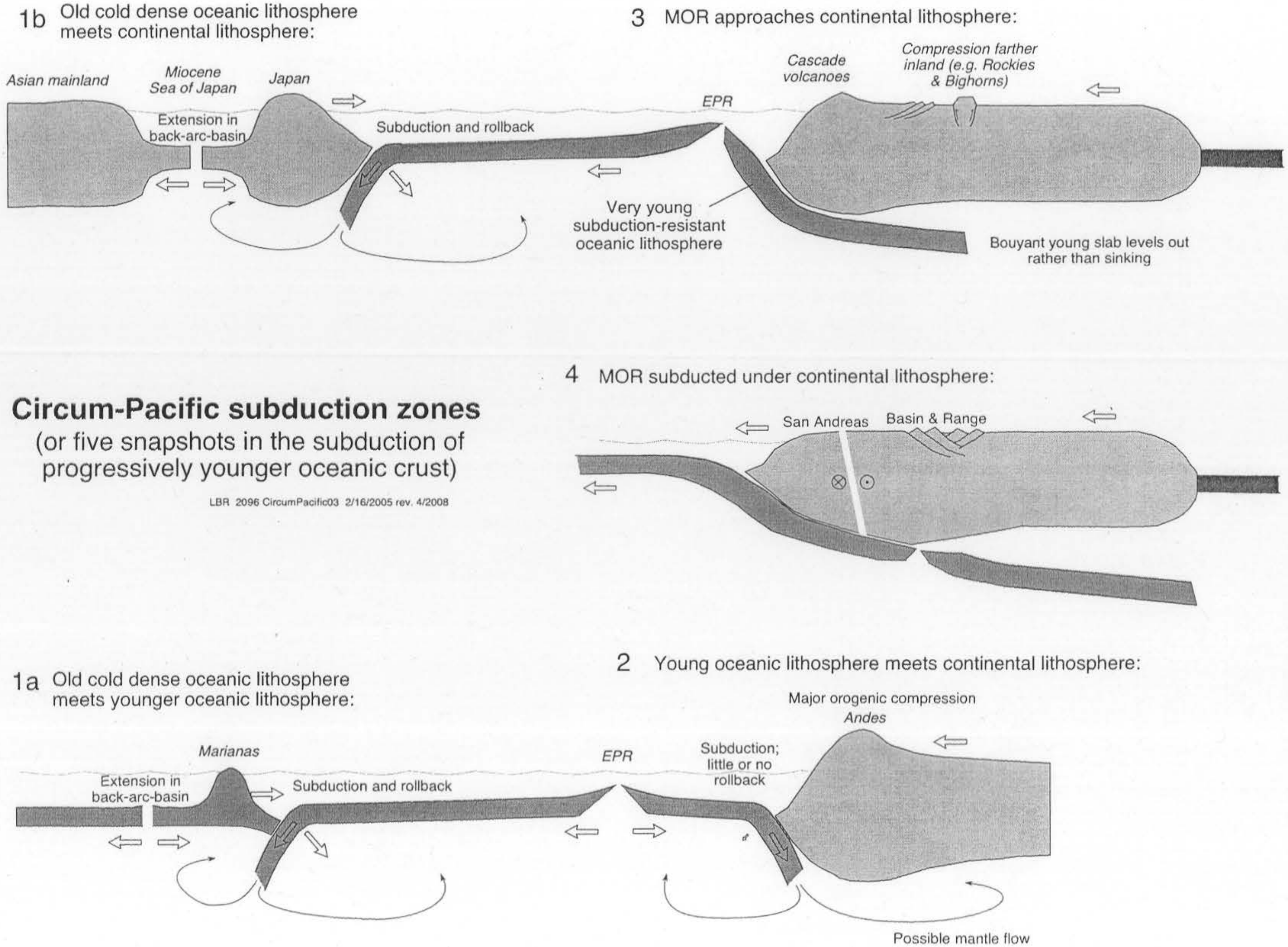

Figure 15-8: Five schematic cross-sections summarizing Chapter 15.
See also the later stages of Figure 9-7.

CHAPTER 16: GEOLOGIC GENERALIZATIONS ABOUT MOUNTAINS AND MOUNTAIN RANGES

This chapter seeks to reach generalizations from the previous seven chapters. One should perhaps be reminded of Oliver Wendell Holmes's generalization that "no generalization is worth a damn, including this one".

Mountains are sites of erosion and are thus sources of sediment, and thus they are inevitably flanked by plains or basins of sediments derived from those mountains (Fig. 16-1). In the case of thrusted mountains, we can often speak of a foreland basin that lies in the direction of thrusting and is a basin, a downwarped feature accepting sediment, because of the loading of thrust sheets. The modern Persian Gulf in front of the Zagros Mountains and the Pennsylvanian-age Appalachian Basin in front of (to the west of) the Appalachians are examples of foreland basins. In other cases, sediments accumulate simply at the margins of mountain ranges (as the Bighorn and Powder River Basins flanking the Bighorn Mountains, and all the basins flanking the ranges of the Basin and Range). In either case, accumulations of sandy and coarser siliciclastic sediments are the inevitable results of orogeny. On geologic maps, modern mountain ranges are thus flanked by areas of yellow, the areas of Tertiary to Quaternary sediments.

Across geologic time, such sediments may be the only evidence of mountains that have long since been eroded to obscurity. The study of the composition of sandstones has largely focused on the question of how one interprets ancient mountain ranges from the sands that survive long after the mountains are gone.

There are many ways to build mountains, as we have seen in Chapter 10, but collisions of continents after the closing of ocean basins are the events producing Earth's largest mountains on land. The reasons for this in part lie in the thickening of crust that results from the partial subduction of the leading margin of the approaching (Indian-type) continent. A general model can be seen at the top of Figure 16-2. In this general model, a suture zone is the surficial expression of the major thrust fault boundary between the two masses of continental crust. Both pro-thrusting (thrusting in the same direction as that along the suturing thrust) and retro-thrusting (thrusting opposite that direction) are possible, and they both can contribute to general elevation of the resulting mountains.

Comparison of the Alps and Appalachians would suggest that the most general model would involve uplift of high-grade metamorphics in the core of a mountain range (the Altkristallin and Piedmont, respectively) surrounded or led by a zone of less metamorphosed rocks (the quartz-phyllites of the Alps and the Blue Ridge of the Appalachians), in front of which would lie folded and thrusted marine sedimentary rocks (the Calcareous Alps and the Valley and Ridge), with a foreland of undeformed siliciclastics (the German Foreland and the Appalachian Basin). However, the presence of thrusted sedimentary rocks high in the heart of the Himalayas (in the Tethyan Himalaya) clearly brings into question how general such a model can be – unless one views the uplift of the High Himalaya as the anomaly that makes the Himalayas so anomalously spectacular.

Another feature common to the Alps and Himalayas is extrusion tectonics in the overriding plate. In both the Alps and the Himalayas, strike-slip faulting in the overriding plate to the north has allowed extrusion of continent to the east toward a convenient ocean basin (the Black Sea and the western Pacific, respectively). One might wonder why no similar feature developed in the Appalachian orogeny, and the answer may lie in the Appalachians' development during the collision of much continental area to make Pangaea, leaving no similarly convenient pathway to an ocean basin.

The positive topographic nature of any mountain range is mirrored by a thicker crust below, and that thicker crust allows rebound of the mountain range as it is eroded (Fig. 16-3). In a very simple model, the continental crust can be analogized to an ice cube riding on the "water" of the mantle, and isostatic equilibrium requires that the higher the ice cube, the greater its submerged extent. In continent-continent collisions, subduction of continental crust (as with Indian crust under the Himalayas) can double crustal thickness. The result is a deep mountain root. As the top of a mountain range is eroded, that root provides the buoyancy needed for the mountain range to rebound, so that the lifetime of a mountain range can be far greater than the time required to erode a thickness of rock equal to its elevation. However, although the heights of the mountain range may not diminish greatly through time because of this rebound, the geology at its surface will change as more and more deeply buried and intensely metamorphosed rocks are exposed.

Thrust faults are essential to the building of large mountain ranges, and they generally develop in the direction of thrusting (Fig. 16-4), even though it would theoretically be possible for thrusts to develop sequentially in the direction opposite that of the thrusting. This prograde direction of thrusting has been observed within the thrusts of the Valley and Ridge of the Appalachians, within the sequence from the Tethyan Himalaya to the Higher Himalaya to the Lower Himalaya to the Sub-Himalaya, and within the sequence of subductive thrusts interpreted in the eastern Alps.

Cross-sections with vertical exaggeration exaggerate the extent to which bodies of rock have been uplifted during orogenies. Most cross-sections of mountain ranges are drawn with a different vertical scale than horizontal scale, so that elevations appear greater and slopes appear steeper. The uplift of sedimentary rocks deposited on the seafloor to the peaks of mountain ranges appears improbable on such cross sections. However, calculations in fact show that the angle up which such strata moved was typically 1.0° to 2.0° and at most 3.0° (Fig. 16-5).

Mountain ranges are commonly the site of concentrations of rare minerals and ores and thus the locus of mining. Many orogenic processes contribute to the development of ore bodies in mountains: intruding magmas are fractionated to produce residua of incompatible elements; the heat of intrusions drives convection of groundwater that concentrates and precipitates certain elements; faults provide conduits for fluids moving upward from deeper in the crust and those fluids precipitate exotic minerals as they cool high in the crust; erosion and isostatic rebound expose deep crustal rocks. All of these concentrations of rarer elements and minerals are the focus of mining. Mining provides a certain romance to the cultural history of mountain ranges and drives local economies, but it also leaves behind huge spoil piles that are often sources of acid mine drainage and sources of other pollutants.

Tectonics builds *mountain ranges*, and gravity-driven surficial processes dissect those mountain ranges into individual *mountains* and slowly tear them down. We have to think about "mountains" from two general perspectives. First, the locations of mountain ranges and then general structure of mountain ranges are determined by tectonics: mountain ranges develop along, or in association with, plate boundaries. Furthermore, the nature of those boundaries determines the structure of the mountain ranges: whether they will be broken only along thrust faults, whether oblique subduction will cause strike slip faults and escape structures, or if plate motion will produce tension and normal faults. The lessons of the circum-Pacific and Himalayan orogens further show the overall height of a mountain range has much to do with plate tectonic history.

With that said, the second general perspective must be that the actual form of the mountains we see, and thus how we experience mountains as we live amongst them or pass through them, is mostly dictated by surficial processes and factors, such as climate, weathering, erosion, and mass wasting. Climate, for example, determines how much precipitation occurs and whether it is dominated by rain or snow. The resulting styles or precipitation determine what kinds of valleys cut through mountain ranges and thus define the form of the individual mountains. Climate also determines the position of tree lines, and so determines whether we traverse forested mountainsides or barren rocky slopes. Landslides commonly take away the whole side of a mountain, changing that mountain's form both with the gaping scarp left behind and the apron of debris that becomes the mountain's flank.

Thus the generation of mountains must be viewed from both of these perspectives, the longer-term and larger-scale tectonic perspective of position and structure of mountain ranges, and the shorter-term and smaller-scale perspective of surficial processes that shape the individual mountains and valleys and rivers within a mountain range. The two perspectives of course overlap, as when the location of a fault (a tectonic feature) becomes the locus of weathering and erosion and stream flow that becomes a river valley (a surficial feature).

Older deformations are commonly hidden within the most recent orogenic deformation of any mountain range. Within the Appalachians, it has long been recognized that there have been at least three orogenies, with the Pennsylvanian orogeny overprinted on those before it and deforming their deformations. Recognition of similar older deformations with more recent deformations, as noted in Chapter 13 regarding the Himalayas, will in the long run be the rule rather than the exception as we study mountain ranges further.

In part as a corollary of the above, one has to remember that **the concept of "a mountain range" is a dangerous thing**. A string of land at high elevation is conveniently viewed as a mountain range, but its geologic character almost certainly varies across its short direction (as evidenced by the zones discussed in previous chapters). It also is likely to vary along its length, so that one end of a mountain range may be very different than the other (compare the French/Swiss Alps to the Austrian Alps). Through time, an elevated land mass changes, either through repeated orogenic events or later through erosion and rejuvenation (compare the Ordovician Taconic Mountains with the Pennsylvanian Appalachian Range with the Cretaceous Appalachian Plain with the modern Appalachians). As with our attempts to put names on rocks, our attachment of names to orogenic masses gives us linguistic power but can lead us both to overgeneralizations and to artificial distinctions as we try to label mountains.

To say that "the concept of a mountain range is a dangerous thing" may seem discouraging. However, it also promises us that, as we grow bored with our generalizations about any one mountain range, we will be able to find new variation and new things to learn as our explorations continue. Bon Voyage!

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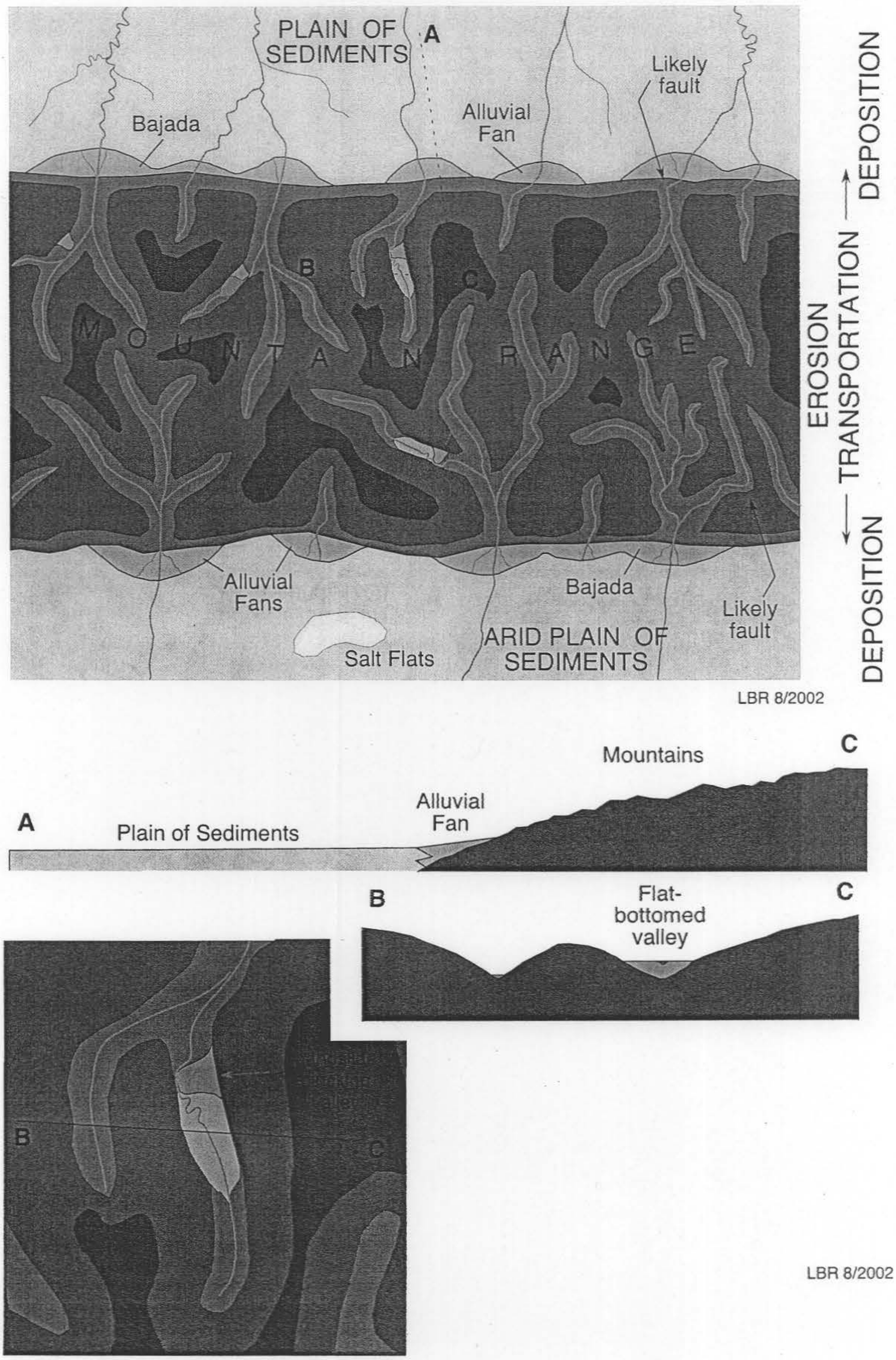


Figure 16-1: A general map of a mountain range, with cross-sections.

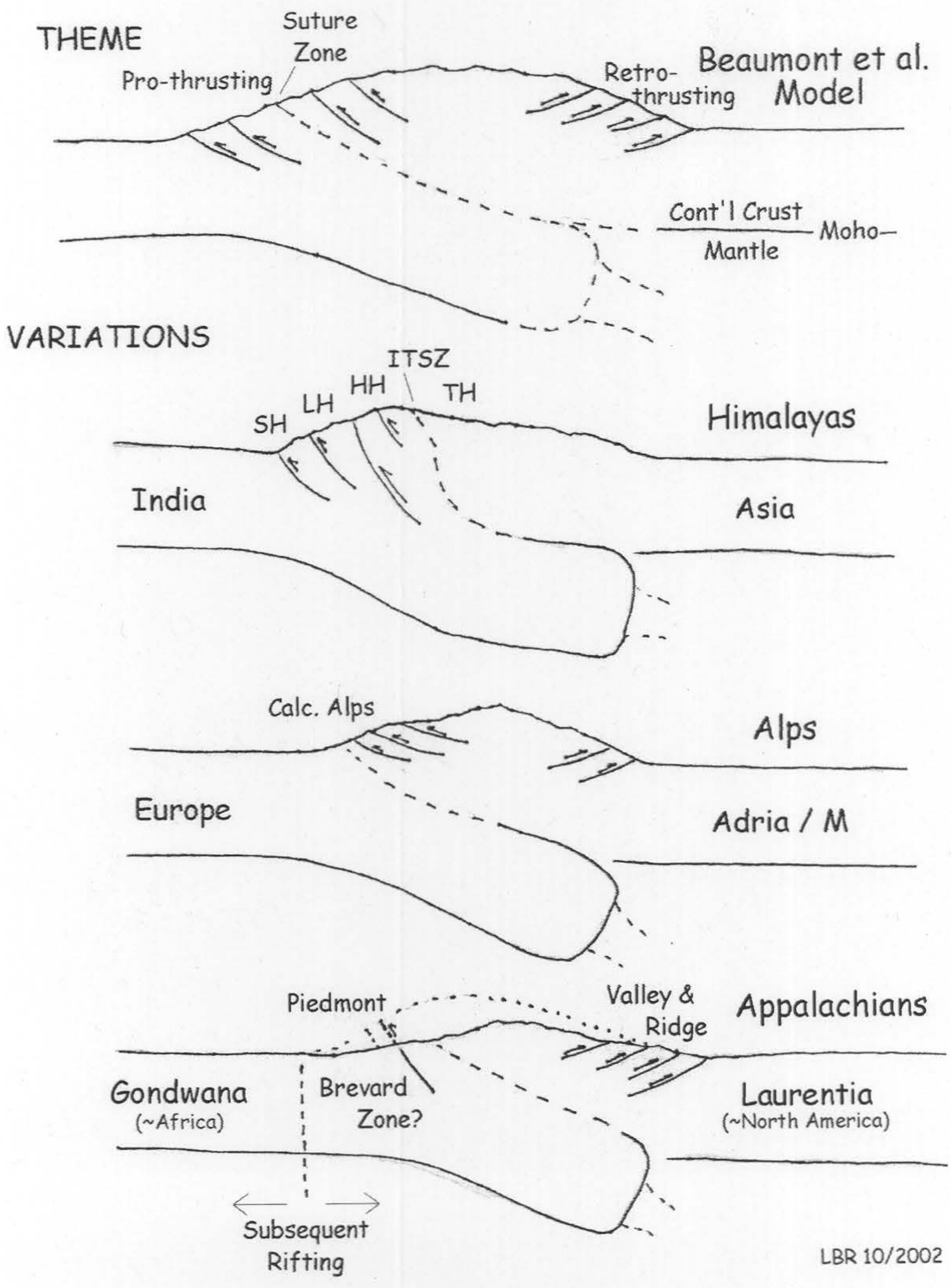
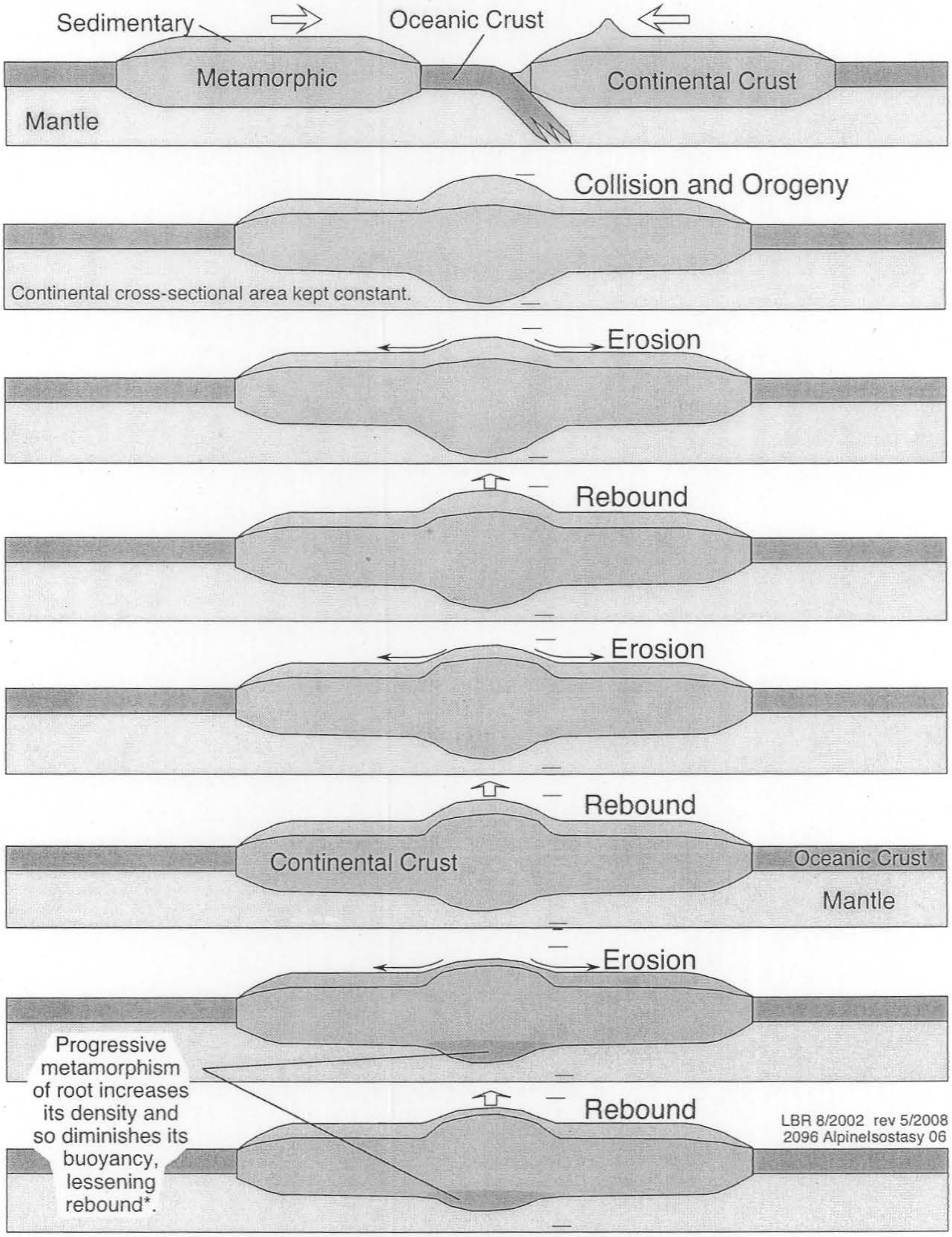


Figure 16-2: A general model ("Theme") and specific examples ("variations") of orogenies via collisions of continents.



