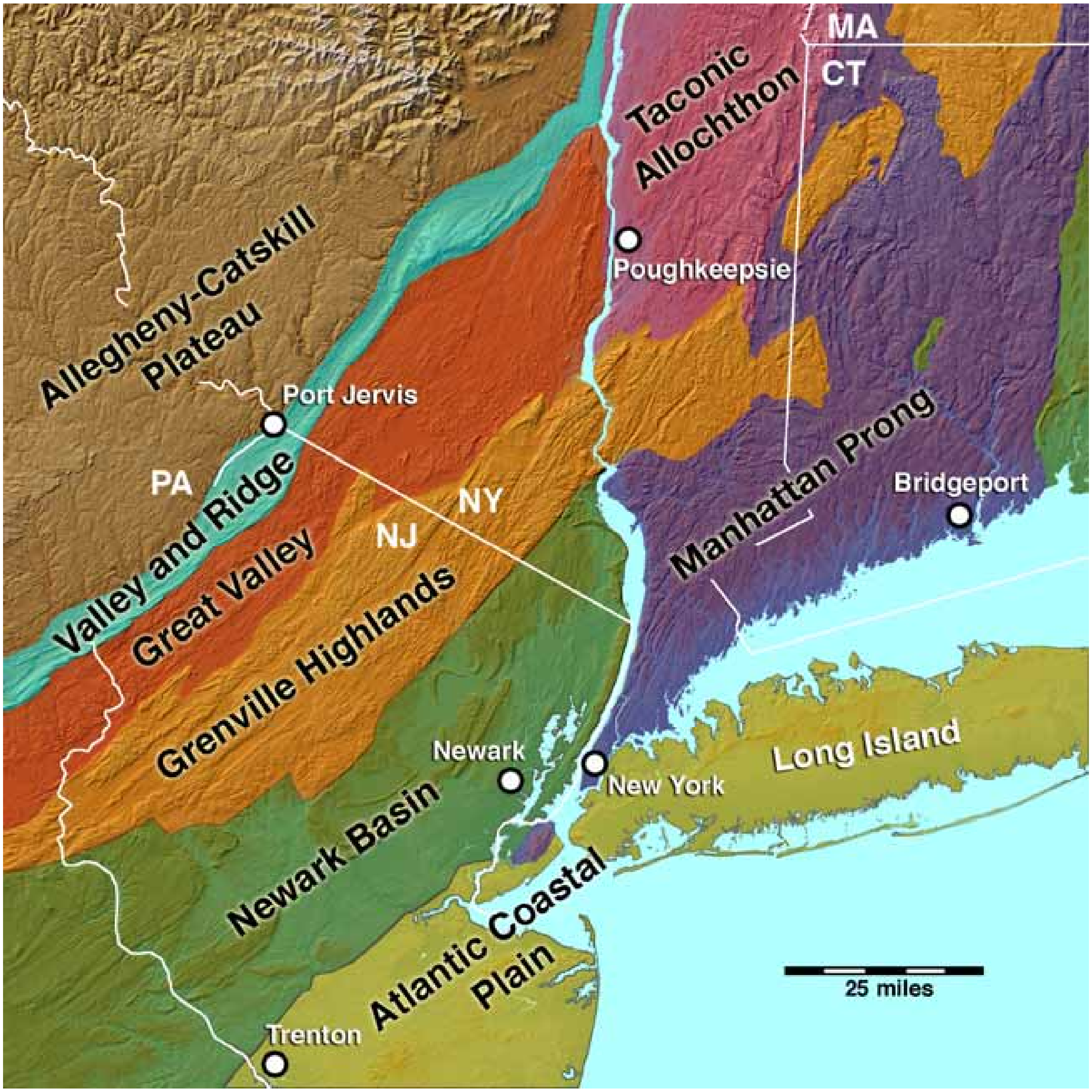
Physical Geology Laboratory Manual



J Bret Bennington, Charles Merguerian and John E. Sanders

Geology Department
Hofstra University

PHYSICAL GEOLOGY LABORATORY MANUAL
Third Edition (Revised)

by

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Department of Geology
Hofstra University

ACKNOWLEDGEMENTS

We thank the entire Geology Department faculty and all of our former Geology 1C students for helping us develop and improve these laboratory exercises and for pointing out errors in the text.

Table of Contents

Lab	Laboratory Topic	Page
	Instructions for Writing Reports	1
1	The Earth as a Planet	5
2	Physical Properties of Minerals	29
	Questions for **[R1]**, Minerals and Mineral Properties	41
3	Mineral Identification	43
4	**Mineral Practicum** and Introduction to the Three Rock Types	53
5	Igneous Rocks	57
6	Sediments and Sedimentary Rocks	75
7	Metamorphic Rocks	83
8	Integrated Rock Identification	91
	Questions for **[R2]**, Igneous, Metamorphic, and Sedimentary Rocks	98
9	**Rock Practicum** and Introduction to Maps	101
10	Introduction to Topographic Maps	105
11	Topographic Contour Maps and -Profiles	123
12	Earthquake Location and Isoseismal Maps	139
	[R3], A Future Earthqake in New York City	153
	Appendix A: Geologic History of the New York Region	159
	References	180
	Mineral and Rock Practicum Test Forms	183

How to Get Help and Find Out More about Geology at Hofstra

Information and assistance for all Geology courses at Hofstra University can be obtained from the following sources:

Email: Geology professors can be contacted via Email: Bennington (**geojbb**), Merguerian (**geocmm**), Radcliffe (**geodzr**), Wolff (**geompw**), Dieffenbach (**geowpd**), Liebling (**georsl**), Rockwell (**geoczr**), Sichko (**geomjs**), and Schaffel (**geoszs**). To Email from outside the Hofstra network, append **@hofstra.edu** to the above addresses.

Geology Club: If you are interested in doing more exploring on field trips and in learning more about geology and the environment outside of class, we invite you to join the Geology Club! Meetings are every **Wednesday** during **common hour** in **Gittleson 162**. See the web site listed above for details about club happenings this semester.

Pay us a visit! The Geology Department is located on the first floor of Gittleson Hall, Room 156 (main office, x5564). The secretary is available from 9:00 a.m. to 2:00 p.m. to answer questions and schedule appointments. Free **tutoring** is available throughout the semester and **lab materials** (mineral and rock specimens, maps) are available for additional study in **room 135**. Consult our web page or ask your lab or lecture professor for a schedule of tutor availability and faculty office hours.

INSTRUCTIONS FOR WRITING REPORTS

GENERAL

The three report-writing exercises **[R1]**, **[R2]**, and **[R3]** are to sharpen the focus of your ideas about some important geologic subjects we shall cover in Geology 01C and to give you experience in preparing formal written material such as you might submit later on to your boss on the job or to an editor of a magazine that you hope will accept and publish an article that you have written. In addition, writing the lab reports will help you learn the information you need to know to do well on the lecture exams.

Your lab instructor will read your papers and evaluate them not only for their geologic content but also for their writing style. Your work will be an example of one famous definition of education: "making mental efforts under criticism." You will make the mental efforts and your instructor will serve as a friendly critic. Your instructor will not only grade your papers but "copy edit" them as well. It is our intention to help you become better at writing, perhaps the most important skill in almost any professional career.

Format: Typed (or word processed), double spaced, ample margins all around, and written on one side of the paper only.

Write your answers in complete, well-formed sentences and paragraphs. Discuss each answer thoroughly, as if you were making a formal presentation or teaching the information to someone who was unfamiliar with it.

SCHEDULE AND TOPICS

[R1 Minerals]: Questions from list at the end of Lab 02 (to be selected by your instructor).
Due: Start of lab on Week 04.

[R2 Rocks]: Questions from list at the end of Lab 08 (to be selected by your instructor).
Due: Start of lab on Week 09.

[R3 Earthquake New York City]: Questions from exercise in Lab 12 (to be selected by your instructor).
Due: Start of Map Practicum lab, Week 13.

You must submit these papers on the dates indicated or suffer grade penalties for lateness as explained in the lab syllabus.

SOME COMMENTS ABOUT WRITING ERRORS TO AVOID

HYPHENS

1. Hyphens are needed where several words form a compound adjective before the noun they modify. [Example: “sea-level rise”; sea-level is the adjective describing the type of rise]. If you write “rise in sea level” then no hyphen is needed because sea level is itself a noun. In other words, sea and level do not automatically call for a hyphen; it depends on how they are used. A good way to remember this is with the shirt example. Referees wear black-and-white shirts; that is, the shirts are striped like a zebra. Other shirts are all white. Still others are all black. Therefore, referees' shirts are "black-and-white shirts." But, if the reference is to the stock of solid-color shirts in a clothing store, for example, then the correct usage is: "black and white shirts.”

2. Hyphens are sometimes used to show words left out. For example, “two-, three-, and four-story houses” because each number has an implied “story” after it. Hyphens are usually only used to show words left out if the last word pair in the list is hyphenated. In the example above, because “four-story” is a hyphenated compound adjective, “two-” and “three-” must also be hyphenated.

SOME NOTES ABOUT WORD USAGES

1. **Above** (Save for position reference; for number references, use “more than” or “greater than”.)

2. **Appear** (Try to avoid "It appears that..."; alt. "evidently.")

3. **Comprise vs.compose** Remember the general rule: The parts compose the whole, but the whole comprises (or is comprised of) the parts.

4. **Coarse vs. course** (Coarse refers to "large" with reference to sizes of crystals or particles in a rock; course means many things, from "golf course" to "of course."

5. **Due to** (Another one to avoid; not wrong, but is more elegant to use "because of" or "as a result of.")

6. **Each other** (Refers to situations where two and **only two** persons or things are being discussed. If more than two, then the correct form is "one another." Sounds backward, but it is correct.)

7. **Farther/further** (Always farther for comparative of the distance word "far." Comparative of "fur" is "furrier." Further is OK for "additional," such as "further developments." The copy writers for the Texaco ads flunk this one; they use the phrase "The energy to go further." If they are referring to distance, it should be "**farther**.")

8. **Imply** (Means "to express indirectly or hint at;" commonly mixed up with **infer** which means “to conclude from evidence or facts.”

9. **Independent** (spelled with final "dent" not "dant.")

10. **Infer** (Means "to conclude from evidence or facts;" commonly mixed up with **imply**.)

11. **Its and it's** (The possessive of the word "it" is "its." [“its mother was looking for it.”] "It's" is a contraction of "it is." [“It’s never too late.”])

12. **Laminae** (Plural of lamina - thin layers; never, never "laminations.")

13. **Latter** (Correct only when the **second of two** is referred to; if more than two, then the last one must be designated as the "**last**.")

14. **Occur** (Be careful not to use this too many times - it becomes annoying. The word occurring is spelled with 2 r's.)

15. **On the other hand** (No one-armed phrases here; to be used only if preceded by "on the one hand...")

16. **Over** (Use as with "above." "More than" for number references.)

17. **"Play a role."** Leave this to the theatrical folk who get paid for doing it. Try "is a factor."

18. **Presently** (Do not use for "currently" or "now;" presently means at some indefinite time in the future. "At present" is OK.)

19. **Principal vs principle** (As a noun, **principal** is the head of a school; as an adjective, it means "main" or "chief." As a noun, **principle** refers to a fundamental truth or primary doctrine; never use principle as an adjective.)

20. **Since** (Reserve for time sense; otherwise "because," or "whereas.")

21. **Stratification** (An attribute word never to be used in the plural and not to be muddled with the material things, strata.)

22. **Striae** (Plural of stria - a thin line; commonly misused in the form "striations".)

23. **There** (Means "in or at that place;" do not confuse with pronoun **their**, "meaning belonging to them." Sentences that include "there is" or "there are" can usually be re-written to become shorter and more direct.)

24. **Transpire** (“To give off through pores or openings.” Often used in the sense of “to happen” as in “The sad events of which we speak transpired last week.” This usage is considered both pompous and pretentious.)

25. **Velocity** (A vector quantity in physics implying **both a magnitude and a direction**. The magnitude part is "celerity" or "speed.")

26. Avoid the following usage: A [singular noun] of [plural noun] followed by a plural form of a verb. Example: "A **group** of minerals **are** green." In this sentence, the word **group** is the subject; therefore, the correct verb form should be **is**, not are.