* Karen M. Fischer, 2002, Waning buoyancy in the crustal roots of old mountains: *Nature*, v. 417, p. 933-936.

Figure 16-3: Cross-sections illustrating isostatic rebound of mountains with erosion.

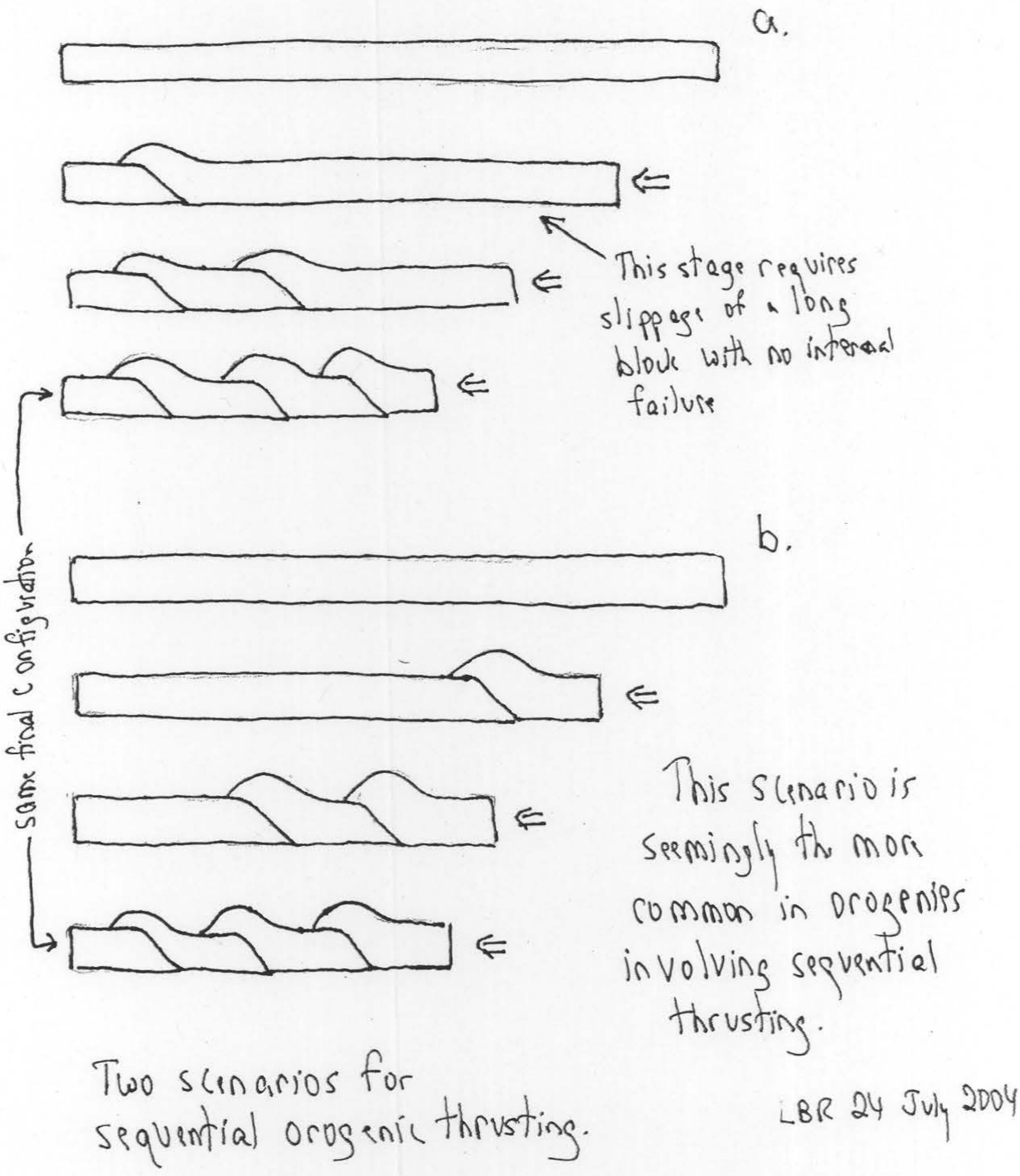
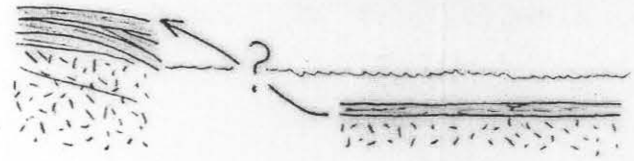
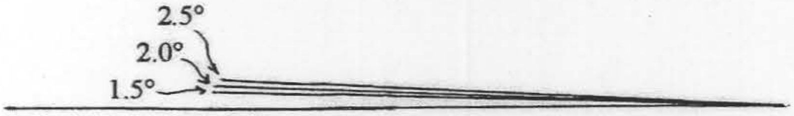


Figure 16-4: Possible sequences of thrusting.



Kilometers of uplift	*Kilometers of horizontal transport*	*Slope*	*Duration of orogeny (millions of years)*	*Rate of horizontal transport (cm/yr)**
4 + 3 = 7 present Alps elevation plus deep-sea depth	300 typical distance	1.34°	10	3.0
7 + 1 = 8 typical max. elevation plus shallow-sea depth	300 typical distance	1.53°	10	3.0
7 + 3 = 10 typical max. elevation plus deep-sea depth	300 typical distance	1.91°	10	3.0
9 + 3 = 12 Himalayan elevation plus deep-sea depth	500 reasonable Him. distance	1.72°	10	5.0
9 + 3 = 12 Himalayan elevation plus deep-sea depth	300 minimal Him. distance	2.29°	10	3.0
7 + 1 = 8 Typical max. elevation plus shallow-sea depth	200 Alps minimum	2.86°	10	2.0



*Present sea-floor spreading = 3 to 15 cm/yr.

Sketched crudely to scale:

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Figure 16-5: A table illustrating the scale and rate of continent-continent collision.

Part III. Glaciers and Glaciation

Chapter 17: Glaciers and Glaciation: Fundamental Principles

Defining and necessary characteristics

A glacier can be defined as a flowing mass of ice and rock on land.[96] It originates from snow, but the white flaky snow must be compressed into clear solid ice to be a glacier, and the solid ice must be under sufficient pressure to flow. A patch of snow high in the mountains that survives into the summer is not a glacier both because it is snow rather than ice and because it is a static unmoving mass. Sheets of sea ice are likewise not glaciers, most obviously because they are not on land but also because they do not originate from snow and they do not flow as the result of internal deformation.

For snow to survive long enough to be converted to glacial ice, and for enough snow to accumulate to exert the pressure to make the basal snow into flowing ice, the snow and resultant ice must survive from year to year. Such survival requires cool summer temperatures, to minimize summer melting, and abundant snowfall, so that at least some snow and ice survive through the summer melt-off. The coldness of winter is irrelevant, so long as temperatures remain below the freezing point of water; it is summer temperature that is critical. The greater the summer temperature, the greater the amount of snowfall must be for a glacier to exist, with a maximum summer temperature at the fern line (see below) of about 5°C (41°F).

The significance of abundant snowfall and cool summers to the survival of glaciers can be seen in three critical examples. One is the presence of glaciers on the top of Mt. Kilimanjaro in equatorial Africa, where both factors allow ice to survive at the equator.[97] A second is the greater frequency of glaciers on the southern rather than northern side of the Himalayas, because greater snowfall on the southern side results from transport of water vapor from the south. The third is the non-existence of glaciers in the dry valleys of Antarctica, where the lack of snow precludes glaciers despite a remarkably cold climate.

A glacier must be defined as mass of ice and rock, rather than just ice, because glaciers inevitably incorporate mineral material within their ice, with particle sizes ranging from clay-particles to single chunks of rock the size of large buildings. We will in fact find that glaciers can grade from ice-rich to rock-rich, and there are alpine bodies called "rock glaciers" because they are seemingly more rock than ice (Fig. 17-1).

The motion of a glacier depends in large part on the flow, or ductile internal deformation, of ice in the lower part of the glacier. Ice at atmospheric pressure is brittle and so breaks if mechanical force is applied to it – bonds between water molecules fail quickly across largely surfaces in such breakage. At higher pressure, however, water molecules or clusters of them move relative to each other, and this internal deformation at fine scale allows ice to deform in a ductile rather than brittle manner. The weight of the snow and ice at the top of a glacier generates sufficient pressure that the lower portions deform ductilely and flow (Fig. 17-1). A glacier can thus be divided into an upper brittle zone and a lower ductile zone. Crevasses, which are large and roughly vertical breaks in the ice, can develop in the brittle zone but can not extend into the ductile zone. Folding of layers of ice can take place in the ductile zone, and folded layers are only observed in the brittle zone when they have been exhumed by melting of the upper parts of the glacier.

Snow, firn, and ice

Snow is snow: loose to loosely packed tiny crystals of ice recently fallen as atmospheric precipitation. The density of newly fallen snow is typically 0.1 gm/cm^3 (hence the generalization that 10 to 12 inches of snow are required to equal an

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[96] There are, in addition, "salt glaciers" where rock salt erupts from the subsurface and flows across a arid landscape. The best-known examples are in Iran (see earthobservatory.nasa.gov/Newsroom/NewImages/images.php3?img_id=16435).

[97] One might more accurately say "where these factors formerly allowed the survival of glaciers", because global warming has nearly eliminated the glaciers atop Kilimanjaro.

inch of rain). Dense new snow can be as dense as 0.2 gm/cm^3 (Fig. 17-2).

Firn or névé is well-bonded snow that is older than one year (i.e., it has survived at least one melt season) . It has a density of about 0.4 to 0.8 gm/cm^3. Firn is thus a transitional form from snow to glacial ice.

Glacial **ice** has a density of about 0.83 to 0.92 gm/cm^3, the density of pure ice. The transition from snow to firn to ice can take as little as a few years, if the transition is accomplished in part by melting and refreezing. On the other hand, it can take as much as a hundred years if the transition depends solely on compaction of snow to ice.

Alpine and continental glaciation

Glaciers can be divided between alpine glaciers and continental glaciers. As the names imply, the most obvious distinction is in location: alpine glaciers are found in mountain ranges and in the valleys descending from them, whereas continental glaciation takes place across continents and reduces them to plains or basins. There is likewise a major difference in scale: alpine glaciers or ice fields are the size of high mountain areas or inter-mountain valleys and are perhaps hundreds of feet thick, whereas continental glaciation covers entire continents or at least large parts of them, and the ice is thousands of feet thick. There is a major difference in age of the ice: ice in alpine glaciers is typically a few hundred years old, whereas some ice in Antarctica today is at least 800,000 years old. The kinds of glaciation also differ in the forces driving their motion: the flow of alpine glaciers is driven by, or at least is aided by, gravity as they move downslope, whereas the flow of continental ice sheets is driven largely by the mass of ice at the center (much like a viscous liquid flowing away from one point at which it is being poured). Finally, the two kinds of glaciation differ in their effects on landscapes. Alpine glaciation typically sharpens and steepens landscapes, producing spectacularly sculpted mountains of which the Matterhorn is probably the most famous example. Continental glaciation, on the other hand, flattens landscapes, generating famously flat plains like those of northern Illinois and Iowa.

Movement of Glaciers

Although flow, as mentioned above, is critical to the movement of glaciers, at least three mechanisms for movement exist and probably work together or in sequence (Fig. 17-3).

Flow of ice in the ductile lower part of a glacier allows the brittle zone to ride along and allows the base of the glacier to remain frozen to an underlying umoving substrate. A mechanical model might be a tablet riding on a flowing layer of very viscous molasses on a sloping cold metal surface.

Slippage of ice can occur, or is more likely, when the bottom of the glacier is lubricated with water. This means that slippage is more common when meltwater is abundant, and thus typically in summer. Slippage may account for many surges of glaciers.

Deformation of an underlying substrate may allow motion of an entire glacier. This may be most common when the glacier is passing over a ground moraine (see below), which is in turn more common with continental glaciation than with alpine glaciation.

Because of the drag of the bottom of the glacier on the underlying substrate, flow of a glacier is typically slowest at the bottom (and at the sides of an alpine glacier), and fastest at the top (and at the center of an alpine glacier) (Fig. 17-3). Along the length of an alpine glacier, flow is typically fastest where the glacier narrows and slower where it is wider. These generalizations are the same as those for a river, and for the same reasons in terms of the behavior of a flowing substance.

Unique sediments and deposits

A last generalization that can be made about glaciation, both alpine and continental, is that it produces a distinctive sediment called till. Till is a mixture of pre-existing mineral material with unsorted particle sizes ranging from clay-sized particles to huge boulders. Most mechanisms of sediment transport lead to the sorting of particle sizes (as in the sorting of gravels from sands from silts by rivers), but glaciers readily carry all particle sizes within their ice and then deposit all that material together as till. The only geological materials readily confused with tills are landslide deposits, although landslides and mudflows typically produce deposits with more angular clasts and with less diversity of particle size. "Diamicton" is a useful purely descriptive term for a matrix-supported poorly sorted coarse sediment and is quite useful when one is unsure whether to use the genetic terms "till" or "mudflow" (see Chapter 5).

"Till" is thus the name of a material, whereas "moraine" is the name of a landscape feature consisting of till. Moraines commonly form at the margins of glaciers and thus are linear to arcuate ridges of till, or they form beneath glaciers (typically continental glaciers) and thus are layers of till (and sometimes they form from material atop glaciers and thus make a layer after the ice melts). Moraines and the till of which they are made are thus our principal evidence of the past extent of glaciers, and they reveal how important glaciation has been in making the alpine landscapes and higher-latitude landscapes on which many of us now live.

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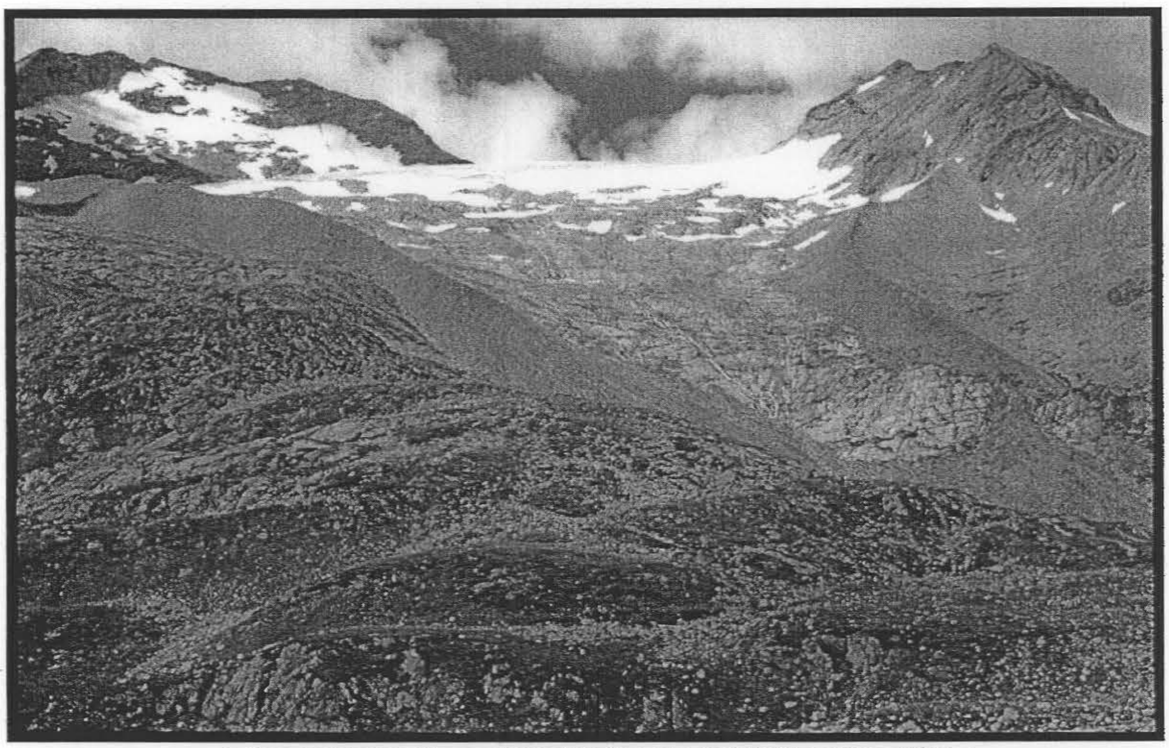
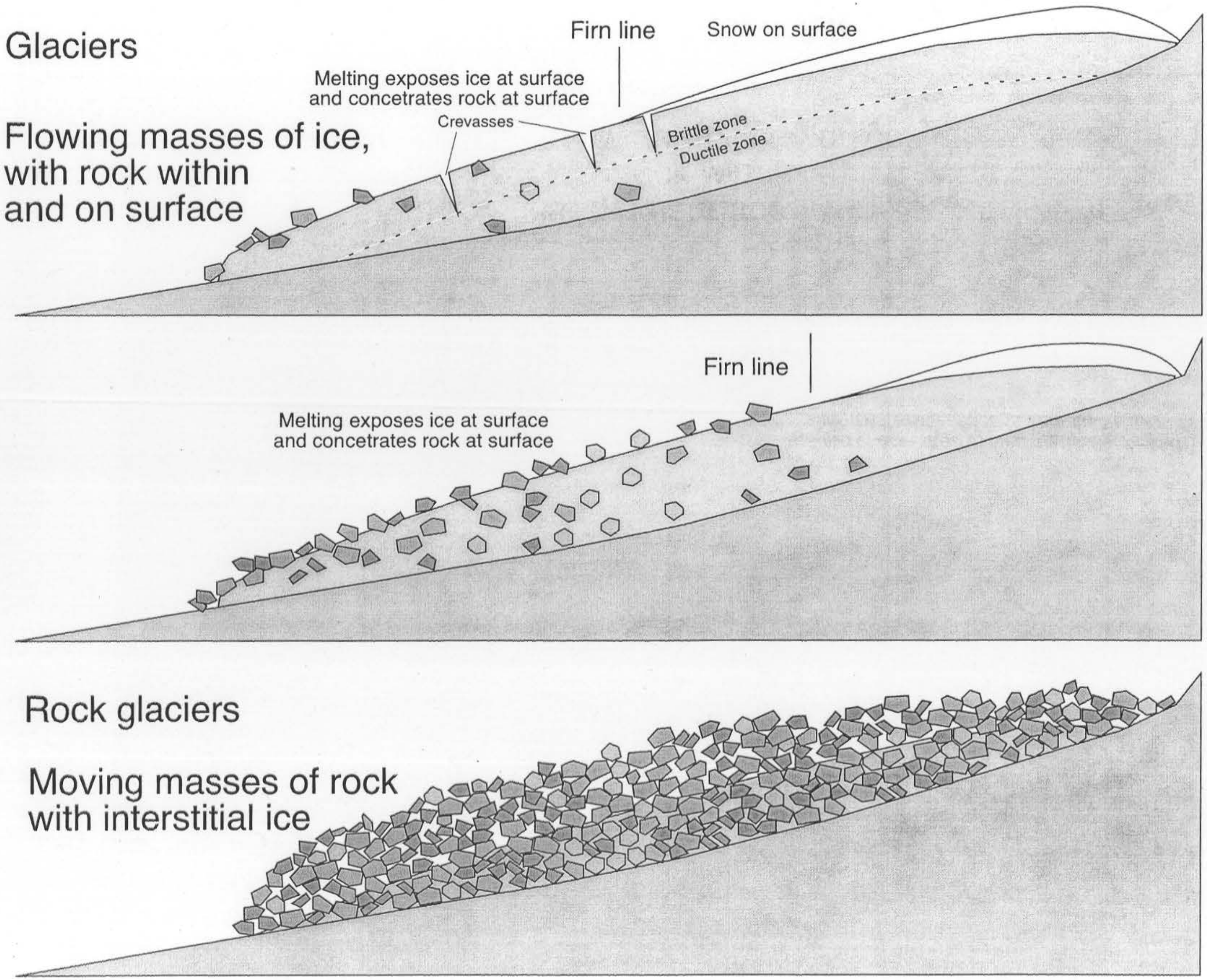


Lateral moraines and barren bedrock left by retreat of the Stampflkees Glacier above Schlegeisstausee in the uppermost Zillertal, above Mayrhofen, in Tirol.