    The presumptive subject, **minerals**, has been relegated to a secondary status as the object of the preposition "of." You can avoid this by using "many" or "numerous" instead of "a group of" or "a large number of."

Lab 01 - The Earth as a Planet

PURPOSE

Our first lab is intended to show you how some very basic characteristics of the Earth such as its shape, volume, and internal composition can be discovered by clever people using simple observations and measurements (no need for satellites, lasers, and other high-tech toys.) In the first part of this exercise, you will read about how the shape of the Earth can be demonstrated from observations made at its surface. You will see the basis for ancient astronomers' using the Earth's pole of rotation as a point of reference. You will also gather some numerical data from a few drawings and get some practice plotting these data on a graph. The first part of this lab ends with an account of how a pioneering genius of mathematical geography, **Eratosthenes of Alexandria**, put together some significant geometric data in ancient Egypt and used them to calculate the **circumference** of Planet Earth. Because the Earth approximates the shape of a sphere (in truth it is an “oblate spheroid” - a sphere with a bulging waist), knowing its circumference allows you to calculate its **volume**.

In the second part of this lab, you will “weigh the Earth” (estimate its **mass**) by making some measurements with a simple pendulum. Having found out the Earth’s volume and mass, you will calculate its **mean density**. Finally, you will weigh some specimens of minerals and rocks--samples of typical Earth-surface materials--in air and in water and use these weights to compute the densities of these materials. By comparing the density of the entire Earth to the densities of common surface materials you will be able to draw some interesting conclusions about the nature of the Earth’s interior.

INTRODUCTION

As a resident of the Earth during the Space Age, you may never have experienced any difficulty accepting the idea that the shape of the Earth closely approximates that of a sphere. Color TV images of the Earth taken from the Moon(Figure 1.1), for example, have convinced nearly everyone of the correctness of this concept of the Earth's shape. (The diehards are card-carrying members of the “Flat-Earth Society”, who even today maintain that those pictures shown on TV of the Earth supposedly from the Moon were only elaborately worked out hoaxes designed to promote the sale of globes.)

Evidence From Early Astronomical Observations

The famous Greek mathematician **Pythagoras** (6th century B.C.) noted in about 520 B.C. that the phases of the Earth's Moon (Figure 1.2) are best explained by the effect of directional light on the surface of a sphere. Therefore, he reasoned, if the Moon is a sphere, so too, is the Earth likely to be spherical.

In 350 B.C., another famous Greek, **Aristotle**, argued that the shape of the Earth's shadow across the Moon at the beginning of an eclipse is an arc of a circle (Figure 1.3). Only a sphere casts a circular shadow in all positions.

Figure 1.1 - The Earth viewed in December 1968 from a lunar-orbiting spacecraft at a distance of 160,000 km. The curved surface at bottom is the surface of the Moon. (NASA.)

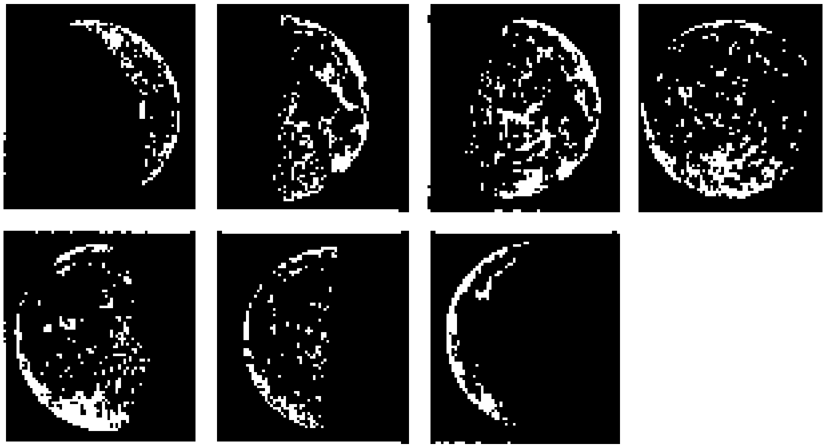


Figure 1.2 - The Lunar Phases

(from top left to bottom right: waxing crescent, first quarter, waxing gibbous, full, waning gibbous, last quarter, waning cresent)

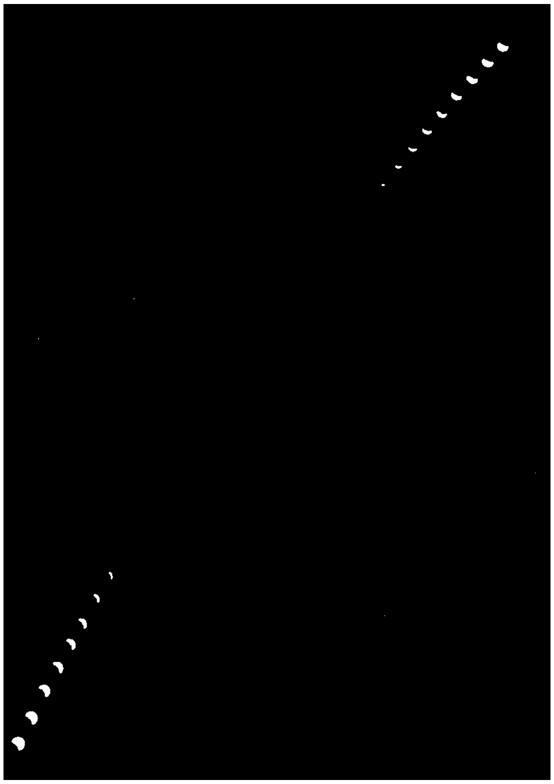


Figure 1.3 - Total lunar eclipse of Wed 09 December 1992 photographed from Jones Beach, NY with the shutter of the camera opened at intervals from 1700 h to 1810 h without advancing the film. (Bill Davis, Long Island Newsday.)

Ancient astronomers found only one star in the heavens that did not change position in the sky throughout the year: the pole star. They named the star directly above the Earth's North Pole **Polaris**. Figure 1.4 shows the results of a modern time exposure: a picture taken from a camera with its shutter kept open for many hours mounted on a telescope and pointed into the nighttime sky at the star Polaris. As the Earth rotates, the stars appear to move in circles across the sky. Notice that moving towards a point in the sky directly above the North Pole the circles become smaller and smaller. The tightest circle is made by Polaris (alas, Polaris is no longer directly above the North Pole - if it were it would appear as a single dot, the only apparently 'fixed' star in the picture.)

Even without cameras, by observing the motions of the stars, ancient astronomers found that they could identify in the sky the position directly above the Earth's pole of rotation and could use this as a fixed point of reference. Once they had established such a reference, the ancient astronomers could then define a plane perpendicular to this line. The plane perpendicular to the Earth's pole is named the Earth's Equator (Figure 1.5).

Once they could identify the pole star, astute ancient observers noticed that when they traveled from north to south, the position of the pole star in the nighttime sky changed. In Greece, for example, they saw that the pole star is higher in the sky than it is in Egypt. If the Earth's surface were flat, then at all points on the Earth the elevation of the pole star should be the same. A change in elevation of the pole star as a function of location on the surface of the Earth could be explained only if the shape of the Earth were spherical (Figure 1.6). This kind of observation enabled the ancient Greek geographers to define **latitudes** - lines on the surface of the Earth made by circles parallel to the Equator (Figure 1.7).

Other experiences relating to the shape of the Earth involved the observations of sailors, who saw ships disappear below the horizon, or found that their first views as they approached land were of the tops of hills. The shorelines appeared only after their ships had approached close to land.

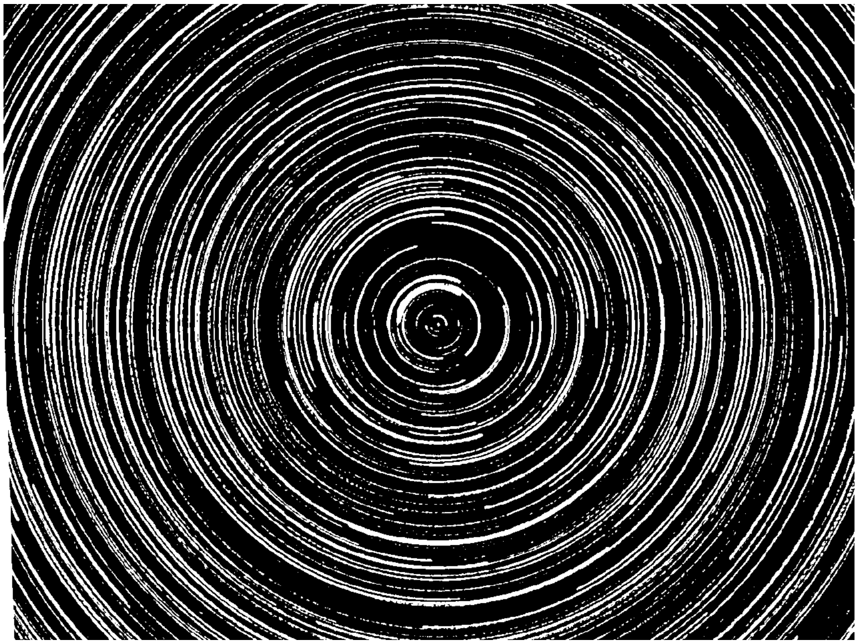


Figure 1.4 - The Earth turns on its axis as shown by this 8-hour exposure with a fixed camera pointed at the North Pole. The heavy trail near the center was made by **Polaris**, the North star. (Photograph by Fred Chappell, Lick Observatory.)

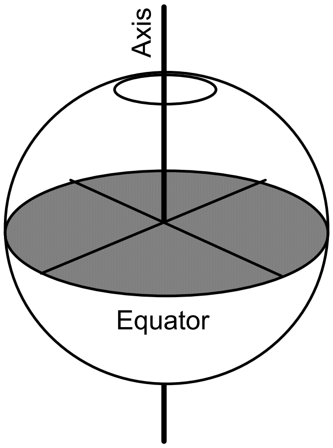


Figure 1.5 - Earth's Equator defined as a plane perpendicular to the polar axis of rotation.

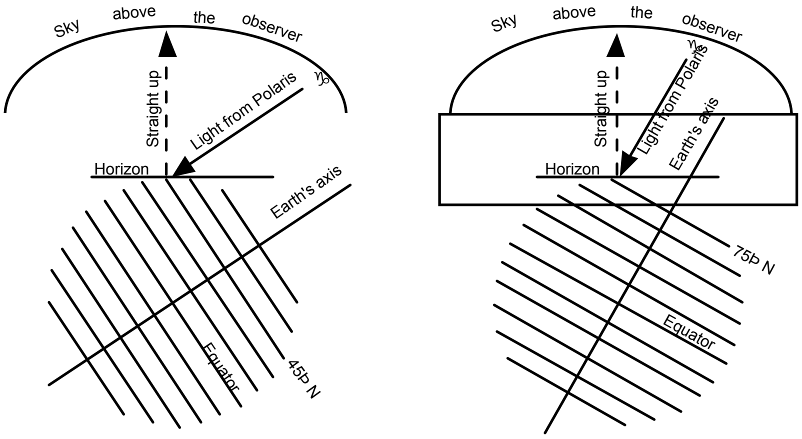


Figure 1.6 - Changing height of Polaris (the pole star) above the horizon as seen by observers at two different latitudes on the surface of the Earth.

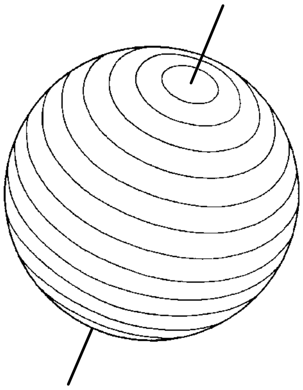


Figure 1.7 - Sketch showing circles drawn on the Earth's surface parallel to the Equator. These are named **parallels** and are the basis of latitude.

PART 1A. THE CASE OF THE "DISAPPEARING SHIP"

History does not record the first person who noticed that a ship sailing away from shore gradually disappears from view. As the ship moves farther from shore, it sinks below the horizon until only the top of the mast is visible. Even that eventually disappears. You will recreate these observations using a series of sketches of a ship traveling away from shore.