Rock glaciers are more common in regions of little snowfall (e.g. eastern Cascades and Rockies)

Figure 17-1: Cross-sections illustrating some basic features of glaciers, and the spectrum from "normal" glaciers to rock glaciers.

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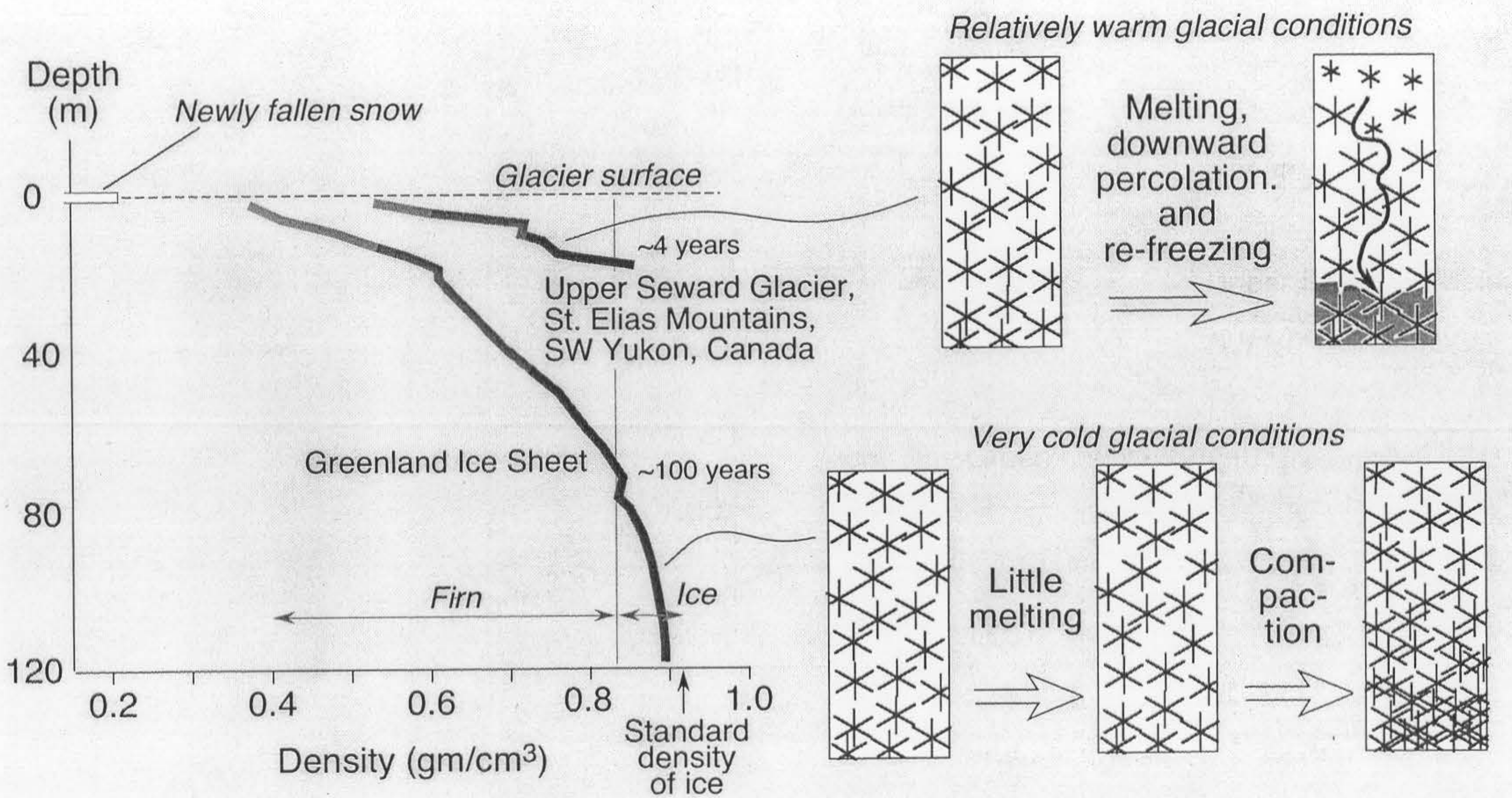

Firn: Snow that has survived one melt season and has begun transition to ice.

Based on Figure 2.3 of Benn and Evans, *Glaciers and Glaciation*

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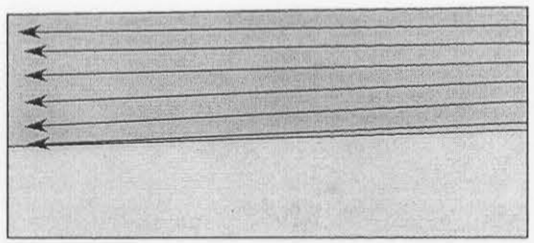
Fig. 17-2: The change from snow to firn to ice with depth and time.

Movement of Glaciers

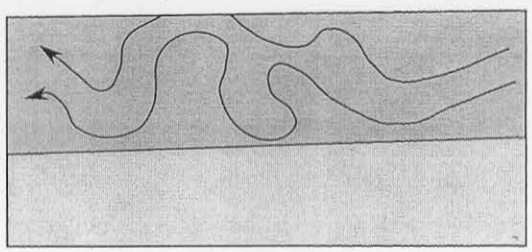
Mechanisms:

Sliding:

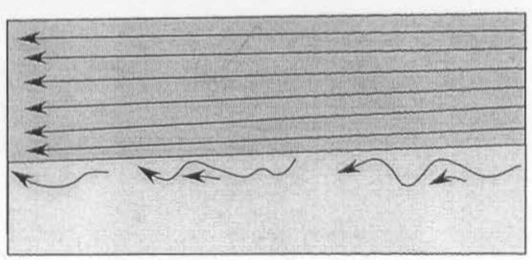


Slip is enhanced by water at base of glacier
(see graph at right)

Flow (deformation) of ice:



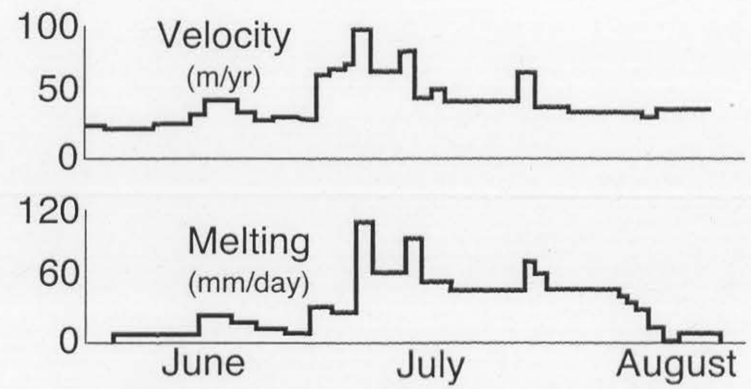
Deformation of underlying earth:



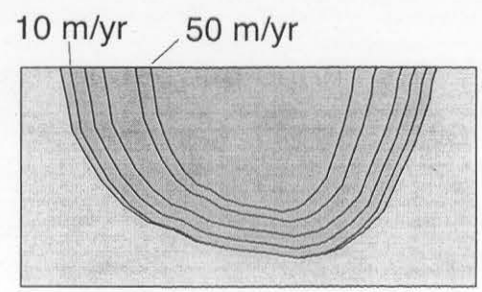
Longitudinal cross-section

Rates:

Typical	10-1000 m/yr	.05-4 in/hr
Surges	400-5000 m/yr	2-20 in/hr



White Glacier, Axel Heiberg Island
(Fig 4.25 of Benn & Exans)



Transverse cross-section
Athabasca Glacier -
Fig. 4.24 of Benn & Evans

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Figure 17-3: Mechanisms and rates of movement of glacial ice.

CHAPTER 18: ALPINE GLACIATION

Alpine glaciers exist in mountains at latitudes from that of Alaska and Antarctica to that of Mount Kilimanjaro at the Equator, and at elevations from sea level to the heights of the Himalayas. Glaciers in themselves are fascinating things, carrying ice, sediments, and artifacts for hundreds of years from alpine heights to the valleys below. However, even if one has no particular interest in glaciers, their effect on alpine landscapes should make them significant to anyone with an interest in mountains.

Incorporation of mineral material

Alpine glaciers accumulate mineral material, which will ultimately become glacial sediment, in at least four ways (Fig, 18-1). Two take place at the surface of a glacier and two at its base. At the surface, wind-blown dust accumulates by settling onto the surface. This happens year-round but leaves its most recognizable traces in the summer, when no snow dilutes it. The second source of accumulation at the surface is from rocks falling from the valleys at the sides of the glacier. The over-steepened valley walls (the sides of arêtes, as will be discussed below) are sufficiently unstable that rocks fall on a daily if not hourly basis, providing a constant supply of potentially large chunks that fall onto the glacier surface but are buried by subsequent snowfall.

Two sources at the base of the glacier likewise introduce fine and coarse mineral material into the glacial ice. Coarse chunks are produced by plucking, a process akin to the freeze-thaw mechanism of mechanical weathering in which water freezes into cracks in the rock, expands and thus opens the cracks farther, and ultimately breaks free pieces of rock that then travel with the ice. These plucked chunks, and perhaps chunks of rock generated by rockfalls, are then dragged along the bottom of the glacier. As they drag on the underlying bedrock surface, fine particles are ground or abraded from the both the dragged chunks and the underlying bedrock. This fine mineral material is commonly called "rock flour", and the grey-green color of the Inn is attributed to the transport of rock flour down the river system from glaciated highlands.

Crevasses and seracs

As the ductile zone of a glacier flows over the bedrock of a valley floor, its shape changes in response to the shape of the floor, much as a flowing river can have a non-flat surface as it passes over shoals. The overlying brittle zone, on the other hand, cannot deform so smoothly and instead breaks. The most common expressions of such deformation are crevasses, cracks in the glacier surface. Crevasses commonly form where a glacier flows over a bump or over an increase in valley steepness, so that the surface of the glacier is extended and breaks. Crevasses can extend only to the base of the brittle zone. In the ductile zone, any such opening would collapse on itself and be sealed back together.

The opposite effect occurs where a glacier flows over a decrease in slope and the surface of the glacier is compressed. This compression forces some of the brittle surface upward in a manner analogous to reverse faulting with the compression of the earth's crust. The result can be jagged protrusions of ice from the glacier's surface. These jagged spikes of ice are called seracs.

Zones of accumulation and wastage

The flow of a glacier down a valley means that it is transporting ice to the lower end of the glacier. The budget of ice in turn means that the glacier has an upstream end where snow must accumulate to make ice. That upper part of the glacier is called the zone of accumulation, and it can be recognized in summer because it will be white with the previous winter's snow (if the previous winter's show had melted, there could be no accumulation).

The downstream end of the glacier is the area where melting removes ice, and it is called the zone of wastage or the zone of ablation. This part of the glacier is dark in summer because the previous winter's snow has been removed and sediment-laden ice is exposed. This part of the glacier is in fact likely to be littered with rock across its surface, or even rock-covered. The extent of rock cover is a measure of how much ice has melted from the surface, stranding the rock that was present within that ice.

Melting on the surface can also produce meltwater streams on the surface of the glacier. Such streams commonly flow down through crevasses and flow under the glacier. Their water then either emerges from under the glacier at its toe or flows out as groundwater in the rocky debris under the glacier.

The boundary between the zones of accumulation and wastage, called the "firn line" or "equilibrium line", can change from year to year. A glacier could withstand a year or few years in which the firn line dropped completely off the glacier and only wastage occurred, but a series of such years would dictate the elimination of the glacier.

Form and length of alpine glaciers

The simplest model of an alpine glacier is that of one linear mass of ice moving out of a single cirque and down a single valley, with the toe within that valley (rather than beyond the mouth of the valley). However, many glaciers drain larger areas in their headlands, so that one can speak of "icefields" high in alpine areas (Fig. 18-2). A glacier draining a larger area presumably carries more ice and so may extend farther down-valley. Greater snowfall in the headland area also presumably allows flow farther down-valley.

At its lower end, a glacier may extend beyond the mouth of a valley and flow out onto a broader surface. Such glaciers are called "piedmont glaciers" because they flow onto the piedmont area below a mountain range in its strictest sense (Fig. 18-2). Well-developed examples not only flow onto the piedmont but expand to be wider than the valley from which they emerge. Piedmont glaciers presumably require a large supply of upland ice and either drain icefields or result from convergence of several upland glaciers. They also may drain areas of heavy snowfall. The Malaspina Glacier in Alaska is the most widely cited example of a piedmont glacier.

Erosional landforms

Alpine glaciers erode landscapes and thereby produce many different landforms. The following is partial list.

U-shaped Valleys are valleys cut by alpine glaciers, in contrast to the V-shaped valleys typically cut by rivers. The reason for the U shape is that the glacier can hold up straight valley walls that would collapse into a fluvial valley, and because the mass of the ice cuts downward across the entire width of a valley, rather than just at the center as would a river. Fjords are U-shaped values flooded by Holocene rise of sea level.

Arêtes are sharp ridges between glacial valleys. They develop as the straight sides of two Us converge to produce a knife-like ridge.

Cirques are bowl-shaped depressions at the upper ends of glacial valleys. They arise from intense plucking of bedrock at the glacier's upper end.

Tarns are small lakes filling cirques.

Horns are very steep three-sided or four-side peaks at the upper ends of three or four glacial valleys. A horn can be viewed as a mountain into which three or four cirques have been cut to sculpt its steep sides, or as the intersection of three of four arêtes.

Hanging Valleys are tributary glacial valleys that enter main or large glacial valleys and have floors much higher than that of the main valley. They form because the smaller glacier of a tributary valley cannot erode as deeply as the large glacier of a central valley. During glaciation, the tops of the glaciers will be at the same elevation, but after glaciation the elevations of the valley floors are very different. Waterfalls commonly develop at the mouths of hanging valleys.

Because erosion by alpine glaciers is so intense, alpine glaciers tend to steepen or sharpen landscapes greatly (Fig. 18-3). A pre-glacial landscape of rolling highlands and gentle valleys can be transformed into a rugged landscape of jagged arêtes and horns. The transition has implications both for the landscape itself (e.g., greater erosion and less vegetation) and for the isolation of the ecosystems and cultures that develop on it.

Depositional landforms

Depositional landforms of alpine glaciation are less impressive and perhaps shorter-lived than the erosional landforms, but they nonetheless provide important information about present and past glacial activity.

Lateral moraines are deposits of till at the edges of glaciers. Within valleys, they are pushed up against the valley walls. At the mouths of valleys, they extend downslope and may essentially be geomorphic depositional extensions of the erosional arêtes.

Medial moraines are masses of till within glaciers that form as lateral moraines merge at the junction of two glaciers. Medial moraines thus appear as dark bands parallel to the length of active glaciers. Medial moraines of former glaciers may be sharp-ridged masses of till sculpted by the ice that flowed on each side.

End moraines are deposits of till at the toe of glaciers. They are most likely to form when a glacier stabilizes its toe at a given location for a few years to deposit till at one place, and then retreats to leave that till as a distinct end moraine. Because much meltwater may pass through an end moraine, many end moraines are better sorted than lateral and medial moraines because their fine sediment has been at least partly flushed away. End moraines are also commonly completely breached by meltwater streams. However, end moraines may survive sufficiently intact to dam the lower ends of glacial valleys, creating glacial (really post-glacial) lakes.

Any one glacial valley may contain several end moraines, usually from the oldest low in the valley to progressively younger ones higher up. Where such moraines can be dated, perhaps by the radiocarbon method, they record times at which the valley's glacier stabilized to deliver large quantities of till before retreating up the valley.

Valley trains are strings of sand and gravel deposited downvalley from the toe of a glacier. Meltwater from the glacier erodes sand and finer gravel from the end moraine or off the glacier itself, and meltwater streams then deposit that material farther downvalley in the sand bars and gravel bars that define the braided pattern of most such streams.

Rock Glaciers

Any alpine glacier incorporates mineral material via the mechanisms described and carries such material with it down its valley. In most cases, we can still think of the glacier as a flowing mass of ice incorporating rock to varying degrees. However, in extreme cases no ice or snow is visible at the surface of an entirely rock-covered glacier, and the presence of ice is only inferred from the glacier-like flow of the mass of rock, and from outflow of meltwater with temperatures between 0°C and 1°C. These ice-cored but entirely rock-covered glaciers are called "rock glaciers" (Fig. 17-1).

The ice core in the middle of a rock glacier probably is sustained by refreezing of meltwater from the surface. There is considerable debate whether such refreezing within a talus pile is how such a core initially forms, or if the core begins as a conventional glacier in which the ice then shrinks to leave only rock at its surface.

Rock glaciers have the general form of conventional icy glaciers. Their most distinctive characteristic is a rumpling at their lower ends as rock still moving down the valley pushes up against non-moving rock at the toe of the glacier. The rumpled terminus is typically free of vegetation on active rock glaciers but is increasingly vegetated on fossil rock glaciers (rock glaciers from which the ice core has melted completely).

Rock glaciers have not been studied as extensively as conventional glaciers, but ongoing work suggests that they form in regions of drier and perhaps warmer climate than that of conventional glaciers (hence the lesser accumulation of ice). Most rock glaciers form in regions of less than 1500 mm annual precipitation as compared to as much as 4000 mm precipitation where some glaciers form. In North America, an example is the presence of conventional glaciers on the west sides of the northern Cascades but rock glaciers in the rain shadow on the east sides.

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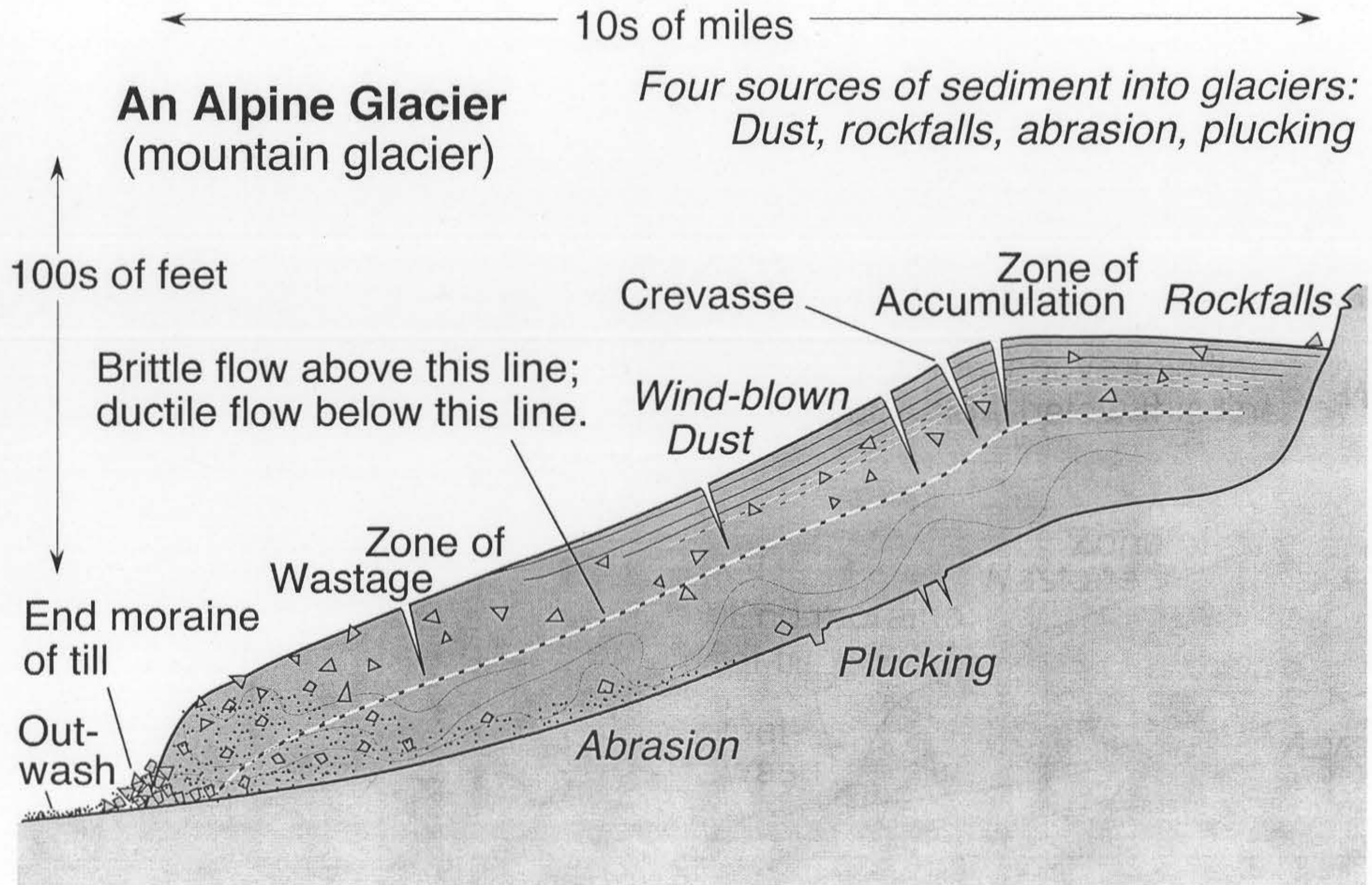

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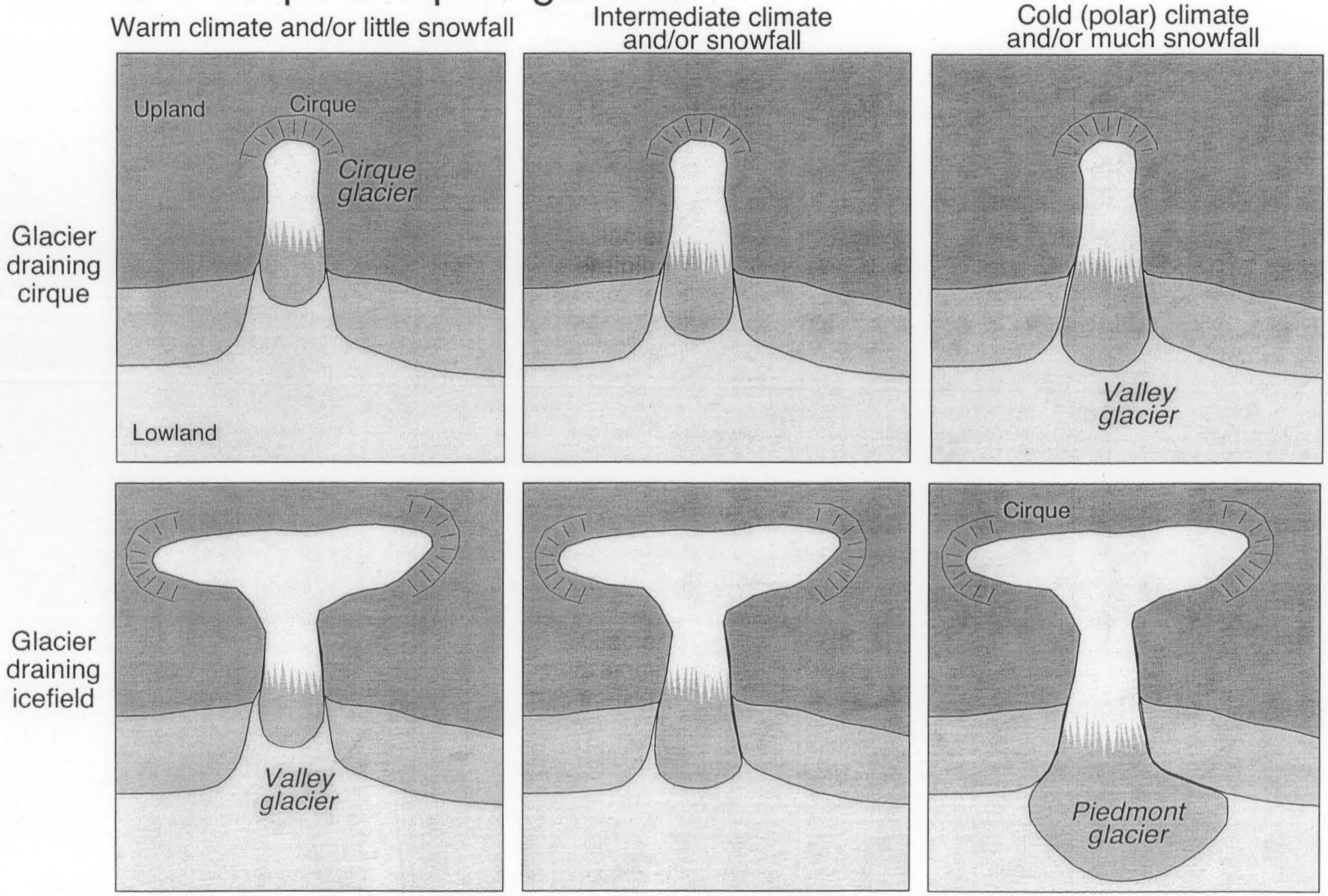
Drafted 11/20/01
rev. 4/2001

Figure 18-1: A schematic cross-section of an alpine glacier.

Size and shape of alpine glaciers



Note: This diagram is solely the author's interpretation of limited observations and should not be taken as well-documented or widely accepted concept.

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Figure 18-2. Size and shape of alpine glaciers.

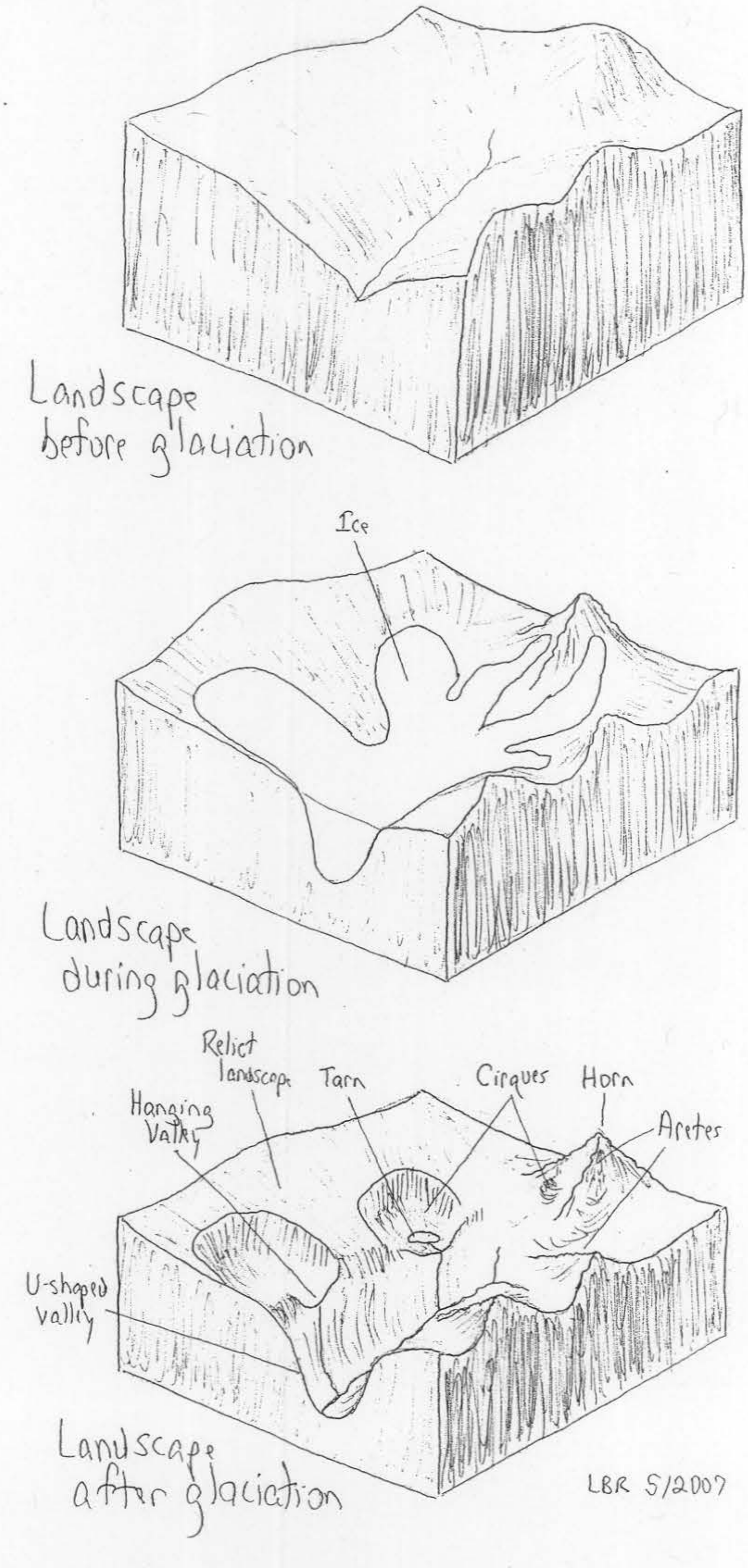

Figure 18-3: Evolution of landscapes with alpine glaciation, and resulting landforms.

CHAPTER 19: CONTINENTAL GLACIATION

On the modern Earth, the ice sheets covering Greenland and Antarctica account for 96.6% of the area covered by glaciers, which is about twice the area of the continental United States. Ice within these masses is as much as three kilometers deep, so that the proportion *by volume* of ice held in these continental ice masses is even greater. These two areas of continental-scale glaciation thus overwhelmingly dominate modern glaciation. However, continental glaciers have in the last two or three million years covered even greater areas, including most of northern North America and northern Eurasia, so that continental glaciation has taken on even greater significance in the relatively recent geologic past. This chapter therefore covers some basic concepts of continental glaciation (Fig. 19-1).

Scale

Continental glaciation[98] involves huge masses of ice moving radially away from a central area of accumulation. Flow is driven by the addition of mass in the middle, much as molasses will spread across a floor when poured at just one location. The edges of the ice mass develop where melting removes all ice or where the ice enters the sea and breaks off to form icebergs.

The two modern continental ice masses on Greenland and Antarctica are immense. In both cases, ice in the center is about three kilometers, or two miles, thick. In both cases, the weight of the ice is sufficient to depress the top of the underlying bedrock down to sea level. If the ice of Greenland and Antarctica could be removed, the bedrock would rebound hundreds, if not thousands, of meters as the overlying mass was removed.

Ice masses of similar thickness are believed to have covered northern North America and northern Eurasia during the last two million years. In North America, ice extended as far south as the modern Missouri and Ohio rivers, and in Europe it extended to about Amsterdam and Berlin. The presence of depositional and erosional features like those discussed below is the evidence for that conclusion.[99]

Depositional features (Fig. 19-2)

"**Drift**" is a very general and now outmoded term for all glacial sediments, regardless of particle size, sorting, or stratification. It arose from the assumption that such sediments were deposited from icebergs drifting across the world during the Noachian flood.

An **erratic** is a boulder, often quite large, that consists of a rock type different than that of the underlying bedrock. Erratics are thus evidence of glacial transport of rock over large differences. In the Midwestern U.S., a region of sedimentary bedrock, most erratics are igneous and metamorphic rocks that can only have come from the Canadian Shield, a large area of igneous and metamorphic rock in Ontario and Quebec.

An **end moraine** is a mass of till deposited at the downstream end of a glacial mass. End moraines are typically tens of feet tall, hundreds to thousands of feet across, and tens to hundreds of miles long. End moraines exist as long arcs across Iowa, Illinois, Indiana, and Ohio and suggest that ice flowed out as huge lobes tens to hundred of miles across.

Ground moraine is till deposited under a glacier as it flows over an area, or perhaps from the

[98] The expression "continental glaciation" will be used here, if sometimes awkwardly, because the expression "continental glacier" implies a distinct single ice mass flowing in one direction and thus fails to characterize the immense radiating mass of ice, thousands of times larger than any alpine glacier, that develops when a continent undergoes continental glaciation. "Ice sheet" is the usual term for one continuous continental-scale mass of ice, and "lobe" is used for a flowing mass protruding from a sheet. For example, the Des Moines Lobe was part of the Laurentide Ice Sheet. Any one lobe would have a volume of ice thousands of times that of an alpine glacier.

[99] Our understanding of the extent of Quaternary glaciation began with the recognition by Jens Esmark in 1824 that glaciers in Scandinavia had been much larger to account for the erratics and fjords there. By the 1830s, Jean de Charpentier concluded from striations, erratics, and moraines in Switzerland that glaciers had been much more extensive. Charptentier convinced an originally skeptical Louis Agassiz (1807-1873) of the significance of past glaciation. Agassiz went on to be a professor at Harvard and correctly recognized widespread evidence of glaciation across North America. He is thus commonly viewed as the principal nineteenth-century proponent of the "glacial theory". However, he also argued that Brazil had been completely covered by glaciers, a claim that weakened his credibility.

top of a glacier as it melts out. Ground moraine is typically tens of feet thick and extends over hundreds of square miles. Ground moraines and end moraines cover many areas of the Midwestern U.S. so thoroughly that no bedrock is exposed for tens of miles, and quarries into bedrock are rare features from which trucks haul crushed stone great distances.

Outwash is sand and gravel carried by meltwater and deposited in stream channels on the plain in front of a continental glacier. Outwash plains can extend across the entire margin of a glacial lobe and tens of miles outward from the margin. Outwash plains can be easily recognized on topographic maps by an abundance of gravel pits.

A **drumlin** is an elongate asymmetrical hill of till with its long axis parallel to the direction of ice flow, a steep slope on the upstream end, and a gentle slope on the downstream end. Drumlins seeming form as the result of ripple-like perturbations in the flow of ice, and they commonly occur in fields of multiple drumlins. This suggests a widespread cryodynamic turbulence during flow of glacial ice analogous to turbulence in flowing water that generates ripples in streambeds.

An **esker** is a sinuous ridge of sand and gravel. Eskers form via deposition of sand and gravel in meltwater channels that develop under a glacier.

A **kame** is a deposit of sand and gravel sorted by flowing water that was in direct contact with glacial ice. Some kames are conical hills of sand, or sand and gravel, formed when meltwater lakes developed on top of glacial ice (i.e., supraglacial lakes) and melting of the ice dropped the sediment to the land surface. Kame deltas form where a meltwater stream enters a proglacial lake. Kame terraces form where meltwater streams flow between a glacier and an adjacent land mass (for example, a valley wall). Moulin kames form where a meltwater stream drops through a glacier to its base and deposits sediment there.

A **kettle** is a pond or lake that formed as the melting of an ice sheet left isolated blocks of ice that melted to leave depressions in which a pond or lake formed.

A **dropstone** is a rock rafted by an iceberg into a lake or (more significantly) the ocean, and then dropped to the lake floor or sea floor as the iceberg melts. They are readily recognized because they are much larger than the fine particles of lake or ocean sediments. Dropstones are historically significant because they can provide evidence of glaciation on land after all evidence on land is eroded, and because their latitudinal distribution through time can be an indicator of the warmth or coldness of global climate.

Erosional features (Fig. 19-2)

Striations are parallel linear scratches or gouges in bedrock surfaces over which a glacier has moved. The scratches are made by chunks of rocks dragged by flowing ice across the underlying bedrock. Striations are significant both as evidence of glaciation and as indicators of directions of the movement of ice.

A **roche moutonee** is a chunk of bedrock that has a gently curved surface on one end and a nearly vertical surface on the other end. A roche moutonee is sculpted by glacial ice, in that ice melts as it impacts the curved upstream end and refreezes on the downstream end, freezing into cracks in the rock and dislodging chunks to produce the nearly vertical surface on the downstream end. Roche mountonees are thus indicators of the direction of flow of ice.

Finger Lakes are parallel elongate lakes gouged by flowing glacial ice. The Finger Lakes of western New York state between Rochester and Syracuse are excellent examples[100].

Glacial geographic features of North America

As was noted above, deposits from continental glaciation are found across northern North America south to the modern Missouri and Ohio rivers. A few deposits or areas deserve special note and are discussed in the hodge-podge below.

The **Canadian Shield** is, from a geological perspective, a region of Precambrian igneous and metamorphic rocks in eastern Canada. From a glaciological perspective, the Canadian Shield is the area from which most ice flowed during Quaternary glaciation of North America. These two ways of viewing the Canadian Shield seem to be related, in that evidence from tills across North America suggests that the source area for tills and erratics was originally covered much more with

[100] The Finger Lakes are, from west to east, Conesus, Hemlock, Canadice, Honeoye, Canandaigua, Keuka, Seneca, Cayuga, Owasco, Skaneateles, and Otisco Lakes.

Paleozoic sedimentary rocks but eroded through time until those rocks were gone and only a landscape of underlying Precambrian igneous and metamorphic rocks was left.

Glacial Lake Missoula was a large lake dammed in western Montana by a lobe of the Cordilleran ice sheet between 15,300 and 12,700 years ago. At its maximum extent the lake extended east 200 miles and contained more the 500 cubic miles of water. The lake repeatedly broke through the ice dam to send jokulhlaups (the Iceland term for huge floods of meltwater) across northern Idaho and eastern Washington. The flow of water during the largest of these floods is estimated to have been about ten times the flow of all the world's present rivers combined. The floods generated the Scablands that cover much of eastern Washington, eroding hundreds of feet of soil and sediment in a few hours. The Scablands feature fluvial channel features (ripples, point bars, etc.) at scales hundreds of times of those of analogous features in normal rivers.

Glacial Lake Agassiz filled the basin of the modern Red River in eastern North Dakota and the basin of Lake Winnipeg in Manitoba. The Red River basin drains north, and Lake Agassiz was thus trapped in front of the retreating Quaternary ice sheet. Lake Agassiz covered about 110,000 square miles (larger than all the Great Lakes combined), was nearly 700 miles long, and at its maximum width in Manitoba was probably more than 250 miles wide. Its greatest depth was more than 700 feet. Its shorelines and deltas have been mapped across Minnesota, North Dakota, and Manitoba. The bed of Glacial Lake Agassiz remains so flat that the Red River plain remains particularly prone to flooding.

Minnesota, the "Land of Ten Thousand Lakes", actually has something more like 15,000 lakes or ponds, almost all of them kettles or pockmarks in the Quaternary till that covers most of the state.

Of the Great Lakes, at least **Lake Michigan and Lake Erie** were in part gouged by lobes of ice that flowed along their lengths and out onto the plains at their ends. This claim is supported by the presence of concentric arcs of moraines around the south end of Lake Michigan and the southwest end of Lake Erie. The locations of the lakes was dictated by the presence of easily eroded shale in the underlying bedrock, but the lowlands of that easily eroded rock provided an avenue for flow of ice and thus further erosion.

The **Des Moines Lobe** is a semi-circular area in northern Iowa (and southern Minnesota) onto which a late Wisconsinan (late Pleistocene) glacial lobe flowed and left a strikingly flat landscape. The term is applied both to the lobe-shaped area of glacial till and to the lobe of ice that deposited it.

The **Coteau des Prairies** (the "slope of the prairie") in eastern South Dakota is a north-pointing topographic promontory (or at least a promontory by South Dakota standards). A topographic low to the north splits into two troughs around the Coteau des Prairies. Those troughs were the paths of flowing ice, one of which leads to the Des Moines Lobe.

The **Driftless Area** is an area in southwestern Wisconsin and adjoining southeast Minnesota, northeast Iowa, and northwest Illinois with no glacial deposits. Despite their many advances, Quaternary glaciers seemingly never entered this area. It thus remained free of drift and retained a more rugged topography than that of the surrounding glaciated region.

Finally, one little-known but spectacular feature of North American glacial geology is **a 13,500-ton erratic** in southeastern Ohio. The erratic is located along the north fork of Olive Branch, a tributary of the Little Miami River near Oregonia, Ohio. It is a block of Silurian-age Brassfield Limestone sitting on top of Illinoisan-age till. When deposited, it had an area of 45,000 square feet (more than an acre), a thickness of five to as much as 17 feet, a volume of more than 225,000 cubic feet, and a weight of about 13,500 tons.[101] These dimensions must be stated in the past tense because the erratic was so large that a quarry was excavated in it, and the quarry removed enough rock to separate the original block into three pieces.

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[101] Wolford, J.J., 1932, A record size glacial erratic: *American Journal of Science,* Fifth Series, v. 24, p. 362-367. UGA Geology Professor Steven M. Holland, an Ohio native and famed stratigrapher of the Ordovician and lower Silurian, is thanked for bringing this piece of the literature to the author's attention.

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Quaternary glacial till near Iron Mountain, Michigan. Note boulders hanging from cliff face.

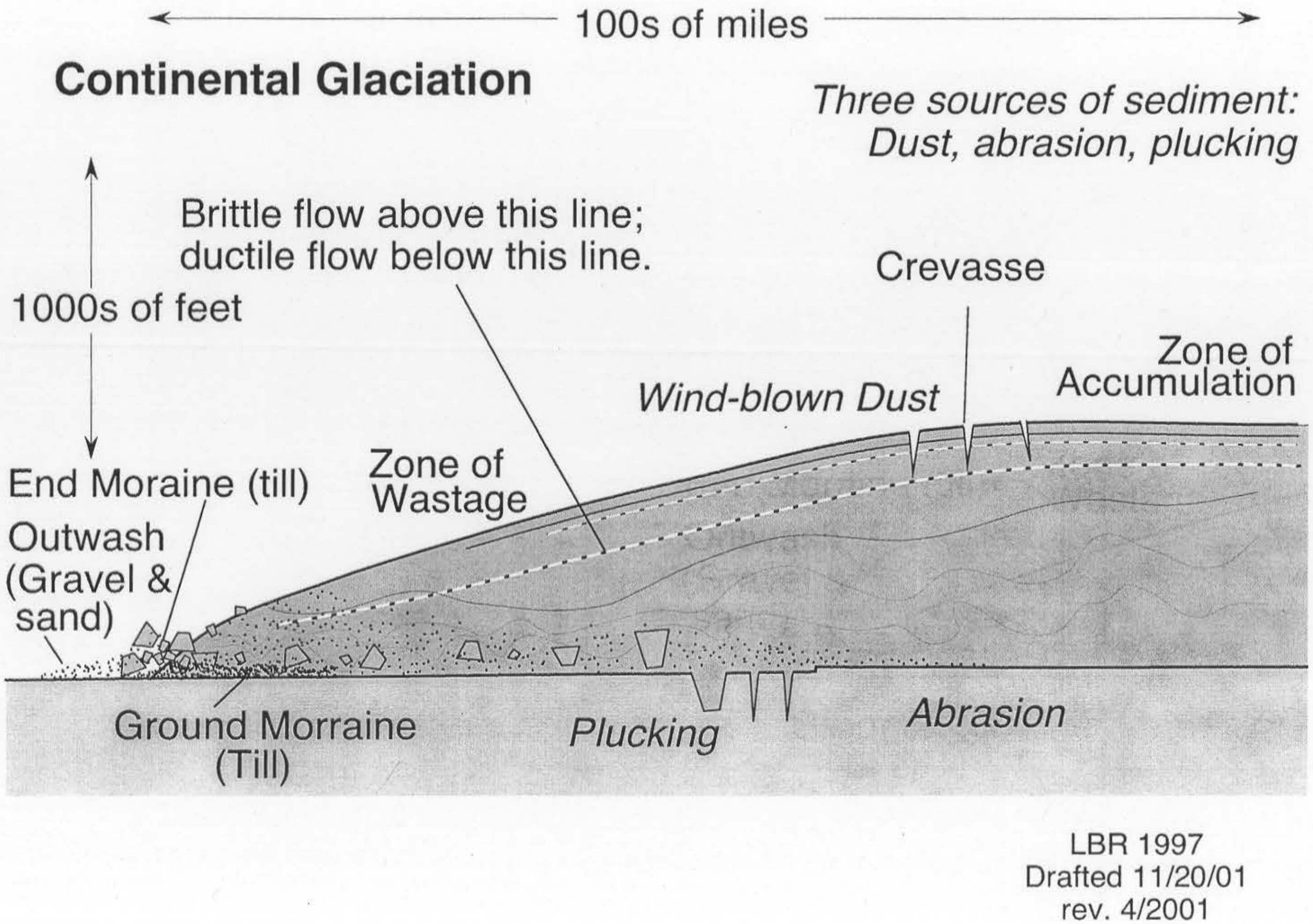

Figure 19-1: A schematic cross-section of continental glaciation.