First, we need to decide on some units of measurement. We do not know what, if any, units were used in the ancient days. Accordingly, we shall adopt some hypothetical units defined as follows:

Length: Our unit is the **sinbad**, equal to the height of the Captain of the ship, "Sinbad's Symphony."

Time: Our unit is the **wineskin**, defined as the time necessary for our Captain Sinbad to drink the contents of a standard goat-hide wineskin. (You might question the accuracy and standardization of this length of time. One suspects that the duration of a wineskin might change as a function of number of wineskins drained. Furthermore, Captain Sinbad seems to have been hitting the wineskins with regularity; in the figure notice that he sails his ship out to sea backwards!)

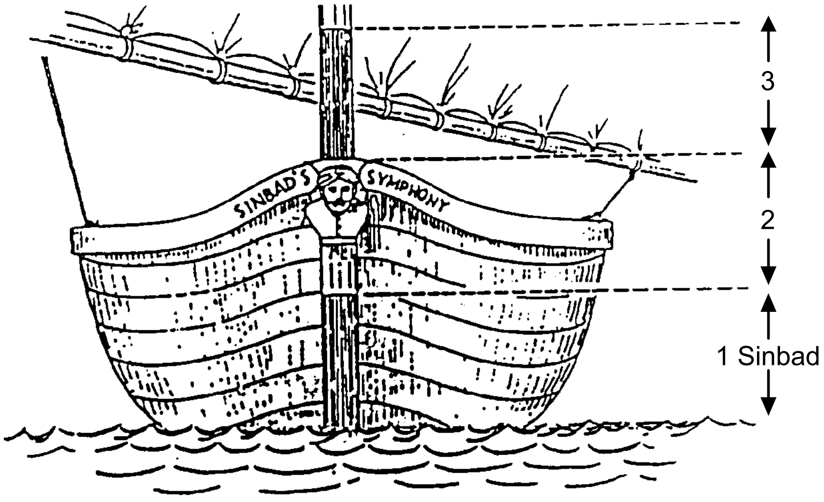
Speed: This is measured as distance traveled per unit of time, for example: miles / hour. Expressed in our sinbad-wineskin system of units, speed becomes **sinbads per wineskin** (distance divided by time).

Knowing the speed of something allows us to determine how far it travels in a given amount of time. If speed = distance / time then distance = speed x time.

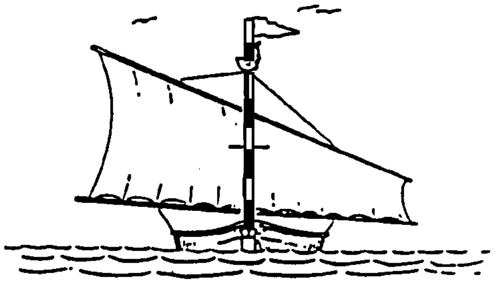
The wind on the day of our experiment allows Captain Sinbad to sail at a constant speed of **360 sinbads per wineskin**. Knowing this, study the sketches in Figure 1.8.and do the following: (report all data on Laboratory Report 1.1.)

1. For each time **t**, calculate the distance traveled by Sinbad’s Symphony and write this in column 3 on the report form.
2. For each time **t**, measure the height of the visible part of the ship above the horizon and write this in column 5.
3. For each time **t**, calculate how far the ship has dropped below the horizon and write this in column 6.

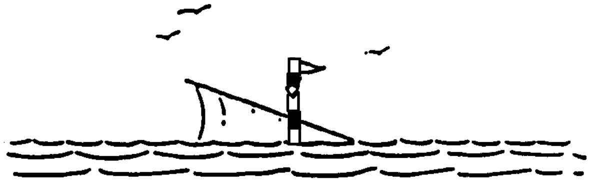
Figure 1.8 Views of the ship Sinbad's Symphony sailing away from the dock (backwards).



t=0 wineskins distance = 0 sinbads, height of ship = 10 sinbads.



t=7 wineskins. Note stripes on mast, each 1 sinbad in length.



t=18 wineskins.

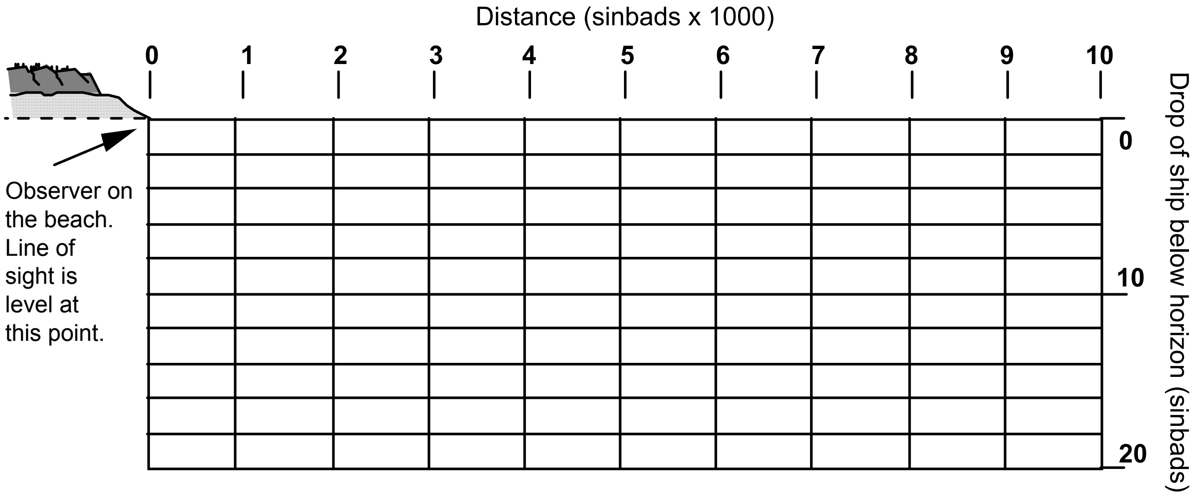
t=25 wineskins

LABORATORY REPORT 1.1
LAB 1 - EARTH AS A PLANET

DATA FORM, Ship over the horizon

Horizontal Distance			Vertical Drop		
Column 1	Column 2	Column 3	Column 4	Column 5	Column 6
Speed (sinbads per wineskin)	Time (wineskins)	Distance traveled (sinbads) Col. 1 x Col. 2	Height of ship at dock	Observed height	Drop of ship below horizon Col. 4 - Col. 5
360	**0**	**0**	**10**	**10**	**0**
			10		
			10		
			10		

After you have filled in the above table, plot on the graph below the computed horizontal distance that the ship has traveled (Column 3) vs. the drop of the ship below the horizon (Column 6).



PART Ib. CIRCUMFERENCE OF THE EARTH CALCULATED BY ERATOSTHENES OF ALEXANDRIA

Geologic books abound with versions of the remarkable calculation of the circumference of the Earth made by **Eratosthenes of Alexandria** (ca. 226 to ca. 194 B.C.). It would not be surprising if they all come from the same place, namely the 11th ed. of the Encyclopedia Britannica. We quote from that great reference, v. 9, p. 733:

> "...His greatest achievement was his measurement of the earth. Being informed that at Syene (Assuan), on the day of the summer solstice at noon, a well was lit up through all its depth, so that Syene lay on the tropic, he measured, at the same hour, the zenith distance of the sun at Alexandria (known to be 5000 stadia) to correspond to 1/50th of a great circle, and so arrived at 250,000 stadia (which he seems subsequently to have corrected to 252,000) as the circumference of the earth."

To understand what it was that Eratosthenes actually did to arrive at his estimate for the circumference of the Earth, we need to define some of the terms used in the above passage and learn a few basic facts about how the Earth is oriented in space relative to the sun.

As you probably learned in grade school, the Earth **rotates** on its **axis** once every 24 hours (to give us day and night) and **revolves** around the Sun once each year (causing the cycle of the seasons). But why are there seasons? It turns out that because of the elliptical shape of the Earth's orbit, the Earth is actually closer to the Sun when we in the Northern Hemisphere are having our winter, and farther from the Sun when we are experiencing our summer. So, distance from the Sun is obviously <u>not</u> the cause of seasons. The reason for the seasons (pardon the rhyme) is found in the **tilt of the Earth's axis**, which is about 23.5° away from vertical. Through the course of one trip around the Sun, the Earth's axis continues to point in the same direction (toward the star Polaris). However, because of the tilt, on one side of the Sun the Northern Hemisphere is tilted toward the Sun and on the other side of the Sun it is tilted away (Figure 1.12). When the Northern Hemisphere is tilted toward the Sun, it receives more direct sunlight than the Southern Hemisphere, giving the former its summer while the later experiences winter. Likewise, when the Northern Hemisphere is pointed away from the Sun, we experience our winter while the Australians are enjoying their summer.

On or near June 21st the northern pole of the axis of the Earth is pointing directly at the Sun. On this day the direct rays of the Sun (those that strike the Earth perpendicular to the surface) are at their farthest north above the Equator. At the North Pole, the Sun never sets (notice that the pole is completely covered by the day side of the summer Earth in Figure 1.12). The rest of us in the Northern Hemisphere spend the longest time in daylight of any other day in the year. For example, in Edinburgh, Scotland the Sun doesn't set until around 11:00 pm. This is the **Summer Solstice** - commonly called the "first day of summer". On the other side of the Sun on December 22 the northern pole of the Earth's axis points directly away from the Sun. The direct rays of the Sun are now striking the Earth well below the Equator. At the North Pole the

Sun never rises and the rest of the Northern Hemisphere experiences short days and long nights. This is the **Winter Solstice**.

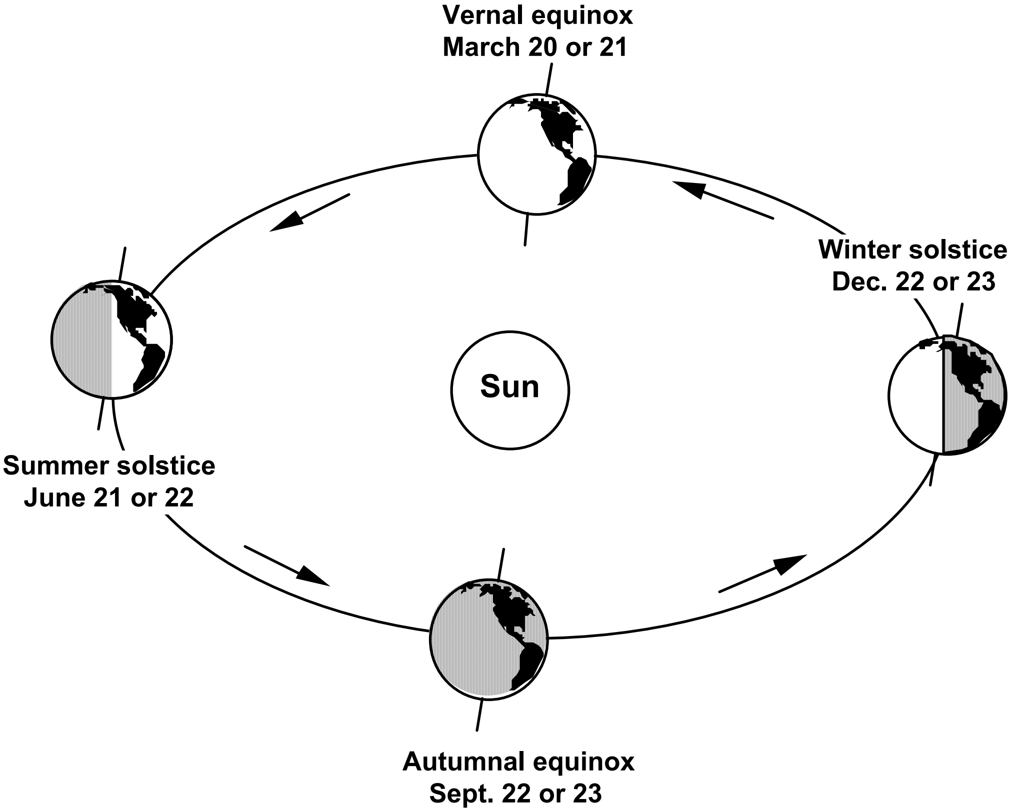


Figure 1.12 Position of the Earth during each season.