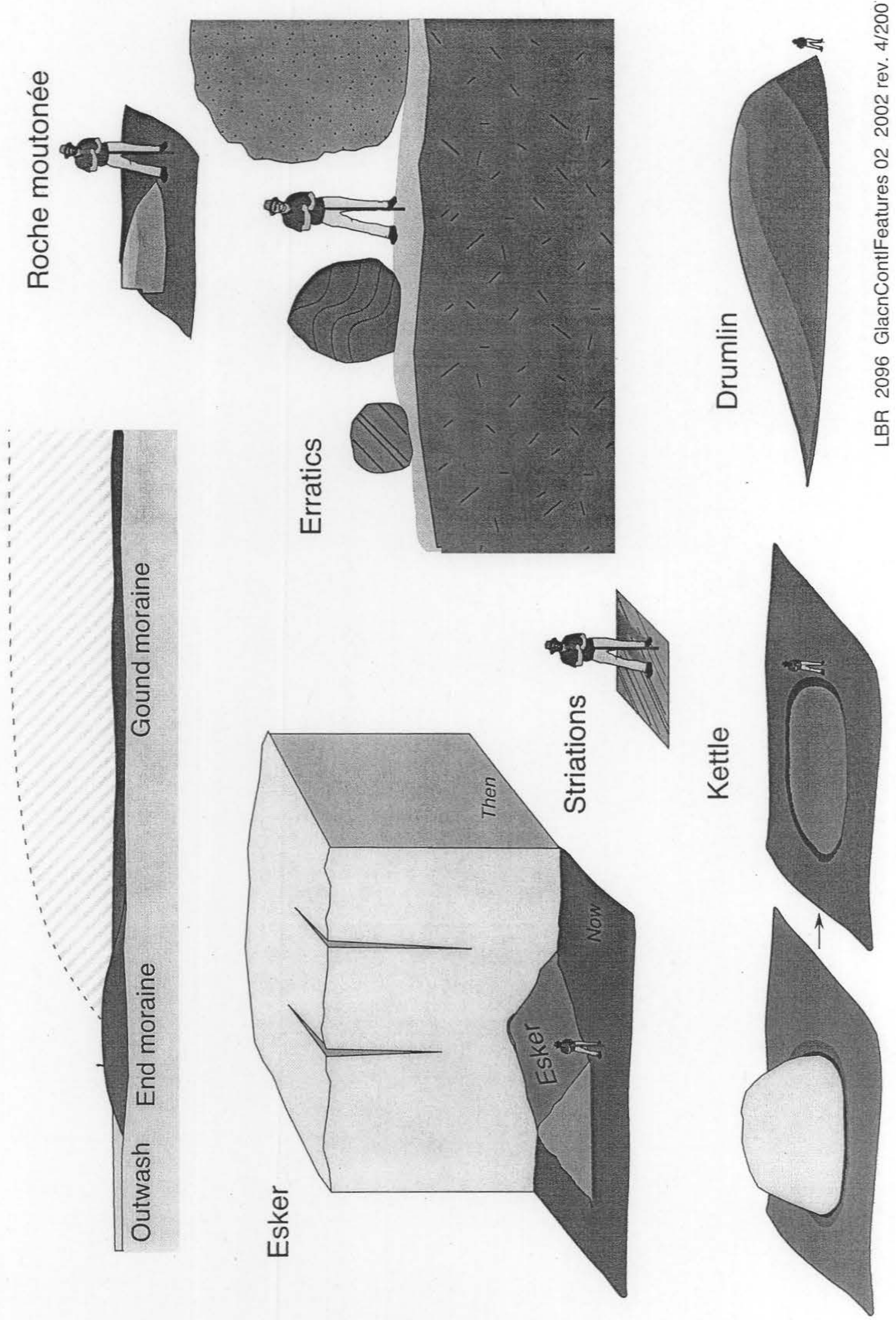

Figure 19-2. Landscape features generated by continental glaciation. Each sketch includes a human figure for scale.

CHAPTER 20. GLACIATION THROUGH TIME

The evidence of continental glaciation across northern North America discussed in the previous chapter and the presence of empty but glacially-sculpted U-shaped valleys in mountains like the Alps clearly indicate that the extent of glaciation has waxed and waned. This chapter addresses the history of glaciation, beginning at the largest scale and quickly moving to the most recent glacial periods or "ice ages".

Glaciation in Earth History

People commonly speak of "ice ages" without appreciating that the Earth has undergone major glaciation several times and at several different time scales. At the broadest scales, the Earth has undergone at least five major glaciations in its history, one (or more) in the early Proterozoic, one (or more) in the late Proterozoic, a lesser one in the early Paleozoic, a major one in the late Paleozoic, and one in the last two million years, or during the Quaternary period.

The **early Proterozoic** glaciation about 2.5 to 2.1 billion years ago is the least well understood of these five glacial periods, simply because it happened so long ago. The evidence for this glaciation is Earth's oldest glacial sediments, the Gowgonda Tillite in the Huronian Supergroup of Ontario, and other glacial sediments in Quebec, Wyoming, Michigan, Wyoming, and Canada's Northwest Territories. The huge range of time spanned by these deposits, about 400 million years, suggests that we might better think of multiple glacial events within this period rather than one, but we are unable to resolve events so long ago.

Late Proterozoic glaciation took place between roughly 850 and 600 million years ago. Glacial sediments from this period are widespread, and some occur on continents believed to have been near the equator. This has prompted the hypothesis of a late Proterozoic "Snowball Earth" that was entirely glaciated from poles to equator. However, our limited temporal resolution of these Late Proterozoic glacial events and the limited geographic resolution of the paleolatitude of Late Proterozoic continents have left most geologists unconvinced of the Snowball Earth hypothesis.

Less widespread glacial events took place in the Ordovician and Devonian. This episode is sometimes called **"Saharan" glaciation** because its earliest evidence comes from glacial erosion in the modern Saharan region. The locus of this glaciation appears to have moved across Gondwanaland, the Paleozoic southern supercontinent, and it may have tracked the position of the south pole as Gondwanaland shifted its location.

By contrast, **Late Paleozoic (largely Pennsylvanian) glaciation** was a major global event. Glacial striations radiating across the Gondwanan continents indicate widespread glaciation; the direction and extent of those striations are familiar to most geology students as evidence of continental drift. Other evidence includes Gondwanan tillites, including the Dwyka Tillite in South Africa's Karoo Basin (in fact, this glaciation is sometimes called the "Karoo" event). Cyclic sediments in the Northern Hemisphere indicate that glacial advances and retreats led to repeated rises and falls of sea level on roughly 100,000-year time scales like those of Quaternary glaciation.

Quaternary glaciation, or glaciation in the last two million years, is the subject of the remainder of this chapter. The reason for this focus is that we know so much more about the most recent event and have much better temporal resolution from its beginnings 38 million years ago to its present state today.

From the Cretaceous Greenhouse to the Quaternary Icehouse

The Cretaceous, the last period of the Mesozoic and thus the last period in which dinosaurs lived before their demise 65 million years ago, seems to have been a period of warm global climate with little or no glacial ice. The evidence for this claim comes from the widespread distribution of warm-climate fossils in Cretaceous sediments, the dearth of dropstones in Cretaceous marine sediments, and oxygen isotope evidence[102]

[102] Oxygen has two relatively abundant stable isotopes, ^{16}O and ^{18}O, of which ^{16}O is most abundant. All the oxygen of air, in water, in oxides, in carbonates, in sulfates, etc., is thus mostly ^{16}O with lesser ^{18}O. Proportions of ^{18}O and ^{16}O in substances are quantified as $\delta^{18}O$ ("delta-18-Oh") values, wherein larger numbers indicate larger proportions of ^{18}O. Hydrogen has two stable isotopes, ^{1}H and ^{2}H, where ^{2}H is deuterium (D), so hydrogen isotope compositions are expressed with an analogous δD value.

that indicates equator-to-pole temperature gradients less than those of the modern Earth. Much the same can be said of the Paleocene and Eocene, the first two epochs of the Tertiary, and thus until about 34 million years ago.

O isotope evidence from marine fossils and paleobiogeographic evidence indicate that global climate cooled abruptly about 34 million years ago, or in the early Oligocene. It appears that sea ice formed around Antarctica then, in response to Australia's splitting from Antarctica and Antarctica's resultant isolation in the cold waters of the West Wind Drift. Sufficient ice formed on Greenland and far northern Canada for icebergs to carry glacial dropstones out to sea.

The Northern Hemisphere outside Greenland and far northern Canada remained relatively ice-free until about two or three million years ago. The onset of glaciation in the Northern Hemisphere appears to have been triggered by closing of the Mid-Americas Seaway between North America and South America with the emplacement of Central America.[103] Previously, warm waters from the east-to-west Atlantic equatorial currents could flow into the Pacific, but the blockage of the Atlantic-to-Pacific passage forced some of that water north to enhance a previously weaker Gulf Stream. The greater delivery of warm water to high latitudes in the North Atlantic seemingly led to more snowfall on the continents, providing more snow to begin the glaciation of the Northern Hemisphere that defines the Quaternary Period.

Episodicity of Quaternary glaciation

The last chapter discussed the extent of glacial deposits across northern North America and northern Eurasia. Study of those deposits in the late 1800s and early 1900s suggested that there had been four major episodes of Quaternary glaciation. The principal evidence for this concept was the presence of widespread tills and outwash, atop which were paleosols. The paleosols could have developed only on ice-free land, so the paleosols presumably represented major interglacial periods separating four glacial events. In North America, those four glacial periods in the Quaternary were called, from first to last, the Nebraskan, Kansan, Illinoisan, and Wisconsinan events. The intervening three paleosols were named Afton, Yarmouth, and Sangamon. In Europe, the four glacial episodes were called Günz, Mindel, Riß, and Würm.[104] By the mid-twentieth-century, radiocarbon dating confirmed that the Wisconsinan or Würm event had occurred in the last 100,000 years and peaked about 20,000 years ago. Dating of the earlier events remained more difficult to determine.

In the late 1900s, drilling in the seafloor to recover sequences of deep-sea sediments and drilling in Antarctica and Greenland to recover ice cores provided more continuous and chronologically detailed records of global climate change. These records coincided remarkably to show roughly 100,000-year cycles of greater and lesser

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Chemical precipitation of a crystal of a mineral (e.g., calcite) in colder conditions results in greater $\delta^{18}O$ of that crystal. In contrast, physical precipitation of snowfall in colder conditions results in lesser $\delta^{18}O$ and δD values of the snow. Thus both ancient calcite fossils and ancient ice can be used as paleo-temperature records. In addition, in times of glaciation, ^{16}O is preferentially removed from the oceans in evaporation and stored in glacial ice, leaving seawater with a greater $\delta^{18}O$. Calcite fossils from seawater in glacial times thus have two reasons to have greater $\delta^{18}O$ values: water of greater $\delta^{18}O$ from which they formed, and greater $\delta^{18}O$ values because of the colder water.

[103] Evidence for the emplacement of Central America about three million years ago between North America and South America comes from the geology of Central America itself, and also from the fossil record of the continents and oceans. South America's mammals were largely marsupials until about 3 million years ago, when placental mammals arrived from the north across the new land bridge and drove most marsupials to extinction. At about the same time, corals in the Atlantic and Pacific, which had been largely the same species in both oceans, diverged to yield new species unique to each ocean, seemingly as they were separated by the land blockage.

With all that said, a recent review concludes that, although the causal linkage of closure of the Central American Seaway and Quaternary glaciation is possible, it may have only been a coincidence. See Molnar, P., 2008, Closing of the Central America Seaway and the Ice Age: A critical review: *Paleoceanography*, v. 23.

[104] The North American names of the glacial periods come from the names of states; the European names come from the names of river valleys draining the Alps northward to the Danube and the Black Sea. The Würm, for example, flows out of the Starnberger See due north of Innsbruck and into the Isar, the river on which Munich is located, and the Isar flows into the Danube.

glaciation (Fig. 20-1). In these cycles, times of greater glaciation ("glacial maxima") were times of lower sea level as water was stored on land, whereas times of lesser glaciation ("interglacials") were times of higher sea level. Within these cycles, the last 10,000 years emerge as an interglacial, with the present as the end of an interglacial with the world seemingly poised for a gradual return to increasingly glaciated conditions.[105]

Trends and events within the 100,000-year cycles

One of the striking features of these glacial cycles is their asymmetry. After each interglacial there is a gradual increase in $\delta^{18}O$ of marine microfossils over several tens of thousands of years, suggesting a slow cooling of the Earth and/or a gradual accumulation of glacial ice. After the glacial maximum (the low point of the curve) is reached, the subsequent warming and/or retreat of glacial ice seems to have happened much more quickly, over a period of about 10,000 years. This implies a rapid meltout and release of freshwater. This pattern of slow glacial advance and rapid glacial retreat on a 100,000-year basis is much like the pattern inferred from sea level change during the Pennsylvanian ("Karoo") glaciation.

Closer examination of the isotope records shows that the record from glacials (but not interglacials) is spiky (as, for example, from 70,000 years ago to 15,000 years ago). The spikes suggest rapid change from full glacial conditions to conditions halfway between glacial and interglacial conditions. These rapid (multi-decade) warming events are called Dansgaard-Oeschger events, and between them cooling was again slow and warming was rapid (the spikes are asymmetrical in the same way that the large 100,000-year cycles are asymmetrical). Cores of deep-sea sediments show that Dansgaard-Oeschger warming events were often immediately preceded by Heinrich events in which icebergs rafted unusually large supplies of dropstones to the oceans. The picture that emerges is one of slow cooling for a few thousand years and growth of ice sheets and then surging and collapse of the ice sheets to yield armadas of icebergs generating a Heinrich layer, just before or at the beginning of a period of warming. Glacials of the 100,000-year cycles were thus not sustained events for tens of thousands of years but featured climatic variations within them.

The transition from the last glacial maximum to the present Holocene interglacial featured two major pulses of melting. Between those two pulses, about 13,000 to 11,600 years ago, was one last gasp of more glacial conditions. That last cold period is called the "Younger Dryas". Since the Younger Dryas, ice cover has diminished to where "only" Antarctica and Greenland remain covered by continental ice sheets.

Ultimate and proximal causes of periodicity of Quaternary glaciation

The 100,000-year periodicity or cyclicity of Quaternary glaciation has been attributed to cyclic changes in Earth's orbit around the sun, which would cause cyclic changes in the seasonality of global climate. The longest cycle is that of variation in the eccentricity or ellipticity of Earth's orbit around the sun (the extent to which Earth's orbit deviates from a circle). In this cycle, the deviation of Earth's orbit from a circle varies from 1% to 5%. This cycle repeats itself every 98,000 years. A shorter cycle is that in which the tilt of earth's axis changes from its present 23.5° relative to the ecliptic plane to as little as 22° and as much as 25°. This cycle repeats itself every 41,000 years. A still shorter cycle is that in which the position of the Earth at any one season (for example, the vernal equinox) moves around Earth's orbit from one side of the sun to another. This cycle, called the Precession of the Equinoxes, repeats itself every 23,000 years. The mathematical summation of these cycles, which are called Milankovitch cycles[106], yields a variation in solar radiation to the earth closely matching the pattern observed in the isotope records from deep-sea cores and ice cores.

Milankovitch cycles account for the timing of glacial cycles, but the changes in insolation

[105] The last 10,000 years is the geologic epoch called the Holocene (so that one lives today in the latest Holocene), whereas the rest of the Quaternary Period (the last two million years) prior to the Holocene is called the Pleistocene.

[106] The linkage between these cycles and the timing of glaciation was first discussed by J.A. Adhemar in 1842 in terms of precession of the equinoxes and by James Croll in 1864 in terms of eccentricity and precession. However, the cycles are named after Milutin Milankovitch (1879-1958), a Serbian mathematician who linked all three cycles to changes in glaciation.

(incident solar radiation) through these cycles are not sufficient to explain the changes in global temperature and climate. In other words, Milankovitch cycles of orbital change seem to be the clock determining the periodicity of glaciation, and are thus the ultimate cause of the periodicity, but they are inadequate to be the proximal driving cause of changes in the extent of glaciation. The direct cause seems to have been revealed by analysis of ice cores from Antarctica, which show that the concentrations of CO_2 (carbon dioxide) and CH_4 (methane) in the atmosphere have varied in perfect synchrony with global climate over at least the last 400,000 years (Fig. 20-2). Milankovitch cycles have seemingly controlled changes in ocean circulation and oceanic burial of sedimentary carbon, which have controlled the amount of carbon in the atmosphere, which have controlled the extent of the greenhouse effect trapping energy in Earth's atmosphere.

CO_2 and CH_4 concentrations have thus varied with the same periodicity as that of the extent of glaciation. CO_2 concentrations have typically been about 180 ppmv (parts per million by volume) during glacial maxima and about 280 ppmv during interglacials. CH_4 concentrations have typically been about 350 ppbv (parts per billion by volume) during glacial maxima and about 700 ppbv during interglacials.

It is worth noting that concentrations of these gases have increased over the last 150 years to about 380 ppmv CO_2 and 1750 ppbv CH_4 (increases of 36% and 150% respectively relative to the natural interglacial levels) (Fig. 20-2). Both increases are clearly anthropogenic. For example, the increase in CO_2 is chronologically coincident with the burning of fossil fuels, *and* the carbon isotope composition of CO_2 in the atmosphere has changed exactly as one would expect if ^{13}C-poor carbon from fossil fuels were added to the atmosphere. If one accepts that the striking correlation of greenhouse gases and climate indicates that the concentrations of greenhouse gases has controlled global temperature, the rapid increases in the concentrations of those gases over the last 150 years would imply that one could expect rapid warming of the earth (opposite the gradual return to "increasingly glaciated conditions" mentioned above). The implications of these extremely rapid increases in concentrations of greenhouse gasses relative to their interglacial concentrations have been the subject of widespread concern in the scientific community.

Effects of deglaciation

Retreat of alpine and continental glaciers from their greatest extent at the Last Glacial Maximum (about 20,000 years ago) to the present has had several important effects.

Global sea level has risen. Global sea level was near its present level at the last interglacial 120,000 years ago, dropped as glaciers advanced, and has risen about 125 meters since the Last Glacial Maximum. This has flooded what we today consider the continental shelves, flooding the land bridge across which the people who would become Native Americans crossed from Siberia and similarly flooding the land bridge from Europe to Britain.

Seawater salinity has decreased with the return of glacial water to the ocean. Overall, global average salinity has decreased from a likely maximum of 35.7 parts per thousand at the Last Glacial Maximum to its present value of 34.7 parts per thousand.[107] The relative magnitude of that shift is apparent if one considers that almost all modern seawater has salinity between 34.0 and 35.2 parts per thousand.

In the shorter term, episodic outbursts of meltwater from the Laurentide ice sheet about 10,000 years ago appear to have diluted the surface waters of the North Atlantic significantly. North Atlantic Deep Water presently sinks from the surface of the North Atlantic and is the main source of water to the deep oceans. The pulses of meltwater to the North Atlantic from the failing Laurentide ice sheet appear to have choked off the formation of North Atlantic Deep Water, causing short-lived but profound disturbances in global deep ocean circulation and climate.

Isostatic rebound of land with the removal of masses of glacial ice has caused the uplift of areas once weighed down by glacial ice. Central Sweden has risen as much as 700 meters in

[107] The estimate of a 1 part per thousand change in oceanic salinity during late Pleistocene to Holocene deglaciation is from Keigwin et al. 1991 (*Journal of Geophysical Research*, v. 96C, p. 16,820) and is in accord with the author's calculations.

the last 10,000 years, with resultant emergence of the Baltic shoreline. The land near and under eastern Hudson Bay has risen about 120 meters, so that Hudson Bay has grown smaller as its shorelines have risen. In southern Alaska, isostatic rebound occurs in places at the rate of one to two inches per year, leaving docks and wharves useless above the water level after a few decades.

Exposure of vast expanses of glacial till has provided the foundation for soil development that has allowed extensive agriculture in North America and Eurasia. Sustained agriculture requires (i) relatively flat land so that erosion does not remove the soil, (ii) soils from which nutrients have not been leached by previous rains, and (iii) sufficient rainfall to support plant growth. Exposure of till provides the first two of these and thus helps account for the sustained "breadbasket" of production of corn and wheat in North America over the last two centuries.

Mass wasting has modified previously glaciated landscapes. This effect is most noticeable on the steep sides of U-shaped valleys, which collapse to yield rock falls and landslides. Human lives are lost and property destroyed each year in such events. On the gentler slopes generated by continental glaciation, slumps and earthflows modify landscapes of till, generally with less harsh human consequences but with noticeable effects on the landscape.

Some anthropologists would argue that the transition from the Last Glacial Maximum to the present interglacial also played a major role in the development of human civilization, as humans adapted to the changing climates that developed and to the new regions made available for settlement as ice retreated. Such claims may be overstated, but the Pleistocene-to-Holocene deglaciation has certainly had, and still has, major effects on the world we know today.