As the direct rays of the Sun migrate from their farthest point below the Equator to their farthest point above the Equator and back again through the year, there must be two days of the year when they are falling directly on the Equator. These two days are called the **equinoxes** (Figure 1.12) and fall near March 21st and September 22 (the **Vernal** or spring equinox, and the **Autumnal** or fall equinox, respectively). Again, we commonly refer to the equinoxes as the first days of the spring and fall seasons.

The latitude lines on the Earth's surface where the Sun is directly overhead during the solstices have been given special names (Figure 1.13). The northern line (summer-solstice position) is called the **Tropic of Cancer** and the southern line (winter-solstice position) is called the **Tropic of Capricorn**.

Eratosthenes must have known that the Earth was a sphere (for reasons discussed earlier in this lab). Legend has it that word reached Eratosthenes that on the longest day of the year (the summer solstice) a vertical pole in the city of Syene in southern Egypt (Figure 1.14) cast no shadow. Eratosthenes knew this meant that the Sun's rays were perpendicular to the ground at Syene on that day (a well was also dug at Syene to show that the sun's rays penetrated to its bottom.) Being a clever Greek well inclined toward geometry, Eratosthenes realized that if he could measure the angle of the Sun’s rays at Alexandria where he lived (the “zenith distance of the sun”), then he would know what fraction of the Earth’s circumference was represented by the distance from Syene to Alexandria.

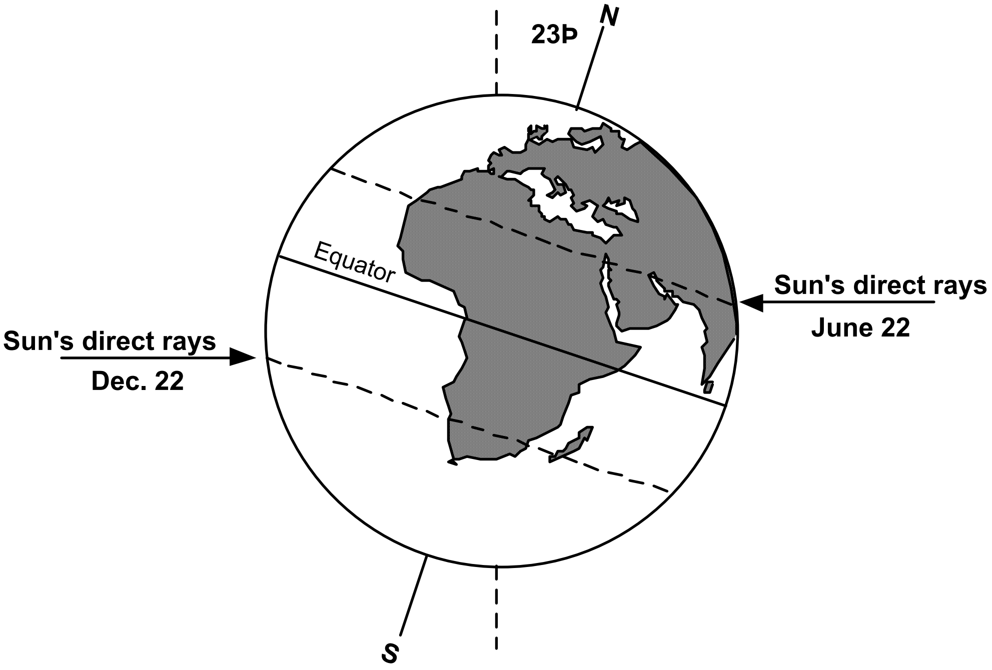


Figure 1.13 The positions of the Earth’s Tropic lines.

How do you suppose Eratosthenes figured out the angle of the Sun’s rays at Alexandria (angle a in figure 1.15)? Legend has it that he did this by noting the length of the shadow of an obelisk (see Figure 1.15). Even without knowing the height of the obelisk, he could find the angle he needed by measuring the vertical angle between the tip of the shadow and the top of the monument (angle b in Figure 1.15). If the obelisk was not leaning, then it would make a 90˚ angle to the ground. All of the angles in a triangle must add up to 180˚, so 180 - 90 - b = a. Eratosthenes’ result was 1/50th of a circle for angle A. So he reasoned that the distance from Alexandria to Syene must be 1/50th the distance around the Earth.

(Note: In fact, Eratosthenes did NOT base his calculations on the length of the shadow of an obelisk. He was more clever than that and, contrary to legend, used a small, sundial-like instrument called a **scaphe** to measure the angle of the sun’s rays at Alexandria. Nevertheless, obelisk or scaphe, the principles he employed are the same.)

For the distance from Syene to Alexandria, Eratosthenes estimated 5000 stadia. The 5000 stadia supposedly is based on the speeds of camel caravans. Given an average speed of 100 stadia per day and 50 days to make the journey from Syene to Alexandria, one arrives at 100 x 50 = 5000 stadia. (In modern units, each ancient

Egyptian stadium is reckoned to be 0.1 mile or 0.16 kilometers). Thus, 5000 stadia become 500 mi or 800 km. Multiply by 50 and the circumference of a spherical Earth is 250,000 stadia (25,000 mi; or 40,000 km; if we take the corrected number of 252,000, then this converts to 25,200 mi; or 40,320 km). Emiliani (1992) cites 40,008 km as circumference of the Earth; very close to Eratosthenes' result. So, in 250 B.C., not only did the Greeks know that the Earth was round, they also knew how large it was!

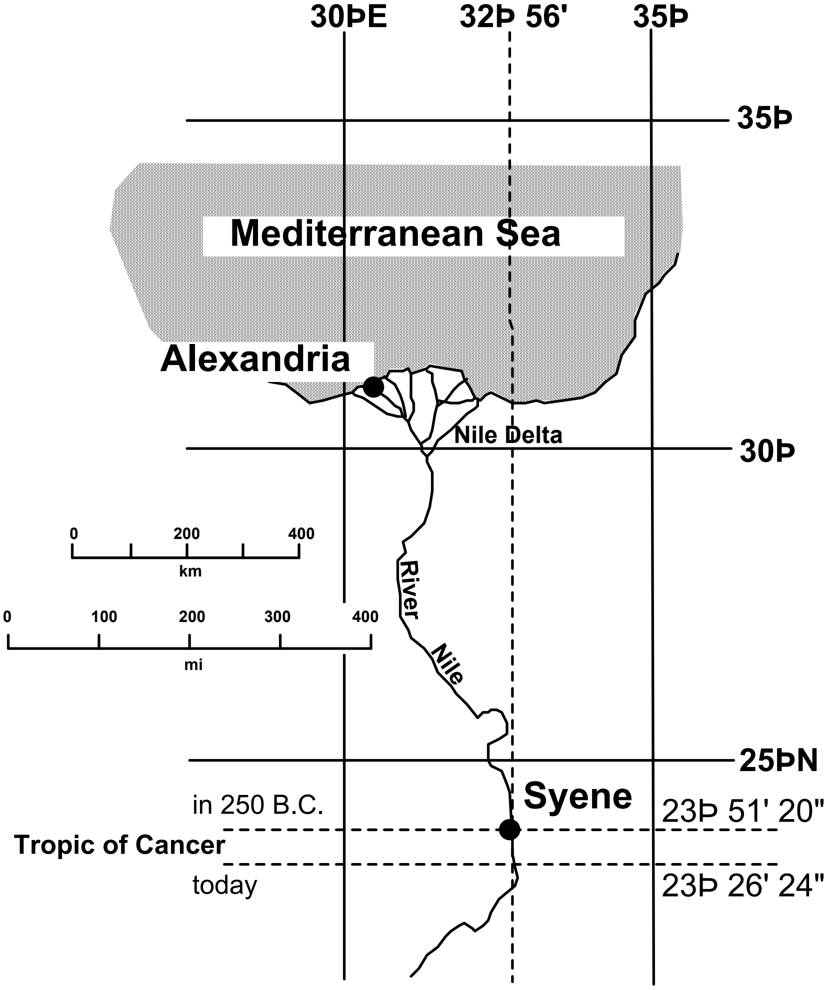
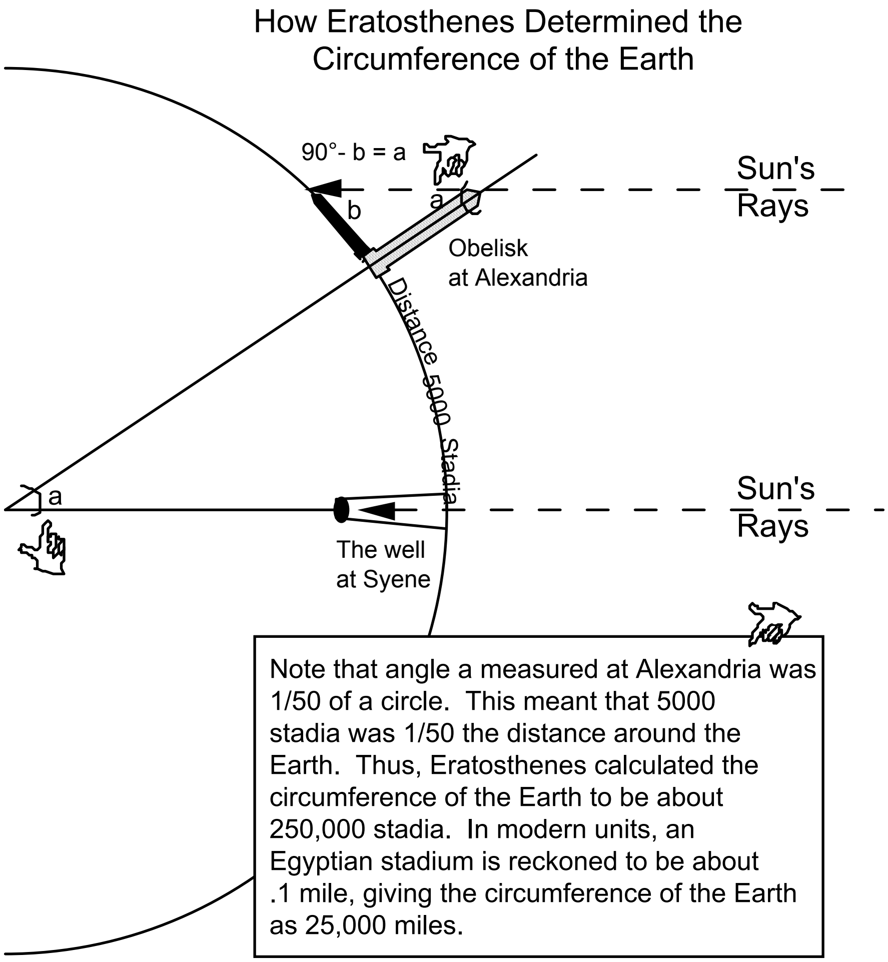


Figure 1.14 Map showing the poitions of Alexandria and Syene (now Aswan) along the River Nile in Egypt. Tropic of Cancer as in 250 B.C.


How Eratosthenes Determined the Circumference of the Earth

Note that angle a measured at Alexandria was 1/50 of a circle. This meant that 5000 stadia was 1/50 the distance around the Earth. Thus, Eratosthenes calculated the circumference of the Earth to be about 250,000 stadia. In modern units, an Egyptian stadium is reckoned to be about .1 mile, giving the circumference of the Earth as 25,000 miles.

Figure 1.15 Diagram showing the geometry that made Eratosthenes' estimate of the circumference of the Earth possible.

LABORATORY QUESTIONS 1.1
LAB 1 - EARTH AS A PLANET

Questions on Eratosthenes' calculations:

1. What would be the effect on Eratosthenes' estimate of the Earth's circumference if the camel caravan did not travel in a straight line?

2. Notice from the map of Figure 1.14 that Syene and Alexandria are not on the same longitude line (Syene is east of Alexandria). Does this geographic relationship affect Eratosthenes' calculation? Explain.

3. What is the Tropic of Cancer?

4. If the reports about the Sun's rays shining directly down the well at Syene on the day of the summer solstice are accurate, where was this well with respect to the Tropic of Cancer of 250 B.C.?

PART II. "WEIGHING" (DETERMINING THE MASS OF) THE EARTH

How can anyone possibly “weigh” something as large as the Earth? Normally, when we weigh something we put it on a scale or balance of some sort. When we do this, we are not actually measuring the mass of the object, rather, we are measuring the downward **force** exerted by the Earth’s gravitational field on the object. That force is directly proportional to the mass of the object. This is the key to our problem. One property that all mass has is **gravitational force** or **gravity**.

The critical relationship of mass to gravitational force was formulated by **Sir Isaac Newton** in the late 1600's. Before becoming something of a religeous mystic later in life, Newton performed such minor feats of mental fitness as inventing calculus (many will never forgive him for this) and determining the fundamental equation describing gravity. Gravity is the force of attraction between two bodies. Expressed as Newton's fundamental equation, the force of gravity exerted between two solid spheres (whose mass can be considered as concentrated at a single point at the centers of the spheres) is <u>proportional</u> to the product of their masses and <u>inversely proportional</u> to square of the distance between their centers:

$$\text{force} = \frac{\text{G} \cdot \text{mass}_1 \cdot \text{mass}_2}{\text{distance}^2} \qquad \text{(Eq. 1.1)}$$

In abbreviated notation we use the first letter of each component in the above equation:

f = force of gravity
m_1 and m_2 = masses of the two spheres
d = center-to-center distance between the two spheres.
G = **universal gravitational constant** (the value of gravity exerted between two spheres of 1 g each at a center-to-center distance of 1 cm).

Wait a minute, you say. What’s this **G** thing? **G** is the universal gravitational constant, which is essentially an invariable physical constant written into the fabric of the universe. It cannot be derived theoretically, it must be directly measured. The first person to successfully measure **G** in the laboratory was **Lord Henry Cavendish** (a British physicist) in 1798. Lord Cavendish built an extremely sensitive instrument that allowed him to measure the twisting of a thin wire caused by the gravitational attraction between two heavy iron balls.

Let’s now use Newton’s equation and Cavendish’s constant to estimate the mass of the Earth. First, we will modify our equation to account for the Earth and an object at its surface.

$$f = \frac{G \cdot m_{\text{Earth}} \cdot m_{\text{object}}}{r^2} \qquad \text{(Eq. 1.2)}$$

$\mathbf{M}_{\text{Earth}}$ = mass of the Earth.
$\mathbf{m}_{\text{object}}$ = mass of an object on the Earth's surface.
r = radius of the Earth (distance from the center of the object at the Earth’s surface to the center of the Earth.)

Now, we need to perform one small piece of algebraic slight of hand. If we move m_{object} to the left side of the equation by dividing each side we get:

$$\frac{f}{m_{object}} = \frac{G \cdot m_{Earth}}{r^2} \qquad \text{(Eq. 1.3)}$$

Newton's second law of motion states that the force (F) on a body equals its mass (m) times acceleration (a):

$$\mathbf{F = m \cdot a} \quad \text{or, to solve for a:} \qquad \text{(Eq. 1.4)}$$

$$\mathbf{F/m = a} \qquad \text{(Eq. 1.5)}$$

So, in Equation 1.3, f/m_{object} is an acceleration. Specifically, it is the acceleration experienced by a falling body at the surface of the Earth as a result of the Earth's gravitational attraction. We call this Earth's **g** (little g, not to be confused with big G, the gravitational constant.)

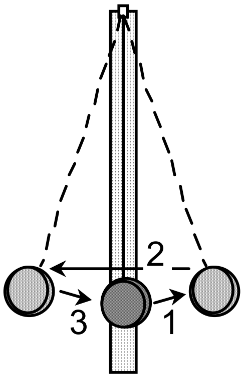
As of now, two unknowns are present in our equation, **g** (= F/m_{object}) and $\mathbf{m}_{Earth}$. We can rearrange Eq. (1.3) to solve for $\mathbf{m}_{Earth}$:

$$m_{Earth} = \frac{g \cdot r^2}{G} \qquad \text{(Eq. 1.6)}$$

We can determine a value for **g** by carrying out some measurements with a **simple pendulum** (which you are going to do) and then use this value in Eq. (1.6) to calculate $\mathbf{m}_{Earth}$, the mass of the Earth.

Using a pendulum to measure g (Earth's Gravitational Acceleration)

A pendulum is a weight on the end of a string (line, pole, etc.) permitted to swing or oscillate freely, to and fro. **Galileo** (Italian astronomer, 1564-1642) is given credit for realizing that there is a regularity to the swings of a pendulum. As a result of this regularity, pendulums have long been the basis for clocks. The complete **oscillation of a pendulum** is the swing from one side to the other and back to the original position (1-2-3 in Fig. 1.16). The time for such an oscillation is called the **period of the pendulum**. Let "**T**" represent the period of the pendulum; "**L**" the length, and "**g**" the Earth's gravitational acceleration:



The relationship between the Earth's gravitational acceleration (**g**) and the period of the pendulum is:

Fig. 1.16

$$g = \frac{4 \cdot \pi^2 \cdot L}{T^2} \qquad \text{(Eq. 1.7)}$$

g	is the Earth's gravitational acceleration.
π	is a numerical constant, equal to 3.14.
L	is the length of the pendulum measured in cm from the point of attachment to the center of mass of the weight.
T	is the period of the oscillating pendulum in seconds. This you will determine using a stopwatch.

You are now ready to determine the period T of the pendulum.

Set up the pendulum apparatus at the end of a lab table (see Fig. 1.17). For this experiment, make **L** equal exactly 100 cm. Start the pendulum oscillating (swinging) through an arc of about 30° (15° on each side of the vertical). The pendulum really only has to swing a little bit. If it swings too wildly the equation will not be accurate.

Another important precaution is to keep the pendulum swinging in a single vertical plane that is parallel to the edge of the tabletop. After the pendulum has begun to swing evenly, wait for the string to pass the (vertical) rest position, then start the stopwatch. Remember your count is **zero** when you start timing. Count **10** oscillations of the pendulum through the rest position (one oscillation equals motion of the string from the vertical to one extreme, back through the vertical again, to the other extreme, and finally back to vertical a second time) and stop the watch. Repeat this experiment three times and record the results on the computation table below. Calculate an average period for your pendulum. Record your measurements on Laboratory Report 1.2.

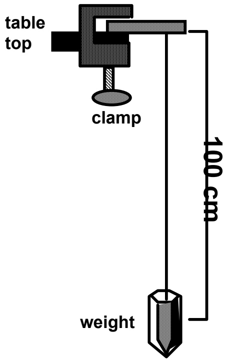


Fig. 1.17

After determining the average value of **T**, use this value to determine **g** using the equation (Eq. 1.7) below. Record your answer on Laboratory Report 1.3.

$$\mathbf{g}\ (\text{cm sec}^{-2}) = \frac{4\,(3.14)^2 \cdot 100\ (\text{cm})}{T^2\ (\text{sec}^2)} \qquad \text{(Eq. 1.7)}$$

Now, use Eratosthenes' value for the circumference of the Earth to compute the radius of the Earth. Record your answer.

$$\mathbf{C}\text{ircumference} = 2\cdot\pi\cdot\mathbf{r} \quad : \quad \mathbf{r} = \frac{\mathbf{C}}{2\cdot\pi} \qquad \text{(Eq. 1.8)}$$

Enter in Eq. 1.6 your calculated value of **g** and numbers for **r**, and **G**, the universal gravitational constant (known from precise laboratory experiments to be **$6.670\text{x}10^{-8}$ dyne cm^2/g**) to calculate $\mathbf{m}_E$, the mass of the Earth (Eq. 1.6). Record your answer.

$$\mathbf{m}_E = \frac{\mathbf{g}\ (\underline{\quad}\text{your no.; cm sec}^{-2}) \cdot \mathbf{r}\ (\underline{\quad}\text{your no.; cm})^2}{\mathbf{G}\ (6.670 \times 10^{-8}\ \text{dyne-cm}^2/\text{gm}^2)} \qquad \text{(Eq. 1.6)}$$

Congratulations! You have just determined the mass of the Earth (equivalent to weighing the Earth) by using a simple apparatus and some simple calculations!

LABORATORY REPORT 1.2
LAB 1 - EARTH AS A PLANET

A. Compute average pendulum period.

Computation of Period (T), Pendulum Experiment

Trial #	# Oscillations	Time for 10 oscillations	Time for 1 oscillation
1	**10**		
2	**10**		
3	**10**		
		Average =	

B. Use the pendulum period to compute **g**.

$$\mathbf{g}\ (\text{cm sec}^{-2}) = \frac{4\ (3.14)^2 \cdot 100\ (\text{cm})}{T^2\ (\text{sec}^2)} \qquad \text{(Eq. 1.7)}$$

Answer: g = _____________ (cm sec^{-2}).

C. Use Eratosthenes' value of 40,000 km circumference to compute **r**, the radius of the Earth in cm.

$$\mathbf{r}\ (\text{km}) = \frac{40{,}000\ \text{km}}{2 \cdot (3.14)} \qquad \mathbf{r} = \_\_\_\_\_\_\_\_\_\_\ (\text{km}) \qquad \text{(Eq. 1.8)}$$

Answer: r (km) · 100,000 cm/km = _________________ (cm)

D. Use **r**, **g** and **G** to compute the mass of the Earth.

$$\mathbf{m_E} = \frac{\mathbf{g}\ (\_\_\_\text{your no.; cm sec}^{-2}) \cdot \mathbf{r}\ (\_\_\_\text{your no.; cm})^2}{\mathbf{G}\ (6.670 \times 10^{-8}\ \text{dyne-cm}^2/\text{gm}^2)} \qquad \text{(Eq. 1.6)}$$

Answer: m$_E$ = _________________ (gm)

Worksheet for Calculations

PART III. THE DENSITY OF THE EARTH AND COMMON EARTH MATERIALS

The objective of this part of the lab is to measure the densities of samples of materials that make up most of the outer layer or **crust** of the Earth and obtain an average density for crustal material. Then you will estimate the density of the Earth as a whole and compare the two values. Before beginning, however, we should review the concepts of density and specific gravity.

Most people have a pretty good intuitive feel for what is meant by **density**. Dense objects are heavy, but in the sense that they are heavy relative to their size. Given a bowling ball and a basketball we have no trouble pronouncing the bowling ball to be the more dense of the two.

Density therefore, is the ratio of mass (or weight) to volume (Eq. 1.9). It answers the question; how much matter is packed into a given space?

$$density = \frac{mass}{volume} \qquad \text{(Eq. 1.9)}$$

Unfortunately, precise measurements of density can be tricky because it is often very difficult to measure volume accurately. For box-shaped or spherical objects volumes are easy to estimate, but how do you measure accurately the volume of something irregular, such as a rock? To get around this problem, we use another concept, similar to density, called **specific gravity**. In fact, if measured in grams per cubic centimeter, density and specific gravity are exactly the same.

Specific gravity is a unitless ratio that compares the mass of an object to the mass of an equal volume of water. The reason that it works out to be the same as density is based on the following very useful property of water:

1 cubic centimeter of water = 1 milliliter (ml) of water = 1 gram of mass

For water then, if you measure its volume you also know its mass, and vice versa. If you immerse an object in water, the object will **displace** a volume of water equivalent to the volume of the object. Furthermore, the weight of the volume of water displaced will push back against the object, causing it to become lighter by as many grams as cubic cm of volume it occupies. This is very useful because it means that to measure the volume of an object we need only measure how much less it weighs in water as opposed to air. The equation for calculating specific gravity is thus:

$$spGravity = \frac{wt_{air}}{wt_{air} - wt_{water}} \qquad \text{(Eq. 1.10)}$$

You are now ready to determine the specific gravities (densities) of some common crustal materials.

Using a triple beam balance, string, and a beaker of water, weigh each of several specimens of rock and mineral both in air and in water. Your lab instructor will show you how to do this. Record your measurements in the table on Laboratory Report 1.3. and calculate the specific gravity of each type of Earth material measured.

Calculate the overall density of the Earth on Laboratory Report 1.3 by dividing your estimate of the mass of the Earth by an estimate of the volume of the Earth based on Eratosthenes' calculations.