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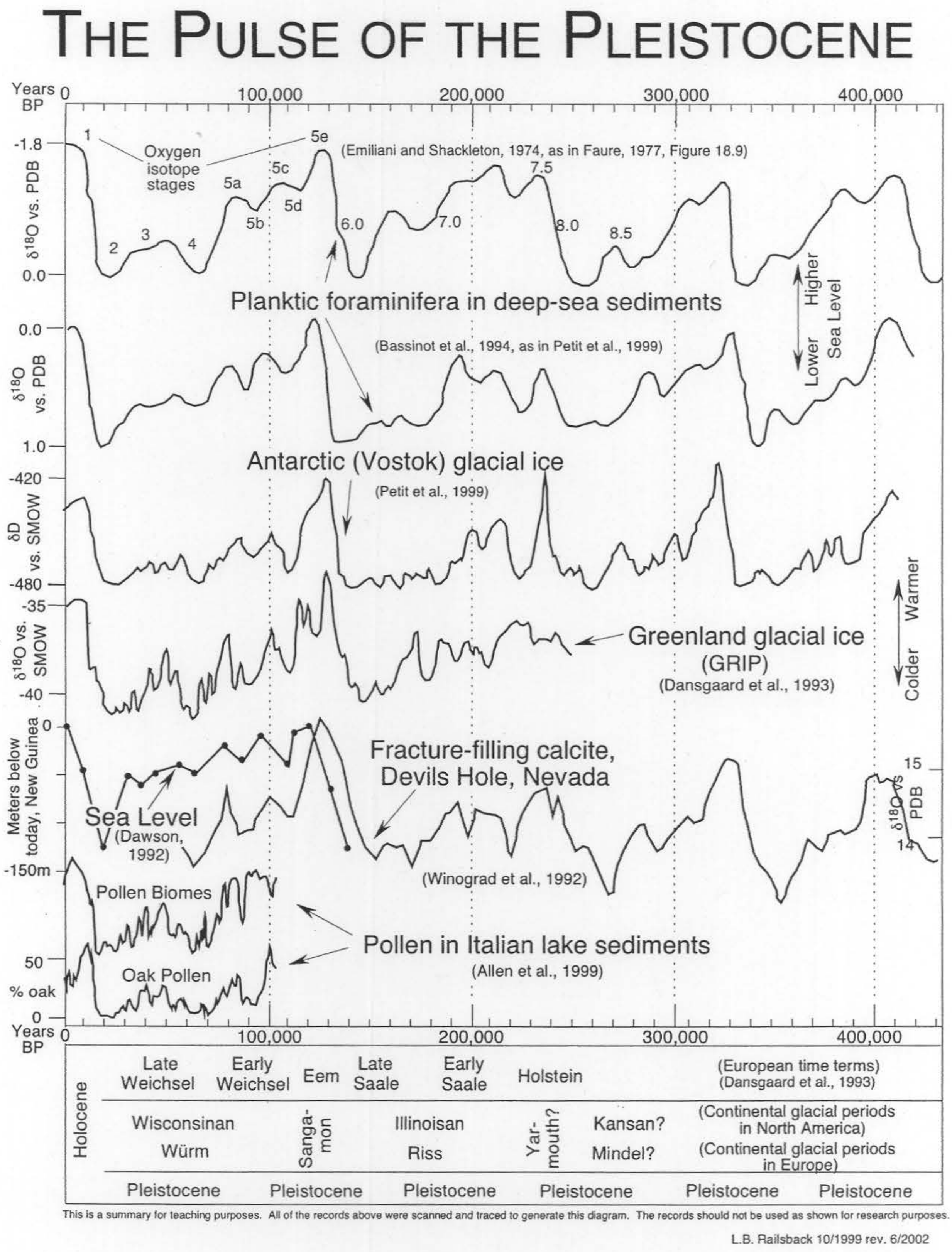


Figure 20-1: Trends in paleoclimatic indicators over the last 400,000 years.

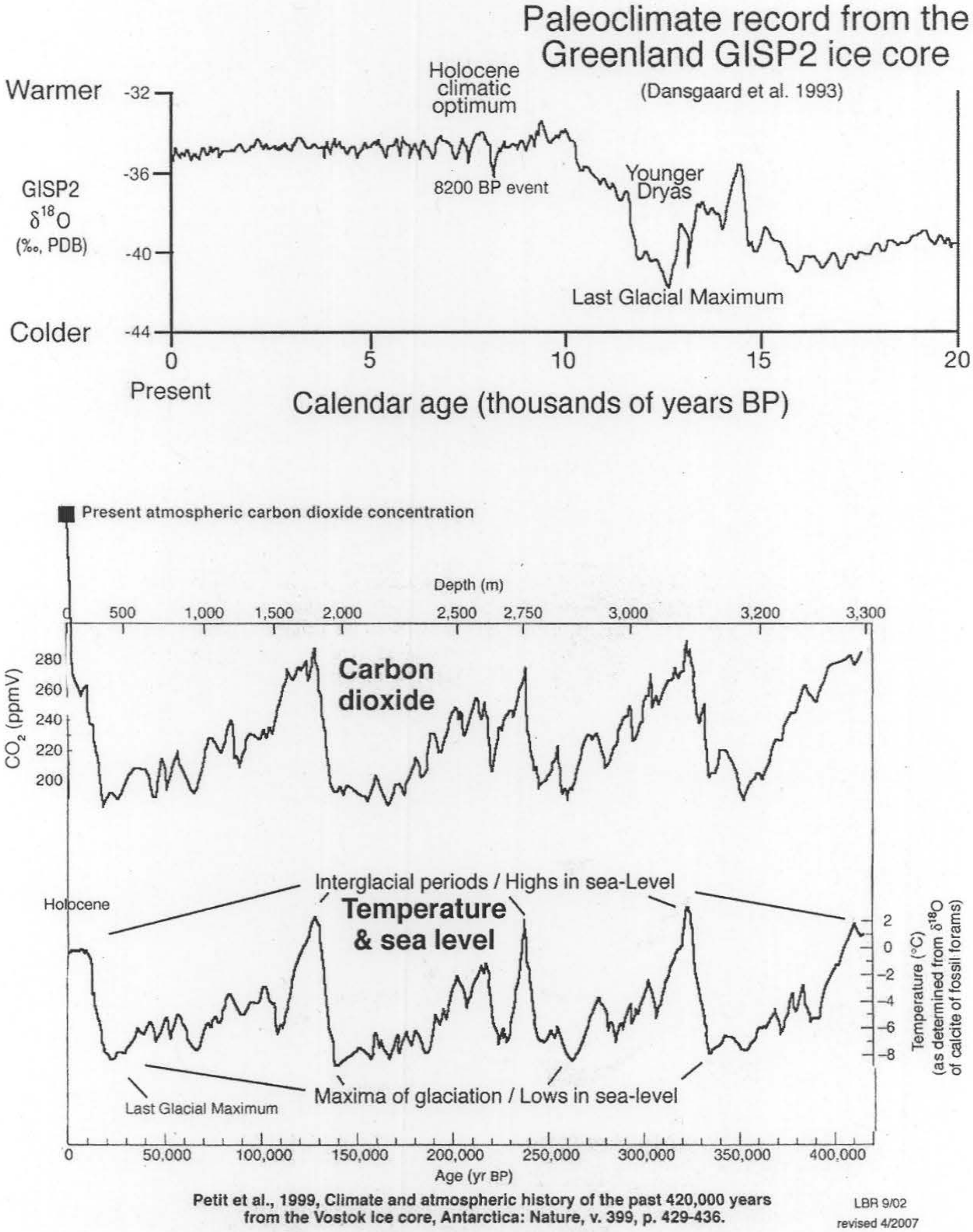


Figure 20-2: Trends in inferred climate and sea level, and PCO_2 of the atmosphere over the last 400,000 years. Top panel shows details of the last 20,000 years.

Part IV. Summary, and More

CHAPTER 21: CONCLUDING THOUGHTS: BIG MOUNTAINS AND BIG GLACIERS

The connection between mountains and glaciers

The preface of this book pointed out that the book covers the topics of alpine geology and glacial geology to meet the needs of a course taught in the summers in Innsbruck, in an alpine and glaciated setting. However, the conclusion of this book is a good place to point out that coverage of the two topics is not just a marriage of convenience, for at least three reasons.

Firstly, high mountains almost inevitably have alpine glaciers, or at least they had them until the last few decades. Mountain ranges at higher latitudes, such as the Alps and northern Rockies, clearly have had extensive coverage by alpine glaciers, with some surviving today. Very high mountains, such as the Himalayas, have extensive glacial coverage even at lower latitudes. The point can be extended further, however, in that the survival of glaciers on Mount Kilimanjaro into the twenty-first century demonstrates that high mountains can even have alpine glaciers at the equator.[108] Big mountains thus almost inevitably mean glaciers in the Quaternary.

Secondly, any understanding of the non-ice landscapes of mountain ranges requires an understanding of alpine glaciation, because erosion by glaciers modifies mountain landscapes so much. Alpine glaciation steepens slopes and broadens valleys, radically transforming how mountain ranges look. It is instructive to note that our two iconic examples of rugged mountain peaks, the Matterhorn and Mount Everest, are glacial horns that would not be nearly so impressive without the sculpting performed by alpine glaciers.

Finally, there are probably causal links between the uplift of mountains and the onset of glaciation at the global scale. On the one hand, geologists have long recognized that times of extensive mountain building at the global scales coincide with lower sea level. Lower sea level inevitably lessens earth's absorption of solar radiation, causing global cooling, and lower sea level may be correlative with slower seafloor spreading and thus less volcanism and less resultant emission of carbon dioxide from volcanoes, leading to a lesser greenhouse effect. On the other hand, and more directly, huge uplifts of crustal rocks inevitably lead to more chemical weathering, which may consume enough carbon dioxide to cause global cooling and glaciation. In either case, directly or indirectly, large orogenies like those of the Appalachians and Himalayas have been coincident with major glaciations (the Pennsylvanian/Karoo and Quaternary glaciations, respectively). This suggests that we cannot consider orogeny without glaciation and glaciation with orogeny, so that the topics of alpine geology and glacial geology are indeed inevitably linked.

The downfall of mountains

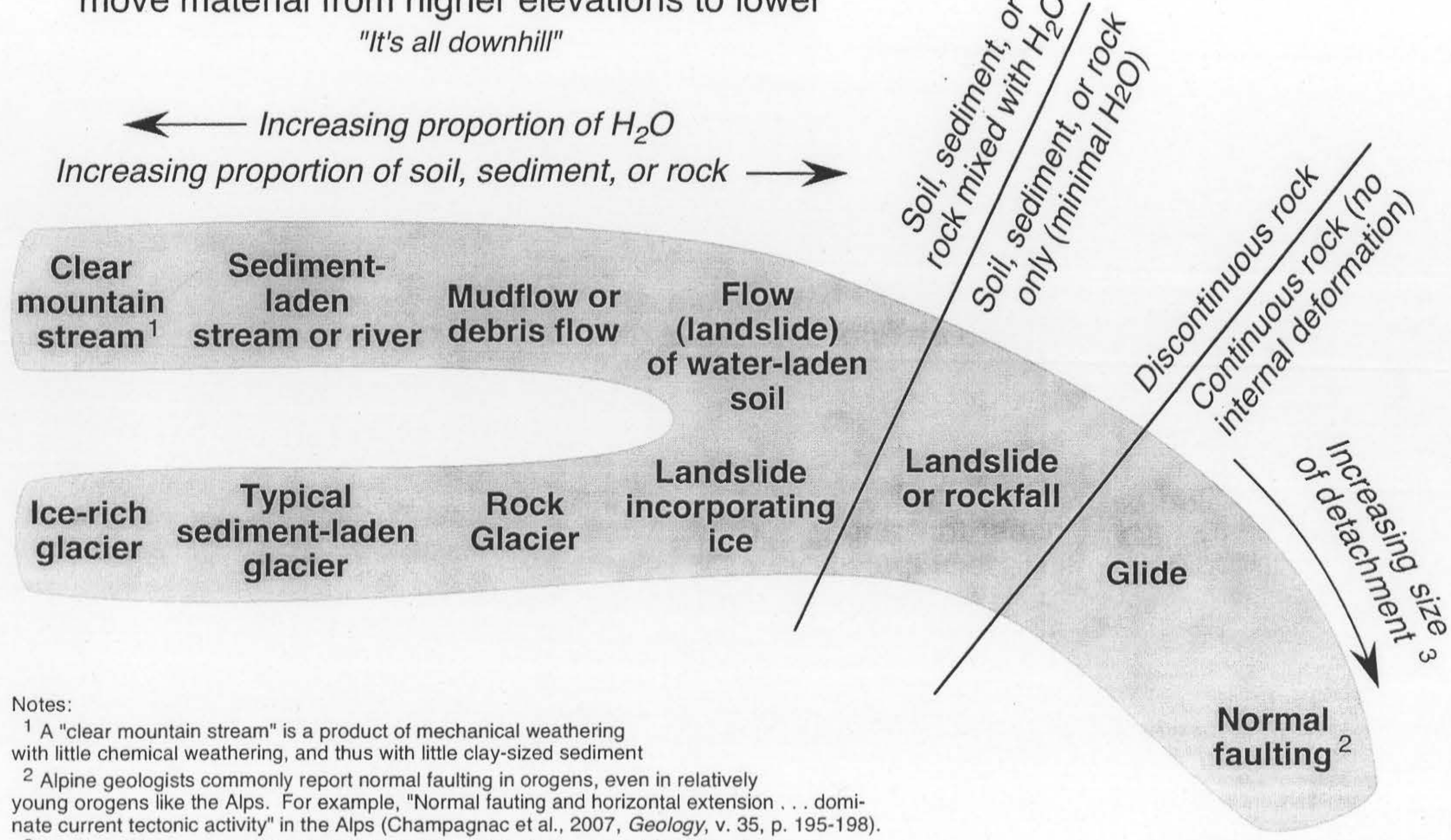
If the linkages above are causal and positive, in that one thing leads to another, there also a linkage that is destructive, at least in terms of elevation of mountains. Uplift of mountains provides a playgound in which gravity-driven geological processes romp unabated. These processes include erosion by water, erosion by ice, mass wasting, and normal faulting. These processes can be viewed as something of a continuum from the H_2O-driven movement of water in rivers and glaciers at one end, to movement of rock along normal faults at the other (Fig. 21-1). All of them carry mountain material downhill, and across geological time they destroy mountains. However, in the shorter term, these processes sculpt otherwise massive uplifted orogens into peaks, horns, ridges, arêtes, and valleys of all sorts. In that sense, we can hardly complain about this destruction of orogens, because it generates much of what we find interesting in the mountains.

[108] Another example of alpine glaciers in equatorial settings is the glaciers at elevations of 4000+ meters on the flanks of Puncak Jaya (Carstenz's Pyramid) in New Guinea. The individual glaciers include the Carstensz Glacier, the Meren Glacier, and Northwall Firn.

How nature destroys mountains

A spectrum of gravity-driven processes that move material from higher elevations to lower

"It's all downhill"



Notes:

[1] A "clear mountain stream" is a product of mechanical weathering with little chemical weathering, and thus with little clay-sized sediment

[2] Alpine geologists commonly report normal faulting in orogens, even in relatively young orogens like the Alps. For example, "Normal fauting and horizontal extension . . . dominate current tectonic activity" in the Alps (Champagnac et al., 2007, *Geology*, v. 35, p. 195-198).

[3] Another difference would be that glides typically detach along weak bedding planes, whereas a fault might detach along a surface not parallel to bedding or other pre-existing structure.

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Figure 21-1: A spectrum of gravity-driven processes that destroy mountains.

CHAPTER 22: ALPINE ECOLOGY

Anyone who reads this book will surely be interested in going into the mountains. In doing so they will see ecological, as well as geological, things of interest. With that in mind, this chapter presents a brief introduction to alpine ecology.

The alpine environment (Fig. 22-1)

The environment at high altitudes poses challenges to life beyond the conditions at lower elevations. Perhaps most notably, atmospheric pressure is lower because of the lesser load of overlying air. At 3000 meters, atmospheric pressure is only about two thirds that at sea level; at 5000 meters it is barely more than half that at sea level, and at Everest-like elevations it is only a quarter to a third that at sea level. Air at higher elevations is also relatively dry, so that organisms must protect themselves from desiccation more.

Another feature of air at high altitude is that it is clearer, with less particulate matter. As a result, a viewer at altitude sees a sky of darker blue above, because less of the sun's incoming light has been scattered. In fact, pictures from heights approaching that of Everest show a dark blue to nearly black sky. The lack of scattering also means that sunlight is more intense at high elevations, with implications both for plants and for humans.

The low air pressure at elevation and paucity of water vapor in that air lead to one of the most striking features of the alpine environment: it is cold, especially at night. As a result, conditions at high elevations are similar to those in polar regions, and many plants common in the Arctic are also common at high elevations.[109] In the northern Alps, elevations above 1800 to 2000 meters never see conditions reasonably considered "summer", and instead spring moves directly into fall. Above 3000 meters, even those seasons disappear, and the climate is effectively year-round winter. These are of course the conditions in which ice can survive year-round, and thus in which alpine glaciers can form.

Finally, one far-reaching environmental aspect of mountains is the effect they have on rainfall. True mountain ranges extend sufficiently high to force winds to rise sufficiently that water vapor condenses, and rain or snow falls. This "orographic effect" leads to greater precipitation on the upwind sides of mountains and less rain on the downwind sides, and beyond. This effect is seen in the Alps, where more rain from North Atlantic vapor falls to the north of the Alps than in the Inn Valley. It is also apparent in western North America, where the Sierras and Cascades (and farther along the Rockies) set up a rain-shadow effect all the way to the Dakotas and Kansas.[110]

The alpine substrate

This book has obviously discussed the geologic structure of mountains, but an understanding of alpine ecology requires an additional understanding of the immediate land surface in mountainous regions. The most striking geo-ecological feature of that surface is the thin soil, or no soils at all. The combination of diminished plant growth and more intense erosion on high slopes, in comparison to conditions on lower and flatter land, means that little soil forms and less stays in the alpine landscape. This has obvious implications for plants needing soil in which to grow. In the Alps, an additional factor in soil development is that limestones of the Calcareous Alps weather by chemical weathering that produces only dissolved solids (Ca^{2+} and HCO_3^-), and so the only clays to form soil are those from minor shale layers and from clays disseminated in the limestones themselves. By contrast, the metamorphic Alps generally have a silicate mineralogy that can weather chemically to generate at least some clay.

A second feature that differs among alpine landscapes is the kind of cover provided by loose rock (as opposed to bedrock). Scree slopes (slopes of small pieces of rock) are continually prone to slippage, and their small clast size results in few inter-clast holes of a size sufficient that animals can burrow. By contrast, talus slopes tend to be

[109] Academically, one sees the similarity of arctic and alpine environments in the name of the journal *Arctic and Alpine Research* and of the University of Colorado's Institute for Arctic and Alpine Research (INSTAAR).

[110] The Big Island of Hawaii sets up a similar and very intense rain-shadow effect, with much more rain on the northeast side of the island (up-wind in the Trade Winds) and dry on the southwest side. This leads to a difference in soils that readers who remember Chapter 7 will appreciate: The northeast side of the island has gibbsite-bearing soils like those of tropical to equatorial settings, and the southwest side of the island has smectitic soils like those of dry to desert-like settings.

more stable because initiating movement of large clasts requires a greater incident force, and the large clasts in talus provide larger inter-clast holes in which animals can find shelter. This is again a difference between the Calcareous Alps and metamorphic Alps, in that the limestones of former tend to shatter to smaller pieces to generate scree slopes, whereas talus slopes are more common in the metamorphic Alps.[111]

Alpine plants (Fig. 22-2)

The non-biological considerations above provide the context for alpine life, and especially alpine plants. Alpine plants face the challenge of a short growing season, minimal soils, and a cold environment. Some common adaptations to the cold include hairy leaf surfaces that trap warmth at a very small scale, low mats of leaves or "plant cushions" that trap warmth at the scale of the entire plant, and growth of plants in clusters or "tree islands" that trap warmth at the scale of plant communities. In addition, the flowers of at least some alpine plants rotate through the day to optimize their capture of the sun's rays and/or to reflect those rays onto the stamens and pistils to the maximal extent possible.

The shape and distribution of alpine trees has much to do with their environments. Alpine trees commonly have thick and/or bent trunks that form as the tree loses limbs to the wind and cold but the trunk continues to thicken, even if its apical branch is lost. Trees with such distorted shapes, compared to the form of their straighter and more symmetrical lowland cousins, are called "krummholz" (the German-language expression for "bent wood"). Strong wind from one consistent direction can remove all the limbs on all but one side of a tree. The result is a tree with limbs only in the downwind direction; such trees are called "flagged trees".

The most striking feature of the distribution of trees in alpine environments is the tree line, the upper limit of the occurrence of trees. The trees at that limit are commonly krummholz and flagged trees, and they commonly are clumped into tree islands. Trees at the tree line are also commonly limited to one or at most a few species, in comparison to different or more diverse species at lower elevation. For example, the tree line in the Alps is characterized by larch and stone-pine (zirben), but not by the other pines or deciduous trees found at lower elevation. Conifers are favored at the tree line because their sugary sap has a low freezing temperature that allows them to withstand cold summer nights better than deciduous trees.

In dry regions, mountains may also have a lower tree line, below which trees are not found. The lower tree line occurs because any orographic effect is too small at that elevation to yield sufficient rainfall for tree growth, and/or because temperatures rise so high that they dry out soils beyond the point where there is sufficient soil moisture to support trees. A similar phenomenon dictates that, in the Northern Hemisphere, trees are commonly more abundant on north-facing slopes than south-facing slopes of drier mountainous regions. Heating by direct sunlight onto the south-facing slopes evaporates soil moisture sufficiently to limit tree growth in areas where north-facing slopes with greater soil moisture support the growth of trees.

A subtle taxon-specific response of many plants, both lowland and alpine, is a tolerance for either acidic soils or alkaline soils. This can be especially significant in mountains, where geology commonly juxtaposes diverse rock types. It is significant in the Alps, where the alkaline soils of the Calcareous Alps favor flowers like gentians and *Scabiosa sp.*, whereas weathering in the metamorphic Alps produces more neutral to acidic soils that favor flowers like bluebells (glockenblumen).