Compare the average densities of the rocks and minerals you measured with the average density of the Earth as a whole. Answer the questions in Laboratory Questions 1.2.

Metals		**Rocks**		**Minerals**	
Gold	19.3				
Mercury	13.6				
Nickel	8.6				
Iron	7.9				
				Olivine	3.3
		Peridotite	3.2	Hornblende	3.2
Aluminum	2.7	Basalt	2.9		
		Granite	2.7	Calcite	2.7
		Limestone	2.7	Quartz	2.7
		Mudstone	2.6	Feldspar	2.6

Table 1.1. Specific gravities of selected metals, rocks, and minerals.

LABORATORY REPORT 1.3
LAB 1 - EARTH AS A PLANET

A. Determine the density of several common crustal rocks and an average overall density for crustal rock.

Relative Density (Specific Gravity) of Earth's Crustal Materials			
Rock or Mineral Type	Weight of Specimen in air	in water	Specific Gravity (Density) $\frac{\text{Wt air}}{\text{Wt air - Wt water}}$
Quartz			
Feldspar			
Granite			
Basalt			
Average Density of Crustal Rocks =			

B. Determine the average density of the whole Earth.

First, calculate the volume of the Earth assuming that its shape approximates a sphere. Use the value for **r** (radius of the Earth) you calculated on Laboratory Report 1.2.

$$\text{Volume of a sphere, } \mathbf{V} = \frac{4 \cdot \pi \cdot \mathbf{r}\,(\text{cm})^3}{3} \qquad \text{(Eq. 1.11)}$$

Answer: $\mathbf{V}_E$= _____________ $(\text{cm})^3$

Now, compute the average density of the Earth (weight per unit volume):

$$\mathbf{D}_E \text{ avg} = \mathbf{M}_E/\mathbf{V}_E = \frac{\text{Mass of the Earth (gm)}}{\text{Volume of the Earth (cm)}^3} \qquad \text{(Eq. 1.12)}$$

Answer: $\mathbf{D}_E$= _____________ gm / $(\text{cm})^3$

LABORATORY QUESTIONS 1.2
LAB 1 - EARTH AS A PLANET

1. Compare the average density of the specimens measured with the average density of the Earth.

2. From a comparison of these two average densities, would you describe the Earth as being a homogeneous body?

3. From this comparison of densities of surface rocks with the average density of the Earth, what can be inferred about the density of the Earth's deep interior?

4. Look over the values of specific gravity given in Table 1.1. What material or materials might the deep interior of the Earth be composed of? Assume that the interior density and the crustal density of the Earth should average out to give approximately the whole density you calculated.

Lab 2 - Physical Properties of Minerals

PURPOSE

The purpose of today's laboratory is to introduce students to the techniques of mineral identification. However, we will not identify minerals this week. Rather, we will define what a mineral is and illustrate the basic physical properties of minerals. By the end of today's lab you will have learned both how to **observe** and **record** the basic physical properties of minerals. Next week, in a true Sherlock Holmesian experience, you will observe physical properties and identify by inference, using tables and flow charts, twenty important rock-forming minerals.

INTRODUCTION

Are minerals and rocks the same? Of course not, although commonly they are mistaken for each other and used interchangeably by those who are not knowledgeable about Earth materials. If we visualize rocks as being the "words" of the geologic "language," then minerals would be the "letters" of the geologic "alphabet." **Rocks are composed of minerals!** That is, rocks are aggregates or mixtures of one or more minerals. Therefore, of the two, minerals are the more-fundamental basic building blocks. Minerals are composed of submicroscopic particles called **atoms** or **elements** (uncharged) and **ions** (charged) and these particles consist of even smaller units of mass called **electrons**, **neutrons**, **protons**, and a host of **subatomic particles** too numerous to mention here. In subsequent laboratory sessions, we will take up the various kinds of rocks. But first, we must study their component minerals.

Our treatment of minerals (and rocks) is by necessity brief and condensed. You are therefore required to supplement the laboratory study of each by the assigned readings in the textbook. Later, in Chapter 4 we offer some hints for identifying minerals in rocks and Chapters 5, 6, and 7 delve into identification of the igneous, sedimentary, and metamorphic rocks in detail. Chapter 8 includes some techniques for identifying rocks in general and for distinguishing among the three types of rock.

MINERALS

Definition:

A **mineral** is a naturally occurring, inorganic, crystalline solid (not amorphous), with a chemical composition that lies within fixed definable limits, that possesses a characteristic set of diagnostic physical properties. Nearly 3,000 different minerals are now recognized yet the average card-carrying geologist can "get by" with the ability to identify a handful of common rock-forming minerals. The essential characteristic of a mineral is that it is **a solid whose ions are arranged in a definite lattice**–no lattice, no mineral. This eliminates any synthetic chemical substances manufactured in a laboratory or in a student dorm room.

To summarize: for a substance to be termed a mineral, it must be:

1. Crystalline - The term “crystalline" means that a distinctive, orderly lattice exists. This orderly arrangement of the particles (ions and atoms) composing the mineral follows laws of geometric symmetry and may involve single atoms or a combination of atoms (molecules).

2. Inorganic - This term excludes from the definition of minerals all materials that organic substances that are not **biocrystals**. The exclusion is particularly aimed at carbon-hydrogen-oxygen compounds, the happy hunting grounds of "organic" chemistry. A substance such as amber, which is commonly used in jewelry, is not considered to be a mineral. For the same reason, coal is not a mineral, yet, carbon in the form of graphite and diamond are minerals because each possesses an important and distinctive characteristic – a crystal lattice.

3. Distinctive chemical composition - Minerals may be composed of a single element (carbon, as in graphite and diamond; gold, silver, copper, or sulfur, for example) or combinations of elements. Such combinations range from simple to highly complex. Among combinations, the composition may vary, but the variation is within specific limits.

4. Occurring in Nature - This is another way of saying "a naturally occurring solid." Rubies, sapphires, etc., normally considered minerals, that have been manufactured as imitation or synthetic gems, are excluded.

To summarize, the lattice and elements present in a mineral control the physical properties of a mineral. To illustrate the importance of the lattice on physical properties, consider the examples of graphite and diamond. Both consist of a single element, carbon. But in graphite, the atoms are arranged in such a way that the mineral is soft and flaky. In diamond, a different geometric arrangement of atoms yields the hardest substance known on our planet. We cannot emphasize too often that the lattice affects all aspects of a mineral's properties. Many kinds of sophisticated instruments are available for determining exact lattice information from which any mineral can be identified. Two such instruments are an X-ray diffractometer and a polarizing microscope. Use of these and other instruments is beyond the scope of this basic course in Geology (but we will show you examples of what these instruments enable geologists to find out about minerals).

PHYSICAL PROPERTIES

We will concentrate on those mineral properties that can be identified by visual inspection and by making diagnostic tests using the simple "tools" (available in your Geology Kits) on small specimens. (If it is possible to pick up a specimen and hold it in one's hand, geologists call it a "hand specimen.") Most geologists routinely use these same tests and tools both in the laboratory and in the field to identify the common minerals that form most rocks.

Below is a list of the important lattice-controlled properties that we discuss with the most-useful properties written in **CAPITAL LETTERS**. The others are of lesser importance, and some can be considered "exotic physical properties." Keep in mind that you should always proceed on the basis that you are dealing with an unknown.

LUSTER
COLOR
HARDNESS
STREAK
CLEAVAGE
Crystal form
Twinning
Play of colors
Specific Gravity
Magnetism
Diaphaneity
Flexibility / Elasticity
Brittleness / Tenacity
Odor
Taste
Feel

LUSTER

The property known as luster refers to the way a mineral reflects light. Luster is a property that can be determined in a general way simply by looking at a specimen. In other words, it is an "eyeball" property difficult to explain in words. We are reminded, for example, how difficult it is to **describe in words** how chicken tastes.

Luster can be treated on two levels: (1) quantitative, and (2) qualitative. Quantitatively, luster is essentially the measured intensity of the reflection of light from a fresh surface of the mineral. Special instruments, known as reflected light microscopes, are available for measuring light reflected from minerals.

For our purposes, however, we can use the qualitative method by noting general categories. The degrees of intensity of reflected light can range from high to low or from splendent to shining, glistening, and glimmering through dull or dead (non-reflective). We will compare the way the mineral reflects light with the qualitative reflectivity of substances known to **most** people. Some examples are:

Qualitative Categories of Luster

Metallic: the luster of metal
Non-metallic:
    adamantine: the luster of diamond
    vitreous: the luster of broken glass
    shiny: just what it sounds like
    porcellanous: the luster of glazed porcelain
    resinous: the luster of yellow resin
    greasy: the luster of oil or grease
    pearly: the luster of pearl
    silky: like silk
    earthy: like a lump of broken sod
    dull: the opposite of shiny

COLOR

The color is generally the first thing one notices about a mineral. Color is an obvious feature that can be determined even without touching the specimen. In some cases, color is a reliable property for identifying minerals. For example, the minerals in the feldspar family can be sorted into categories by color. **Potassium feldspar** (or **orthoclase**) is cream colored, greenish, pink, even reddish. The members of the **plagioclase group** tend to be white, gray, bluish or even transparent and glassy (vitreous luster).

Likewise, minerals in the mica family can be identified by color. Whitish mica is **muscovite**; black - **biotite**; brown - **phlogopite**; green - **chlorite**; etc. However, even here, caution is required because slight weathering or tarnish can alter the color. Biotite can take on a brown or golden hue; chlorite can lighten to be confused with muscovite.

By contrast, **quartz** is an example of a single mineral that boasts numerous colors and hues that are not significant lattice-related properties but functions of small amounts of impurities. Milky quartz is white; smoky- or cairngorn quartz is black; amethyst is purple; citrine is lemon-yellow; rose quartz is pink to light red; jasper is dark red; and just plain old garden-variety quartz is clear, transparent, and colorless. As this list shows, quartz comes in so many colors that color alone is an almost-worthless property for identifying quartz. Calcite is an example of another mineral displaying many colors (white, pink, green, black, blue, and clear and transparent, for example).

Although color is an important physical property that should be recorded, be aware of the minerals in which it is a reliable diagnostic property and of those in which color is not diagnostic but more likely to be a trap for the unwary. The diagnostic mineral charts in Lab 4 deemphasize the importance of color in non-metallic minerals by designating dark- from light-colored mineral categories. As such dark-colored minerals are black, gray, dark green, dark blue, and dark red. Light-colored minerals are white, off-white, yellow, light green, light blue, pink, and translucent. All metallic minerals are considered dark colored. In general, the dark-colored minerals fall into a chemical class called **mafic** (rich in iron and magnesium) and the light-colored minerals form the **felsic** chemical class (poor in iron and magnesium). Rocks are subdivided into these two basic chemical schemes as well.

HARDNESS

The hardness of a mineral is not its ability to withstand shock such as the blow of a hammer, but its **resistance to scratching or abrasion**. Hardness is determined by testing if one substance can scratch another--no more. (A mineral's shock resistance is its tenacity, to be discussed later.) The hardness test is done by using the point or edge of the testing item (usually a glass plate, nail, or knife blade) against a flat surface of the unknown. Exert enough pressure to try to scratch the mineral being tested.

Hardness in a numerical (but relative) form is based on a scale devised by the German-born mineralogist, **Friedrich Mohs** (1773-1839), who spent most of his career scratching away in Vienna, Austria (Vienna's climate is very dry). His scheme, now known as the Mohs (not Moh's) Scale of Hardness, starts with a soft mineral (talc) as No. 1 and extends to the hardest mineral (diamond) as No. 10. (The true hardness gap between No. 10 and No. 9 is greater than the gap between No. 9 and No. 1). The numbered minerals in this scale are known as the **scale-of-hardness minerals**. The numbers have been assigned in such a way that a mineral having a higher numerical value can scratch any mineral having a smaller numerical value (No. 10 will scratch Nos. 9 through 1, etc., but not vice versa). Mohs selected these scale-of-hardness minerals because they represent the most-common minerals displaying the specific hardness numbers indicated. In terms of absolute hardness, the differences between successive numbered scale-of-hardness minerals is not uniform, but increases rapidly above hardness 7 because of the compactness and internal bonding of the lattice. Most precious and semi-precious gems exhibit hardness 8, 9, and 10 and are relatively scarce.

The **Mohs Scale of Hardness** is as follows:

1. Talc (softest)
2. Gypsum
3. Calcite
4. Fluorite
5. Apatite
6. Orthoclase feldspar
7. Quartz
8. Topaz
9. Corundum
10. Diamond (hardest)

You should do whatever you have to in order to memorize the names of these minerals and their Mohs hardness numbers. We will be using over and over again the minerals numbered 1 through 7.

One can purchase hardness-testing sets in which numbered scribers have been made with each of the scale-of-hardness minerals. Lacking such a set of scale-of-hardness scribers, for most purposes, including field identification, it is possible to fall back to a practical hardness scale as is listed below. This simple scale is based on common items normally available at all times. The hardness numbers are expressed in terms of Mohs' Scale. Once you have determined their hardness against materials of known Mohs numbers, you can add other items (keys, pens, plasticware, etc.) to your list of testing implements. (Your Geology Kit contains a small glass plate, a nail, and a knife blade.)

Mohs Practical Hardness Scale

6.0 = Most hard steel
6.0 = Nonglazed porcelain
5.0 - 5.5 = Glass plate
3.0 = Copper Coin (pre-1982 cent)
2.5 = Fingernail

This scale is very useful for making hardness tests. In this lab we will use the Practical Hardness Scale to subdivide minerals into three general groups:

Hard - minerals harder than 5.0 (these will scratch glass).

Soft - minerals softer than 2.5 (these you can scratch with a fingernail).

Medium - minerals between 2.5 and 5.0 (you can’t scratch with a fingernail but will not scratch glass).

One of the columns to be filled in on the exercise sheets for Labs 3 and 4 and on the answer sheet in the Mineral Practicum is hardness. In that column you will write down the results of your hardness tests using the Practical Hardness Scale listed above.

STREAK

Whereas color refers to the bulk property of a mineral, **streak** is the color of a powder made from the mineral. (The property of "streak" in minerals is not to be confused with the definition of "streak" invented a few years ago by college students running around college campuses "in the altogether.") The ideal way to determine a mineral's streak is to grind a specimen into a powder using a mortar and pestle. If we all did this every time we wanted to check on the streak of a mineral, our nice collection would very rapidly disappear. Fortunately, we can obtain the streak of a mineral by rubbing one specimen at a time firmly against a piece of nonglazed porcelain known as a "streak plate." (Your Geology Kit contains a small streak plate but in a streak-test emergency the back side of a common bathroom tile can be used. Be sure to check with the owner or principal user of your bathroom before removing the tiles!)

The reaction between the mineral and streak plate is a tiny trail of colored powder--the streak of the mineral. After many tests, its original white surface may be obscured with the powder of many minerals. It is possible to wash streak plates and reuse them many times. (Rub a little scouring powder on the wet surface of a used streak plate and it will become like new.)

Incidentally, the use of a streak plate serves a dual purpose. Because nonglazed porcelain is made from feldspar, if a mineral leaves a streak it is also softer than feldspar (= 6 on Mohs' Scale of Hardness). Naturally, minerals that scratch the streak plate are harder than 6.

One caution about streak: hard non-metallic minerals may give what looks like a white streak. In reality what is happening is that these minerals are scratching the streak plate; the white powder is not from the mineral but from the streak plate itself. As such, learn to distinguish between a white streak (unknown softer than streak plate), a colorless streak (unknown softer than streak plate), and no streak (unknown harder than streak plate). As a matter of standard procedure, test all metallic minerals with your streak plate as streak is very diagnostic in these instances.

In most cases (75% +), the bulk color of the mineral in hand specimen will be the same as the color of the streak. However, divergences between bulk color and the streak may be so startling as to be a "dead give-away" for identifying that mineral. For example, some varieties of **hematite** display a glistening black metallic luster but the red-brown streak will always enable you to distinguish hematite from shiny black **limonite** whose streak is yellowish brown. The streak of **magnetite** is black. Streak is much more diagnostic than the outward color of a mineral - use it wisely!

THE WAYS MINERALS BREAK: FRACTURE VS. CLEAVAGE

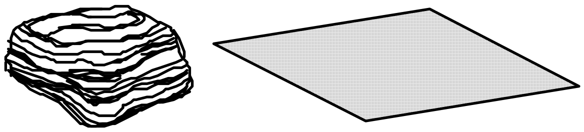
The way a mineral breaks is a first-order lattice-controlled property and thus is extremely useful, if not of paramount importance, in mineral identification. The broken surface may be irregular (defined as "**fracture**") or along one or more planes that are parallel to a zone of weakness in the mineral lattice (defined as "**cleavage**").

Fracture surfaces may be even, uneven, or irregular, including fibrous, splintery, and earthy. (One would describe the way wood breaks as splintery fracture.) A distinctive kind of fracture is along a smooth, curved surface that resembles the inside of some smooth clam shells. Such curved fracture surfaces are referred to as **conchoidal** (from the French word for a shell, namely conch).

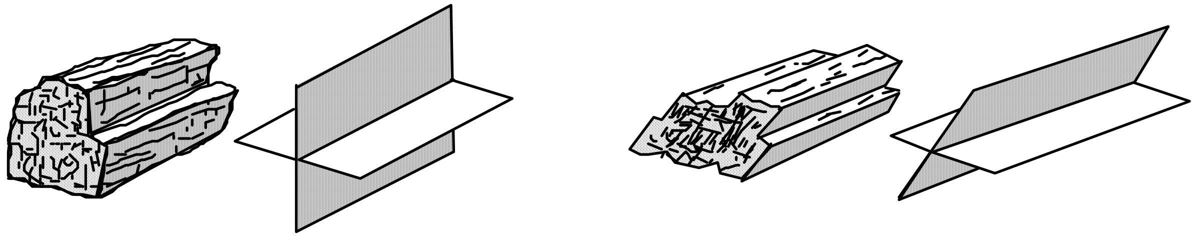
Of all the mineral properties we shall discuss, cleavage is the most difficult for newcomers to the study of minerals to understand. But, be assured that all your efforts to understand cleavage will be richly rewarded. After you have mastered the ins and outs of cleavage, you will use it routinely in the lab or the field as the most-diagnostic physical property.

The first point of difficulty is to realize that "a" cleavage is not just a particular plane surface but rather is "**a direction**." In other words, the concept of cleavage includes not only a single plane surface, but all plane surfaces that are parallel to it. Start by visualizing any given plane. An infinite number of broken segments of a mineral may be parallel to this particular plane (and by definition, the direction that is parallel to a plane of weakness in the lattice). Thus, you must distinguish between the actual plane surfaces along which the mineral splits, which may be numerous, and the one single cleavage direction (i.e. cleavage plane) to which all these surfaces are parallel. For example, the top and bottom of a cube form two parallel surfaces. But because these surfaces are parallel, they can be defined by specifying the orientation of one single plane.

If a mineral cleaves rather than fractures, then the result is two relatively flat smooth surfaces, one on each half of the split mineral. Further, the two surfaces will be mirror images of each other (symmetrical). Because these surfaces (and all other segments of surfaces parallel to them) are smooth, they will reflect light and often do so more strongly than the rest of the specimen.

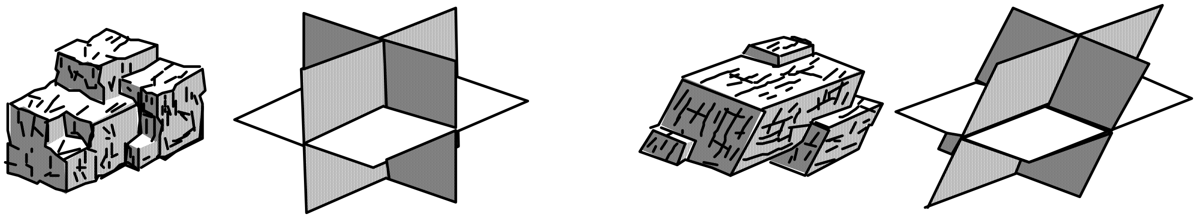


A. One direction of cleavage. Mica, graphite, and talc are examples.



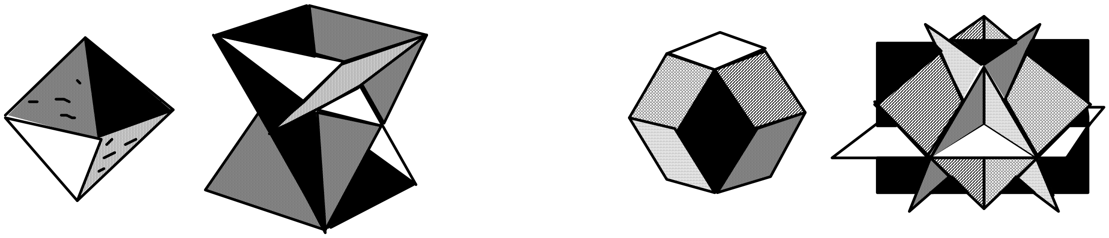
B. Two directions of cleavage that intersect at 90° angles. Plagioclase is an example.

C. Two directions of cleavage that do not intersect at 90° angles. Hornblende is an example.



D. Three directions of cleavage that intersect at 90° angles. Halite and galena are examples.

E. Three directions of cleavage that do not intersect at 90° angles. Calcite is an example.



F. Four directions of cleavage. Fluorite is an example.

G. Six directions of cleavage. Sphalerite is an example.

Figure 2.1 - Examples of cleavage directions found in the common rock-forming minerals.

We repeat again the fundamental point that a cleavage direction is the physical manifestation of a plane of weakness within the mineral lattice. This weakness results from a planar alignment of weak bonds in the lattice. The strength or weakness of these aligned bonds affects the various degrees of perfection or imperfection of the cleavage (degree of smoothness/flatness of the surfaces).

After you have learned to recognize cleavage surfaces, you must deal with two final points of difficulty about cleavage. These difficulties can be expressed by two questions: (1) How many directions of cleavage are present?, and, (2) How are the cleavage directions oriented with respect to each other (if only 2 directions are present) or to one another (if 3 or more directions are present)?

Some minerals, notably those in the mica and clay families having sheet-structure lattices, display only one direction of cleavage and it is likely to be perfect. Such cleavage is commonly termed **basal cleavage**. Other minerals possess two, three, four or six cleavage directions. No mineral exists that displays five directions of cleavage (nor pentameral [five-sided] lattice symmetry, for that matter).

Minerals having three or more cleavage directions break into cleavage fragments that produce distinct, repetitive geometric shapes. Breakage along three cleavages at right angles, as in halite or galena, yields cubes. Not surprisingly, such cleavage is described as being **cubic**. If the three cleavage directions are not at right angles, the cleavage fragments may be tiny rhombs, as in the **rhombohedral cleavage** of the carbonate minerals, calcite and dolomite.

As an exercise in lab, compare the unknown examples in the cleavage sets to the cleavage directions illustrated in Figure 2.1. Convince yourselves that cleavage, although variable in quality, is penetrative, repetitive, and geometrically regular within a sample.

Now that you have come this far, you must be prepared to face one more hurdle: How can one tell these geometric cleavage fragments from **crystals**, which are geometric solids bounded by natural smooth surfaces known as crystal faces? For help on this question, read on into the following section entitled "Crystal Form."

Crystal Form

The external form of a mineral is a function of several factors. In the simplest, ideal case, the external form is a direct outward expression of the internal mineral lattice. In this case, the mineral displays beautiful **crystal faces** that are planes and form regular sharp boundaries with adjoining planes that are arranged in clearly defined geometric solids (and we refer to such an object simply as "a crystal"). But out there in the real world, other factors may be at work that can affect whether or not a growing mineral lattice is able to become a recognizable crystal form. True crystal forms can develop only where (1) the mineral lattice was able to grow uninhibited in all directions, as is the case where the growing lattice is surrounded by empty space or by a liquid; or (2) the power of crystallization of the growing lattice is so great that despite all obstacles, the mineral develops its own crystal faces (and in the process may prevent adjacent growing mineral lattices from developing their crystal faces). Such crystallographic power is common during metamorphic-crystal growth as will be discussed later in this manual.