Alpine animals

Alpine animals also have special adaptations, both physiological and behavioral, that allow them to survive in such a harsh environment. Physiological adaptations include lessening of circulation to limbs to limit heat loss and altered pulmonary systems, including the greater production of nitrous oxide (NO) by humans to increase their uptake of O_2.[112] Life at high elevation nonetheless takes its toll, as can be seen

[111] This may account for the observation that, at least in the author's experience, murmeltiers are more common in the metmorphic Alps than the Calcareous Alps.

[112] Beall C., et al., 2001, Pulmonary nitric oxide in mountain dwellers: *Nature,* v. 414, p. 411-412.

in the lessened reproductive capability of humans at high elevation in the United States.[113]

Behavioral adaptations to the alpine environment include seasonal migration upslope in spring and downslope in fall, and hibernation in winter. Pikas, small alpine lagomorphs, have remarkable behavioral adaptations that include harvesting, drying, and storing grass in summer. They then eat, and insulate themselves with, the grass in the winter. Murmeltiers, larger rodents common in the Alps, by contrast hibernate through the winter.

The alpine environment and human activity (Fig. 22-3)

Humans have also altered their behavior to deal with the alpine environment, and they have altered the alpine environment. Traditionally, most human activity high in mountains consisted of animal husbandry, with flocks or herds led upslope in the spring and downslope in the fall. In Tirol, the spring procession traditionally involved decorating cows with flowers before leading them to high pastures in the metamorphic Alps. Sheep grazing, on the other hand, is more characteristic of the steep slopes of the Calcareous Alps.

These grazing activities represent the topographically upper limit of alpine farming. In the Alps, slopes farther down are used for growing hay, still on slopes so steep that all work must be done by hand or at most with hand-controlled power equipment. Valley floors may be used for hay where not used for row crops.

These farming activities have impacted the alpine environment over the centuries. Grazing at high elevations lowers the tree line as animals eat the small trees and thus bring the tree line down to the elevation of large trees that can't be eaten. Clearing of forested slopes allows erosion, although then covering those slopes with grasses for hay may stabilize them significantly.

Another traditional human activity in mountainous regions is removal of timber for firewood and construction. Where performed by small populations and by people with no ability to transport wood to larger distant markets, such timbering and non-clear-cutting may have not had a great impact on mountainous regions as a whole. Today, with clear-cutting at a large scale driven by distant markets, deforestation can rapidly eliminate tree cover across broad regions and expose fragile thin soils to extreme erosion, essentially destroying the ecosystem for the foreseeable future.

Another activity that traditionally had minor impact was mining. Most mining in mountains before the modern era consisted of underground mines with small openings that caused little destruction of the surrounding land surface. Such mining was also sufficiently inefficient, or sought only such high-grade ores, that it did not produce large piles of tailings.[114] Today, technology allows more effective underground mining and also allows the excavation of large open-pit mines that by definition remove the landscape and its ecosystem. Modern smelting and refining techniques also allow extraction of lower-grade ores, leading to large tailings piles and slag heaps. The tailings piles are commonly the source of acid-mine drainage.

An activity that has no analog in traditional culture but dominates many alpine regions today is skiing and, more broadly, tourism. Skiing and tourism bring affluent people and their money to alpine regions. Impacts on the natural environment include broader and paved roads that replace narrow unpaved roads and footpaths, towers and service roads for ski lifts, cleared ski runs from which all trees are removed, and high-rise hotels in what used to be small and isolate villages. Impacts on the cultural environment include exposure to new languages and loss of traditional ones, availability of new jobs and technologies that allow old ones to be forgotten, and sometimes a resentment of the wealth and presumption of the visitors from the flatlands.

Some villages, valleys, or entire regions can choose to avoid modernization and to limit the influence of visitors by limiting trade and tourism.[115] No alpine area, however, can avoid one

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[113] Grahn, D., and Kratchman, J., 1963, Variation in neonatal death rate and birth rate in the United States and possible relations to environmental radiation, geology, and altitude: *American Journal of Human Genetics*, v. 15, p. 329-352.

[114] Tailings are the pieces of rock that must be removed to bring the ore out of the bedrock. Slag, on the other hand, is the now-solid remains of the melting and extraction process used to purify the ore.

[115] One national-scale example is Bhutan's policy of setting a limit on the number of tourists who may visit that country each year. At a more local scale, residents of the Tirolean Kaisertal east of Kufstein insisted in 2007 that the new highway tunnel into their valley be

modern environmental impact: global warming. Global warming might seem attractive in making the alpine environment more hospitable, but it is causing changes that will greatly impact alpine regions. The most obvious, especially in the Alps, is the melting of glaciers and the scarcity of snow for skiing in the wintertime. For example, Innsbruckers tell of skiing down from the Nordkette to their homes in Innsbruck in snows that routinely covered the Inntal. Today, snow-making machines must commonly be used on the slopes above Innsbruck to put snow on the ski runs, and the first snows of the 2006-2007 winter were months later than in the 1900s. Alpine hüttes for hikers and skiers that were built at the bases of glaciers are now places from which one must squint to see the lower reaches of the glaciers much higher up the mountains.

In short, humans have impacted alpine regions, sometimes knowingly and sometimes not. Today, visitors can do some obvious things to lessen their impact: staying on posted trails rather than tramping down the vegetation and setting off erosion of the soil; not picking wildflowers and not disturbing wild animals; not treating the natives and their homes like oddities or non-humans. Those of us hiking up to glaciers should remember what a remarkable privilege we have to live at this brief moment in the history of the mountains, when roads, buses, and ski-lifts allow us to access the highlands easily, but when global warming has not yet melted the glaciers away. Neither our great-grandparents could have had, nor will our great-grandchildren be able to have, the experience that we have in our visits to the alpine and glacial environment. Enjoy!

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open only to residents, emergency vehicles, and service vehicles. Previously, only an aerial ropeway provided access to the valley. (Landler, M., Modernity Drills Through Rock Toward an Alpine Hamlet: *New York Times*, April 24, 2007).

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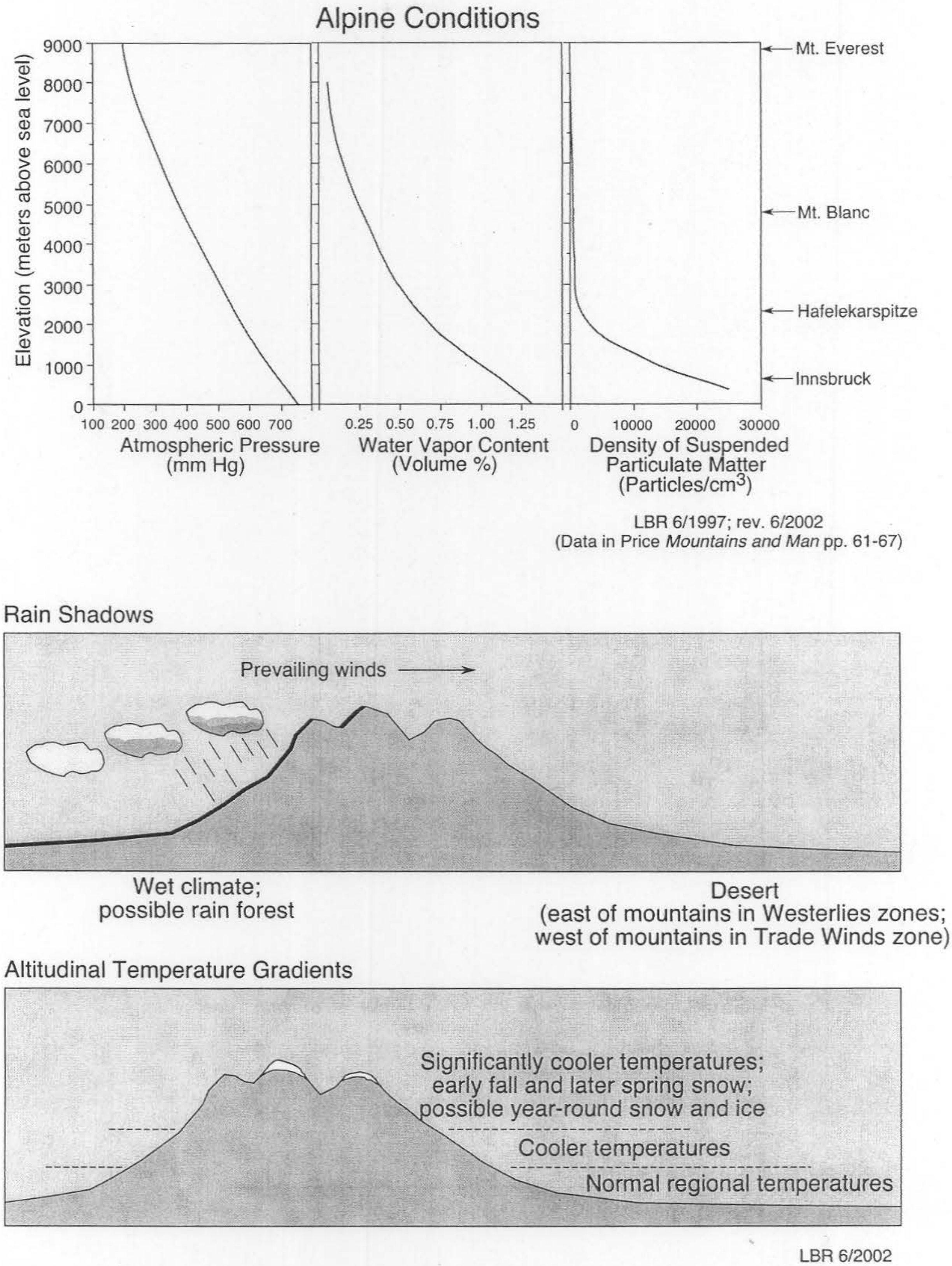
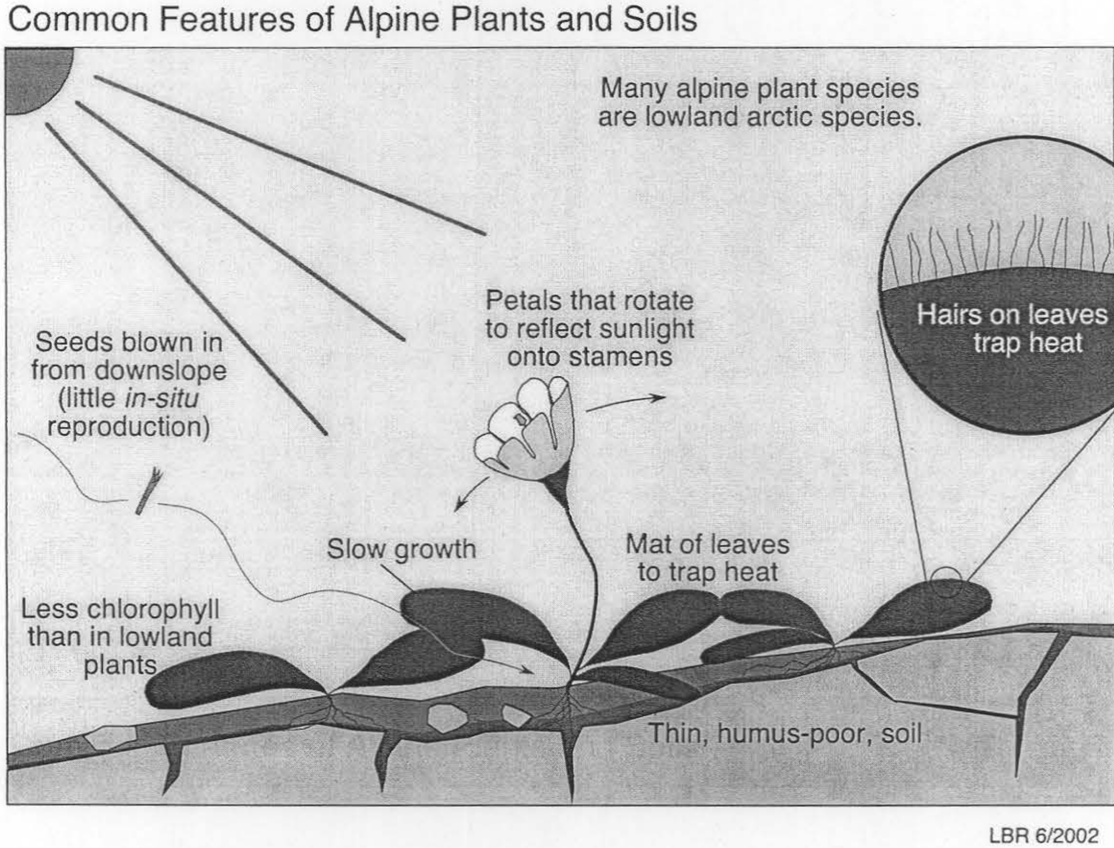


Figure 22-1: Trends in alpine conditions, and impacts of mountains on the climatic environment



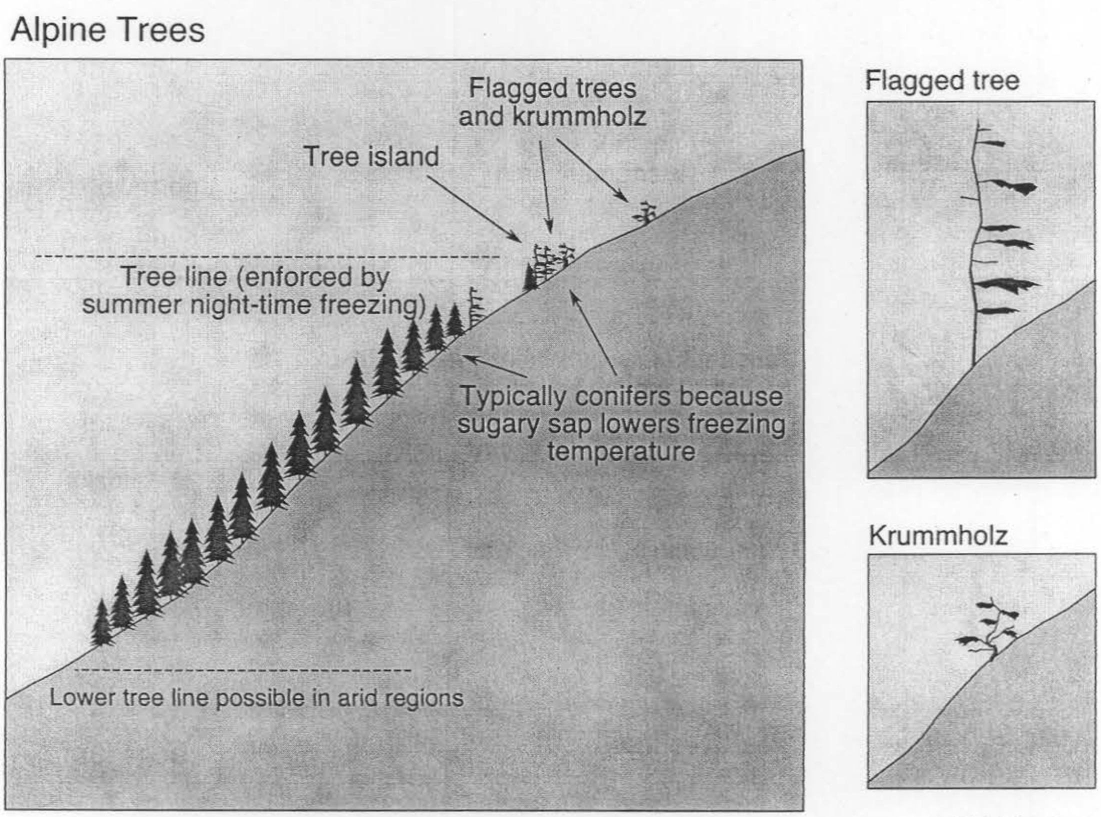
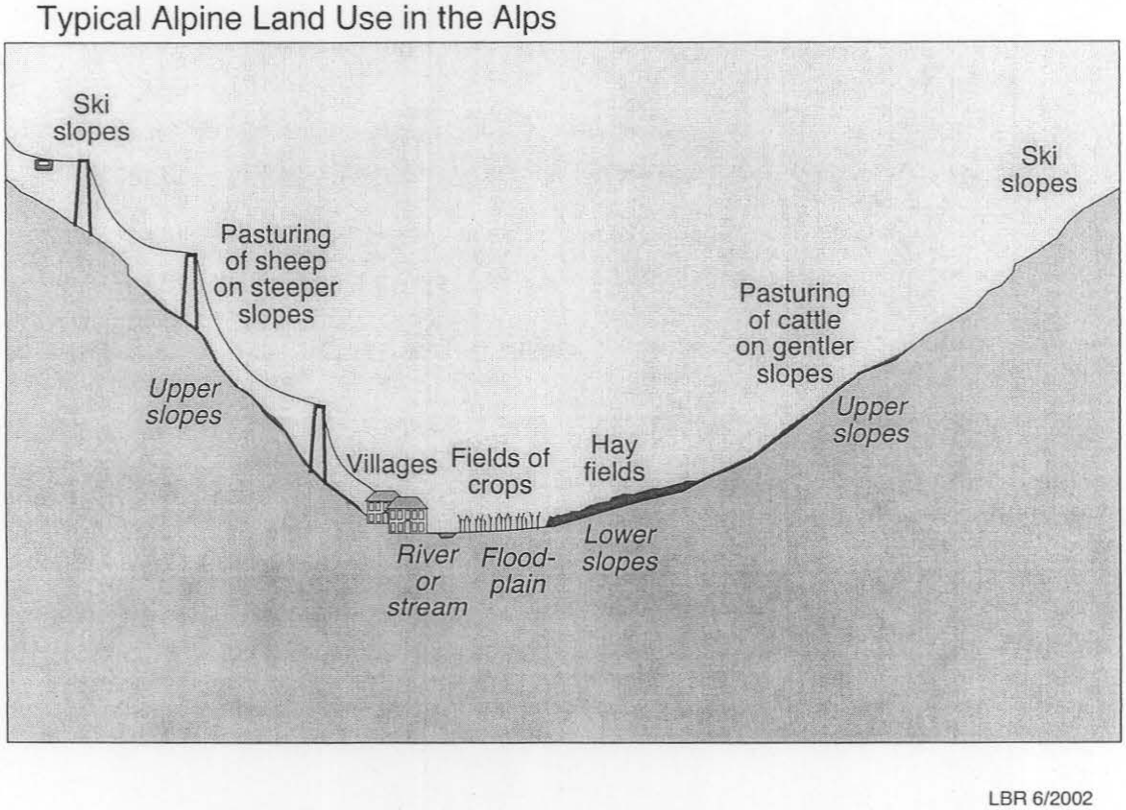


Figure 22-2: Special features of alpine low-lying plants and of alpine trees.



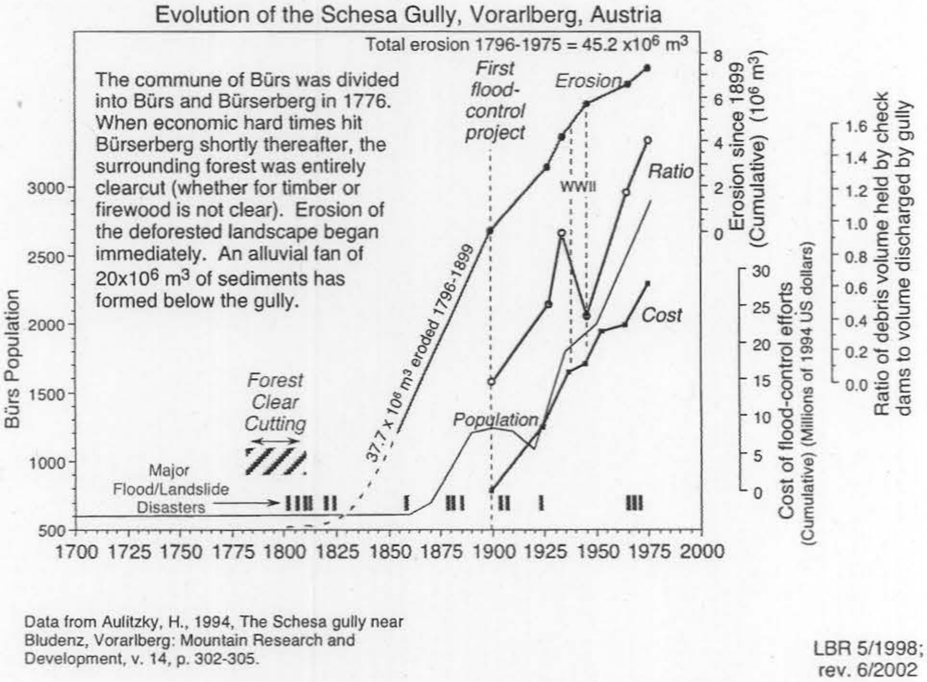


Figure 22-3: Alpine land use by humans, and an example of its potential results.

Appendix A: A List of Noteworthy Mountain Peaks (with a Tirolean-American emphasis)

Peak	Mountain range	Political location	Elevation above sea level (meters)	(feet)	Comments
Mt. Everest / Chomolunga / Sagarmatha	Himalayas	Nepal-Tibet	8848	29,028	Earth's highest mountain peak relative to sea level [1]
K2 / Godwin-Austen	Karakoram - Himalayas	Pakistan-China	8611	28,251	Earth's second-highest peak; Earth's steepest mountain
Kangchenjunga	Himalayas	Nepal-India	8586	28,169	Earth's third-highest peak
Lhotse	Himalayas	Nepal-India	8516	27,939	Earth's fourth-highest peak; neighbor of Everest
Annapurna	Himalayas	Nepal	8091	26,545	Earth's tenth highest peak; Earth's most deadly peak to climb [2]
Shishapangma	Himalayas	Tibet	8013	26,289	Earth's 14th highest peak; lowest of the eight-thousanders [3]
Aconcagua	Andes	Argentina: Mendoza	6962	22,841	Highest peak outside southern Asia; ranks well past 50th if not 100th on Earth
Ojos del Salado	Andes	Chile-Argentina	6893	22,615	Second-highest peak in Andes; highest volcano (a stratovolcano) [4]
Huascarán	Andes	Peru	6768	22,205	Alternate candidate for peak farthest from Earth's center [5]
Chimborazo	Andes	Ecuador	6267	20,560	Peak farthest from Earth's center; a stratovolcano
Mt. McKinley / Denali	Alaska	USA: Alaska	6194	20,320	Highest peak in North America
Mt. Logan	Coast	Canada: Yukon	5959	19,550	Highest peak in Canada; second in North America to Denali [6]
Kilimanjaro		Tanzania	5895	19,340	Highest peak in Africa; Earth's tallest free-standing mountain rise
Mt. Elbrus	Caucasus	Russia	5642	18,510	Stratovolcano; Europe's highest peak
Pico de Orizaba / Citaltépetl	Eje Vocánico Transversal	Mexico: Veracuz - Puebla	5636	18,490	Highest peak in Mexico; third highest in North America
Vinson Massif	Ellsworth	Antarctica	4892	16,050	Highest peak in Antarctica
Puncak Jaya / Carstensz Pyramid	Sundiman	Indonesia: Papua [7] (on New Guinea)	4884	16,023	Highest peak in Oceania; site of equatorial glaciers
Mont Blanc	Alps	France-Italy	4808	15,774	Highest peak in Alps
Das Matterhorn	Alps	Switzerland-Italy	4478	14,693	Spectacular glacial horn
Mt. Whitney	Sierra Nevadas	USA: s. California	4421	14,505	Highest peak in contiguous U.S. [8]
Mt. Elbert	Sawatch (Rockies)	USA: Colorado (near Leadville)	4401	14,440	Highest peak in Rockies; highest peak in U.S. east of Cordillera [8]
Mt. Rainier	Cascades	USA: Washington	4392	14,410	Highest peak of Cascade volcanoes [8]
Mauna Kea	Hawaii	USA: Hawaii	4205	13,796	Peak of world's tallest mountain

Die Jungfrau	Alps	Switzerland	4158	13,642	Well-known Swiss mountain [9]
Cloud Peak	Bighorns	USA: Wyoming	4013	13,167	Highest peak in Bighorns
Der Ortler	Alps	South Tirol - west of Bolzano	3905	12,811	Highest peak in historical Tirol; now in South Tirol
Der Grossglockner	Alps	Austria: Carinthia & Tirol	3798	12,461	Highest peak in Austria & eastern Alps
Fuji-san	none	Japan	3776	12,388	Stratovolcano; Japan's highest peak
Die Wildspitze	Ötztaler Alps	Austria: Tirol	3774	12,382	Highest peak entirely in Austrian Tirol
Aoraki / Mt. Cook	Southern Alps	New Zealnd	3754	12,316	Highest peak in Southern Alps[10]
Pico de Aneto	Pyrenees	Spain: Aragón	3404	11,168	Highest peak in Pyrenees
Die Marmolata	Dolomites	South Tirol	3343	10,968	Highest peak in Dolomites
-	north dome	Greenland	~3290	~10,794	Highest point on Greenland ice sheet (bedrock surface below is below sea level)
Der Habicht	Stubaitaler Alps	Austria: Tirol	3277	10,751	Glaciated peak visible from Innsbruck; "Der Habicht" means "the Hawk".
Kita-dake	Japanese Alps	Japan: Honshu	3193	10,473	Highest peak in Japanese Alps
Die Parseierspitze	Lechtaler Alps	Austria: Tirol	3036	9958	Highest peak in Calcareous Alps
Die Zugspitze	Wetterstein Alps	Germany-Austria (Bavaria-Tirol)	2962	9718	Highest peak in Germany (but only a lesser peak in Tirol)
Monte Corno	Apennines	Italy: Abruzzo	2912	9560	Highest peak in Apennines
Die Serles	Alps	Austria:Tirol	2718	8917	Jagged peak south of Innsbruck; "Der Altar von Tirol" (the altar of Tirol)
Mt. Gerlachovka	Carpathians	Slovakia	2655	8711	Highest peak in Carpathians
Die Hafelekarsptize	Karwendel Alps	Austria: Tirol	2334	7586	Peak on Nordkette above Innsbruck
Der Patscherkofel	Alps	Austria: Tirol	2246	7369	Rounded peak south of Innsbruck
Mt. Kosciuszko	Great Dividing	s.e. Australia	2228	7310	Australia's highest peak
Harney Peak	Black Hills	USA: South Dakota	2207	7242	Highest peak in Black Hills
Mt. Mitchell	Appalachians	USA: North Carolina	2037	6684	Highest peak in Appalachians and eastern U.S.
Clingman's Dome	Appalachians	USA: Tennessee	2025	6643	Second highest peak in Appalachians
Mt. Washington / Agiocochook	Appalachians	USA: New Hampshire	1917	6288	Highest peak in northeastern U.S.[11]
Mt. Marcy	Adirondacks	USA: New York	1629	5344	Highest peak in Adirondacks
Mt. Katahdin	Appalachians	USA: Maine	1606	5267	Northern terminus of Appalachian Trail
Brasstown Bald / Enotah	Appalachians	USA: Georgia	1458	4784	Highest point in Louisiana, Mississippi, Alabama, Georgia, & South Carolina
Ben Nevis	Scottish Highlands	U.K.: Scotland	1343	4406	Highest point in British Isles
Monkey Hill	Audubon Park	New Orleans, LA	~5	~15	Highest point in New Orleans. [12]

[1] "Mt. Everest" is the western name of the highest peak relative to sea level, but Chomolunga ("Mother of the Universe") is the Tibetan name, and Sagarmatha ("Head of the Sky") is the Nepali name. Estimates of the exact elevation vary, both for technological reasons and because some estimates include the topmost ice and snow (which fluctuates) and some do not. As of September 2003, 1924 climbers had summitted Everest, the greatest number of summittings for any eight-thousander (i.e., for any peak more than 8000 meters above sea level).

[2] As of September 2003, Annapurna had been summitted 130 times at the cost of 53 deaths, the greatest ratio of any peak. 130 is the fewest summittings of any of the eight-thousanders.

[3] Reinhold Messner, a native of Brixen in South Tirol, was the first person to summit all fourteen of Earth's eight-thousanders, as well as the first person to summit Everest alone (almost impossible now with the traffic of climbers) and the first person to summit Everest without supplementary O_2.

[4] The highest volcanic peak is necessarily a stratovolcano, rather than a shield volcano, because stratovolcanoes are steeper and because huge shield volcanoes are all oceanic.

[5] Huascaran is the granitic mountain from which a huge earthquake-triggered landslide killed ~20,000 people in the town of Yungay in 1970.

[6] Mt. Logan was named after Sir William Edmond Logan (1978-1875), a famous Canadian geologist. He was the first Director of the Geological Survey of Canada, and the Geological Society of Canada's highest award, the Logan Medal, is named after him.

[7] New Guinea is an island, the eastern half of which is the nation of Papua New Guinea. Papua is one of the provinces of Indonesia on the west half of the island.

[8] It's interesting to note the coincidence that the highest peak in the Sierra Nevadas (Mt. Whitney), the highest peak in the Cascades (Mt. Rainier), and the highest peak in the Rockies (Mt. Elbert) collectively differ in elevation by only 95 meters.

[9] A cog railway climbs through a tunnel in the Jungfrau up to the Jungfraujoch at 3454 m. There, tunnels house restaurants and a hotel, and another walking tunnel allows views of the Aletsch Glacier.

[10] "Aoraki" (or "Aorangi") is the Maori word for "Cloud-piercer".

[11] The Native American name "Agiocochook" means "Home of the Great Spirit".

[12] Monkey Hill in Audubon Park was built of sand, presumably from the Mississippi River, in 1933 by WPA workers. Its diminutive size and climatic location have given rise to T-shirts emblazoned with the words "Ski Monkey Hill". Laborde Lookout in the Couturie Forest and Arboretum on Harrison Avenue in City Park is a higher topographic feature in New Orleans, but it consists of debris from the construction of Interstate 610 (as reported in the Blake Ponchartrain column in the *Gambit Weekly* of January 13, 2004) and thus is doubly unnatural.

The list above includes the Seven Summits, the highest peaks of each of the seven continents (Everest, Aconcagua, Denali, Kilimanjaro, Elbrus, Vinson, and Puncak Jaya). Much fanfare is made of summitting all seven, but in fact there is general agreement that summitting the Seven Second Summits (K2, Ojos del Salado, Logan, Dykh-Tau, Kenya, Puncak Trikora, and Tyree) is more difficult.

Appendix B: A Tirolean-American Earth Scientist's Glossary

die Ache: mountain stream or river (cf. Fluß)

der Ader: (mineral) vein

die Au: wetland meadow or pint bar (as in Mühlau, or as in "Rössl (a horse) in der Au"

die Alm: high clearing or pasture. Many are named for valley towns but are far above them, as in Arzler Alm and Höttinger Alm

der Bach: a stream, as in the Relsbach, a stream south of Siegen in Germany.

der Berg: a mountain

die Bergkette: a mountain chain

der Bergsturz or Bergrutsch: landslide

das Bergwerk: a mine

der Boden: ground or soil (or floor)

dei Bohrung: a well

der Bruch oder der Schichtungbruch – fault (see also Störung)

die Erde: soil, earth, ground

das Erz: ore

das Erdöl: petroleum

die Faltung: a fold; folding (not faulting!)

der Fels: a rock (cf. Stein)

der Fluß (Fluss): a river

der Gang: dike or vein of minerals

die Gebirge: mountains, as in a mountain range.

die Gefüge: structure or texture

der Gletscher: glacier

der Gletschersalz: Epsom salt ($MgSO_4$)

der Geröllstein: pebble, boulder

der Glimmer: mica

der Granat: garnet

der Granit: granite

der Gneis: gneiss

die Grube: mine, quarry, pit, or excavation (zum beispiel, eine Schottergrube)

der Hang: slope, dip

die Höhe: a high place (or altitude).

das Joch: a saddle or high pass.

der Kalk: limestone

der Kalktuff: tufa or travertine

das Kar: a cirque, commonly containing scree or loose rock; commonly used in the Karwendel (e.g., "Haflekar").

das Kees: glacier

die Kluft: fracture, (broken) cleavage

der Kofel: a summit or knoll, as in the "Patscherkofel" just south of Innsbruck.

Kogel: a summit

der Kopf: a head, and a rounded summit

das Lager: layer, stratum, deposit

der Lehm: loamy (sand-and-clay) soil

der Marmor: marble

das Moor: Bog

die Quelle: source (of a river) or spring

die Schichtung: layering, stratification

der Schiefer: Slate, shale, schist

die Schieferung: rock cleavage

die Schottergrube: gravel pit

der Schurf: prospect, test pit (for mining)

der Schutt: debris, talus

der Schwemmkegel: an alluvial fan

der See: a lake.

die Spitze: a peak or mountain top, as in "Zugspitze" and "Hafelekarspitze".

der Stein: a stone

der Steinbruch: a quarry

die Störung: a fault

das Tal: a valley, as in "das Inntal"

der Talboden: a valley floor

der Ton: clay

der Torf: peat

die Trennungfläche: a structural surface, including bedding, cleavage, and faulting

der Tuff oder Tuffstein: Tufa *or* tuff

Geologically significant chemical elements

Atomic number	Symbol	German name	English name
1	H	der Wasserstoff	Hydrogen
6	C	der Kohlenstoff	Carbon
7	N	der Stickstoff	Nitrogen
8	O	der Sauerstoff	Oxygen
9	F	das Fluor	Fluorine
11	Na	das Natrium	Sodium
12	Mg	das Magnesium	Magnesium
13	Al	das Aluminium	Aluminum
14	Si	das Silizium[1]	Silicon
15	P	der Phosphor	Phosphorus
16	S	der Schwefel	Sulfur
17	Cl	das Chlor	Chlorine
19	K	das Kalium	Potassium
20	Ca	das Kalzium	Calcium
22	Ti	das Titan	Titanium
26	Fe	das Eisen	Iron
29	Cu	das Kupfer	Copper
47	Ag	das Silber	Silver
79	Au	das Gold	Gold
82	Pb	das Blei	Lead
92	U	das Uran	Uranium

[1] siliceous = kieselartig; silica = die Kieselerde

Appendix C: Possible Exam Questions

The following is a list of questions that might be asked on exams about the material covered in this book and in associated lectures. This should not be an intimidating list because, if the exam is limited largely to these questions, students know beforehand exactly what material they are responsible for and what material, on the other hand, they can give less attention.

1. What is the age of the Earth? Be sure your answer includes units.

2. List and describe at least six physical properties of minerals.

3. What trends exist in the silicate minerals (and rocks and magmas thereof) across the spectrum from mafic to felsic?

4. Explain the classification system that leads to six fundamental rock names for igneous rocks. Your answer should make evident how one would identify these rocks in the field.

5. Match these generalized sedimentary rock types: chalk. chert, coal, conglomerate, limestone, sandstone, halite, shale
with these generalized depositional environments: continental shelves, deep sea, desert dunes, lakes, marginal seas in horse latitudes, point bars of rivers, river channels, river floodplains, shorelines, swamps, warm shallow seas, wave-swept coasts.

6. Match these metamorphic rocks: quartzite, marble, slate, anthracite
with these possible precursor sedimentary rocks: limestone, bituminous coal, sandstone, shale

7. What is the difference between a conglomerate and a breccia? What is the difference in their origins?

8. Categorize the processes involved in weathering, dividing weathering into two categories and dividing each of those into at least three processes.

9. Explain the differences between flows, glides, rockfalls, rockslides, and slumps.

10. List or sketch the various simple kinds of faults and folds and for each indicate the kind of stress that generates it.

11. What is some of the evidence for plate tectonic theory?

12. What kinds of faults and/or folds are associated with the different kinds of plate tectonic boundaries?

13. We discussed eight general ways that mountains (in the general sense) or mountain ranges can form. List and briefly describe each.

14. What is the origin of features like Atlanta's Stone Mountain, New Hampshire's Mount Monadnock, Rio de Janeiro's Sugarloaf Mountain, and Malawi's Mount Mulange? Are they mountains in the sense that geologists use that term?

15. Sketch at least three stages in the origin of a volcanic chain like Hawaii sufficient to show why these mountains vary in size as they do, and how they relate to their source of magma.

16. Contrast the kinds of mountains formed at convergent plate boundaries between oceanic and oceanic crust, between oceanic and continental crust, and between continental and continental crust. List a modern example of each.

17. What is the world's tallest mountain, from its base to its peak? How did it form?

18. Where are the worlds highest, second highest, third highest, fourth highest, fifth highest, sixth highest, seventh highest, eighth highest, ninth highest, and tenth highest mountain peaks (relative to sea level)? For each, a location down to half of a continent (e.g., northern North America or southern Africa) will suffice. What is the name of the world's highest mountain peak (relative to sea-level), and what is its elevation relative to sea level, in meters or feet?

19. On a map on which at least Munich, Innsbruck, and Venice are shown as reference points, sketch a generalized geologic map dividing this region into at least six different kinds of rocks, and indicate the kinds of rocks found in each.

20. What plate tectonic and structural processes account for the distribution of rocks and thrust faults in the region covered by the map in the previous question? Feel free to include a sketch in your answer.

21. What is the orientation of and relative movement along faults (other than thrust faults) in the eastern Alps (the Alps east of the Sill Valley)? Feel free to use a sketch map in your answer. What process accounts for the development of these faults, and how does it relate to the geology east of the Alps?

22. Sketch a map of the Himalayas dividing this mountain range and its immediately adjacent regions into at least eight parallel bands. Label each band and indicate the kind of rock of which it consists. Where major thrusts separate these bands, label them with names too.

23. Sketch a map of the southern Appalachians dividing these mountains and the adjacent regions into at least five belts. Indicate the kinds of rock and/or styles of deformation typical of each belt. Your map should include Augusta, Georgia, and Knoxville, Tennessee, as reference points, and should likewise show the location of Great Smoky Mountains National Park.

24. Where is the world's highest mountain peak that is not on the continent that is the home of the world's highest mountain peak? What is the name of that peak? What is its elevation relative to sea level? (You can round to the nearest thousand meters.)

25. What, in terms of distance from the coast, is the general relationship between the location of batholiths in the Andes and the location of Cenozoic volcanoes in the Andes? What is the commonly accepted explanation of this relationship?

26. How are volcanoes distributed along the length of the Andes, and to what is this distribution commonly attributed?

27. The mountains of the Basin and Range of the western United States can be said to be mountains generated from previous mountains. Explain this statement in terms of the tectonic history of western North America.

28. Name some of the Laramide mountain ranges of western North America. What is the origin of these mountain ranges? What is the analogous feature in South America, and how is it analogous in origin?

29. How are the Cascade mountains of the northwestern United States related geologically to the Sierra Nevadas of California? With that said, how are they different in origin?

30. How are the Andes and the mountains of the western United States similar, and how are they different? Why are they different?

31. How are the Alps and Himalayas similar, and how are they different?

32. What do straight valleys in mountain ranges usually indicate about the underlying geology? Why does that relationship exist?

33. What kinds of landforms and earth materials are almost inevitably found at the edges of mountain ranges, and why are they there?

34. Define "glacier", and contrast glaciers with things with which they might be confused.

35. How does an alpine glacier vary from upstream end to downstream end, and from surface to base?

36. What are some of the characteristic erosional landforms generated by alpine glaciers? List and describe at least five.

37. How does continental glaciation differ from alpine glaciation? There are at least three major ways in which they differ.

38. What is a piedmont glacier, and what conditions favor the formation of one?. By contrast, what conditions favor formation of a cirque glacier?

39. What are some of the depositional landforms or features left by continental glaciation? List and describe at least four.

40. What are some of the erosional features generated by continental glaciation? List and describe at least two.

41. What are kettles (in the glaciological sense), and how do they form?

42. Describe the periodicity of glaciation over the past million years (or at least the last few hundreds of thousands of years), with attention to both the timing and the rate of glacial advances and retreats. Be sure to indicate where human history (i.e., the last few thousand years) falls in this scheme. A sketch of a graph could be a useful *part* of an answer to this question.

43. What seems to have caused the periodicity of glaciation over the past few hundreds of thousands of years? Be sure to include both ultimate causes and immediate causes.

44. What is the evidence that Quaternary glaciation has been periodic, rather than unchangingly continuous?

45. What implications does the history of glaciation over the last few hundreds of thousands of years have for current concerns about global warming? A sketch of a graph or two could be a useful *part* of an answer to this question.

46. How does the alpine environment differ in its physical (rather than biological) characteristics from the lowland environment?

47. What are some of the distinctive features of trees, both individually and as groups, at high elevations?

48. What human activities affect mountainous regions? Additionally, how do human activities in mountainous regions affect lowland regions downstream?

This is a page for comments and corrections to be used in preparation of later editions of this book. These comments and corrections should be sent to

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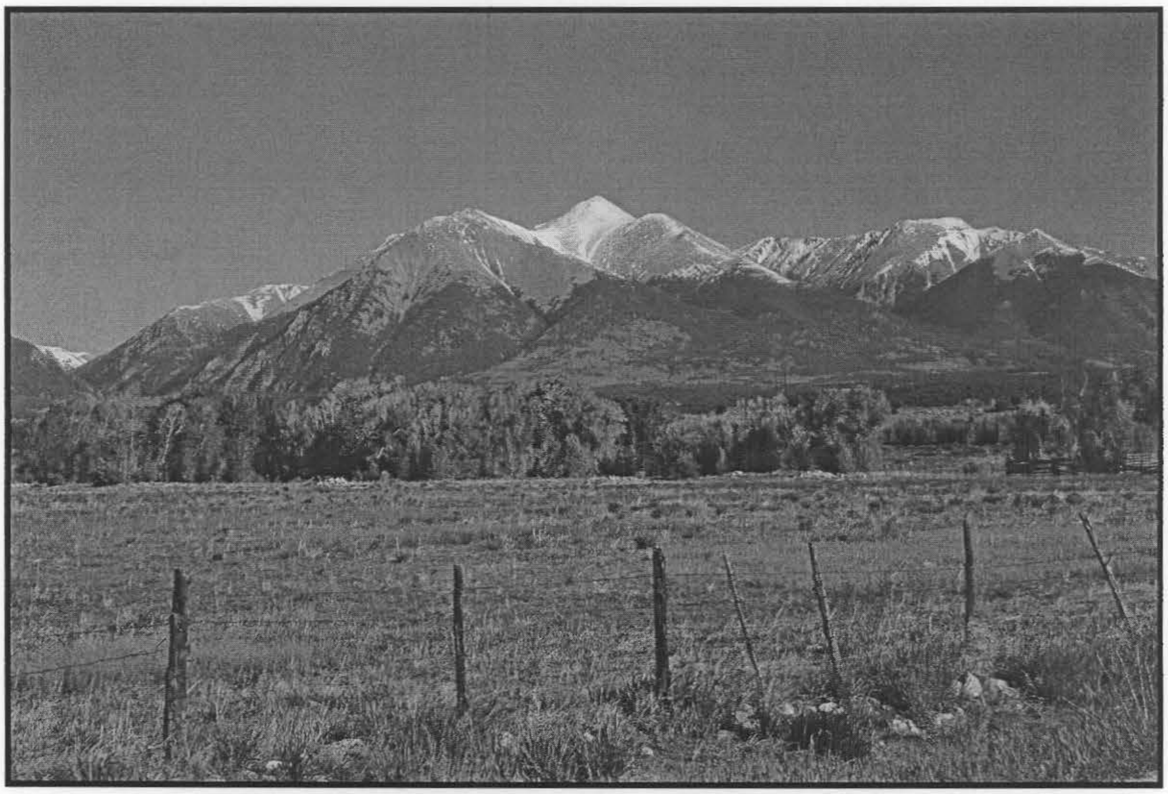

Mt. Shavano in the Sawatch Range west of Salida, Colorado