Each crystal face is a visible external expression of the internal lattice structure of the mineral. The orderly arrangement of the atoms repeated continuously in certain directions within the lattice controls the directions of crystal faces. Adjacent flat crystal faces intersect each other so that they resemble faceted gemstones. But adjoining crystal faces always form specific angles that are constant for any given mineral irrespective of crystal size. This is known as the "**Law of Constancy of Interfacial Angles**" first proposed by Nicholas Steno in 1667.

Organisms are capable of secreting minerals such as calcite. But, organisms have acquired the special talent of being able to shape the outside of the growing lattice into non-crystal faces that suit some particular purpose in the organism--for example, into shapes such as teeth, bones, shells, or eye lenses. Minerals grown within the tissues of living organisms are called **biocrystals**.

A third external category results when many lattices are growing simultaneously and they all interfere with one another in such a way that no lattice develops crystal faces. The chaotic result of such interference is described as "**compromise boundaries**." (If you have ever been packed into a subway car at rush hour, you may have experienced something akin to the "compromise boundaries" of mineral lattices.) The important point from this short summary of what can happen to growing mineral lattices is that on the inside, the distinctive mineral lattice is always present, but external form may be variable. The outside may consist of crystal faces, be biocrystal shapes, or consist of irregular surfaces. **Yet, no matter what the external form, the diagnostic properties of the mineral, determined by the mineral lattice inside (in particular its cleavage), are always present.**

Now on to the big question that we have been postponing: How can one distinguish crystal faces from cleavage surfaces? Let us try to answer this by taking stock of the fundamentals. Both are first-order reflections of significant lattice properties. Crystal faces reflect the entire lattice; cleavages reflect only weak parts of the lattice (the planes of weak bonding). This means that crystal faces typically are more numerous than are cleavage surfaces. But what about cubes? Halite crystals are cubes; so are halite cleavage fragments. What do we do now?

On minerals having vitreous luster, it is sometimes possible to see cleavages expressed as small, incipient parallel cracks extending into the specimen. Look carefully for such cracks using the hand lens from your geology kits or one of the departmental binocular microscopes on the lab benches. These cracks are absolutely diagnostic expressions of cleavage direction(s). Plane surfaces that are not parallel to such internal cracks are crystal faces. Finally, crystal faces are generally flatter and/or contain irregular step-like growth surfaces or surface impurities.

Crystal form can be extremely important in identifying minerals. However, minerals displaying well-developed crystal forms are not very common. In most cases this will not be a useful property for mineral identification.

Twinning

When two or more parts of the same mineral lattice become intergrown, the phenomenon is referred to as "**twinning**." Parts of a twinned crystal may penetrate into another part; such cases are referred to as "penetration twinning" (as in **staurolite**, an important metamorphic mineral). The crystal lattices of one part of a twin can be parallel to the lattice of the other part. Or lattice parts can be rotated through 180˚ with respect to the lattice of the other part. Such parallel twinning can result in twin zones or polysynthetic twin planes that may be visible as planar patterns on the external surfaces of the crystal (crystal faces or cleavage planes).

Parallel-type internal twinning is diagnostic of the plagioclase group of feldspars. The expression of this kind of internal twinning is a series of closely spaced parallel lines known as **twinning striae** that can be seen on cleavage planes. Other twinning striae appear as lines on crystal faces of pyrite.

Play of Colors

The property known as "**play of colors**" is an expression of internal irridescense. To see if a mineral displays this property, rotate the specimen into different orientations under strong light. The "play" is the internal display of some colors of the spectrum. A striking blue irridescense is most peculiar to the varieties of plagioclase named albite (also known as moonstone) and in labradorite (referred to in that mineral as "labradorescense"). Such "play of colors" can be seen in other minerals and may result from various causes. Numerous closely spaced incipient internal cracks (cleavages) or included foreign materials can cause light to be multiply reflected and refracted inside the mineral and thus enable the colors to "play."

Specific Gravity

As you already know from your work in the first week's lab session, the property of specific gravity is a number expressing the ratio between the weight of mineral compared to the weight of an equal volume of water. Practically speaking, specific gravity is a result of whether the specimen that is held in your hand feels inordinately heavy or light. Such "feeling" needs to be used with caution because one must evaluate the "heaviness" of "lightness" against the size or total mass of the specimen being held. Obviously, the larger the specimen, the heavier it will be. The key point is whether the specimen is inordinately heavy or -light with respect to its size (volume).

Magnetism

The property of magnetism refers to what is known technically as **magnetic susceptibility**. This is the ability of a mineral to attract or be attracted by a magnet. Many minerals display such magnetic attraction but not all of them contain iron as one might suppose. To make a valid magnetic test, use a small magnet from your Geology Test Kit and see if it sticks to the unknown mineral. For our purposes, the only common magnetic mineral is magnetite.

Flexibility/Elasticity

The two properties of flexibility and elasticity relate to the outcome of a simple bending test of thin plates of flaky minerals (not by flaky mineralogy students, we hope!). Only a flexible mineral can be bent. The elasticity factor refers to what happens after you bend a mineral flake and then let go of it. If the specimen remains bent, then we say that the mineral is **flexible but inelastic**. By contrast, if you can bend a mineral and after you have let go, it returns to its original position or shape, then that mineral is **not only flexible but is also "elastic."** The clear, transparent variety of gypsum known as selenite or the green mica chlorite are examples of flexible minerals that are inelastic. The light-colored mica muscovite is a flexible mineral that is elastic.

Brittleness/Tenacity

These two properties are at opposite ends of the spectrum describing how a mineral behaves when it is subjected to stress (hammer blow, applied pressure). A brittle mineral disintegrates with relative ease. A mineral that maintains itself (i. e., holds itself together) is referred to as being tenacious.

Odor

Geologists have been called "rock sniffers," probably in jest, but sometimes the old schnozz (as Jimmy Durante would have said) is a useful thing to use. Some freshly broken rocks emit such a distinctive "gassy" odor that German geologists coined the name "Stinkstein" for them. Among minerals, the earthy-musty odor of kaolinite can be enhanced by breathing on the specimen before smelling it. (A reminder: you "smell"; the mineral "stinks.") A distinctive "rotten-egg" odor can be detected from a fresh streak of some sulfide minerals. The odor arises because during the streak test, sulfur from the mineral, such as galena (a lead sulfide with formula PbS) or pyrite (iron sulfide with the formula FeS_2), is dispersed into the air.

Taste

Who would think of tasting a mineral? Among minerals that are easily soluble in water, taste can be useful. The classic example is the salty taste of halite (the mineral name of common table salt). Caution must be exercised because some minerals are toxic and certainly unpleasant to the palate.

Feel

Some minerals simply "feel" distinctive when your fingers touch them. Words useful for describing "feel" are smooth, slick or greasy. Common minerals that feel distinctive are kaolinite, talc, and graphite.

Diaphaneity

The relative ability of a mineral to allow light to pass through it is named **diaphaneity**. Three possibilities exist:

Transparent -	light passes through and one can see through the mineral
Translucent -	light passes through but one cannot see through the mineral
Opaque -	light does not pass through the mineral

So much for a discussion of the myriad of physical properties unique to minerals. Now is the time for you to apply what you have learned with some simple tests and observations on real mineral specimens. Remember, our task today is not to identify the specimens, rather, it is to become familiar with mineral properties by direct observation of mineral samples.

This week your first written assignment **[R1]** will be announced. Each lab professor will pick his/her own choice of short essay questions from the list appearing after the laboratory report sheet. Consult the instructions on writing laboratory reports in the introductory section of this laboratory manual. Feel free to consult outside sources for information to complete your reports but be sure to cite your references so as to avoid plagiarism.

LABORATORY REPORT
LAB 2 - PHYSICAL PROPERTIES OF MINERALS

#	Luster	Hardness	Cleavage	Streak / Other
1				
2				
3				
4				
5				
6				
7				
8				
9				
10				
11				
12				
13				
14				
15				
16				
17				
18				
19				
20				

Questions for Report 1 [R1] on Physical Properties of Minerals and Mineral Identification.

1. Explain at least four physical characteristics a substance must have to be considered a **mineral**.

2. Explain the distinction between a **crystal lattice** and a **crystal**.

3. What are the fundamental building blocks of crystal lattices and how are they bonded together?

4. What is the basic atomic building block of silicate minerals? What are the major structural groups of silicate minerals? List the important rock-forming silicate minerals in two adjacent columns: felsic- and mafic minerals. Alongside each mineral indicate to which structural group it belongs.

5. Define **crystal face** and a **cleavage plane**. How can you distinguish the one from the other?

6. Discuss the diagnostic physical properties that you can use to distinquish between the similar-looking ferromagnesian minerals olivine, pyroxene, and amphibole?

7. Biotite mica and muscovite mica possess similar crystal lattices.
    a. What makes them different minerals?
    b. What physical properties can you use to distinguish biotite from muscovite?

8. You are given a specimen of orthoclase feldspar (K-feldspar) and one of light-colored plagioclase feldspar (albite).
    a. Discuss four physical properties that each of these minerals have in common.
    b. Discuss one physical property that you can use to distinguish the potassium feldspar (K-feldspar) from the plagioclase (albite).

9. You are given three minerals: dark-colored plagioclase feldspar (labradorite), pyroxene (augite), and amphibole (hornblende).
    a. Discuss four physical properties that each of these minerals have in common.
    b. What property can you use to distinguish the amphibole from the pyroxene and the plagioclase?
    c. What property can you use to distinguish pyroxene from plagioclase?

10. Describe three tests you can use to distinguish halite from colorless calcite.

11. Besides cleavage, hardness, and luster, discuss five additional physical properties that can be useful for identifying minerals.

12. What products or important raw materials are made from these minerals? Write out your answers in complete sentences.
    a. muscovite
    b. gypsum
    c. talc
    d. graphite
    e. galena
    f. feldspar
    g. garnet
    h. hematite, limonite, magnetite

Lab 3 - Mineral Identification

What to do:

Having learned the secrets of mineral physical properties last week, in today's laboratory exercises you will identify the twenty mineral specimens found in the large brown mineral collection boxes. Kindly review the concepts from last week, pull out your Geology Kits, and in the best spirit of **Sherlock Holmes** try to logically and systematically identify the twenty minerals. Use the **mineral tests** you learned last week to make observations of luster, hardness, streak, cleavage, etc., that will lead you to the correct identifications. Record your observations on the Lab 3 Data Sheet. Provided on the following pages are a series of tables and charts of mineral properties to help you. Particularly useful are Tables 3.1 and 3.2 which allow you to key out minerals starting from general- and working to more-specific properties. Remember, next week you will be tested on your ability to identify these twenty common minerals.

Table 3.1 - Mineral identification flow chart for non-metallic, light-colored minerals.

Table 3.2 - Mineral identification flow chart for non-metallic, dark-colored minerals.

Table 3.3 - Mineral Tables.

Table 3.3 - Mineral Tables.

Table 3.3 - Mineral Tables.

Table 3.4 (two pages) – Mineral Properties Chart

Minerals Arranged by Hardness

soft Scratched by fingernail **<2.5**	medium **2.5 – 5.5**		hard Scratches glass **>5.5**	
Talc	Calcite	Galena	Quartz	Augite *
Gypsum	Fluorite	Halite	Olivine	Hornblende *
Graphite	Dolomite	Biotite	Magnetite	Hematite *
Kaolinite	Hematite	Sphalerite	Pyrite	Albite
Chlorite	Apatite	Limonite	Garnet	Labradorite
Muscovite				K–feldspar

* indicates that some varieties may be softer than glass.

Minerals Arranged by Luster

Metallic

- Galena
- Graphite
- Hematite (metallic)
- Magnetite
- Pyrite
- Sphalerite #

Nonmetallic

Vitreous

- Quartz
- K-feldspar #
- Garnet
- Calcite #
- Gypsum #
- Olivine
- Halite
- Fluorite
- Augite #
- Hornblende

Pearly

- Labradorite #
- Albite #
- K-feldspar #
- Calcite #
- Dolomite #
- Chlorite
- Talc
- Muscovite

Earthy

- Hematite (earthy)
- Limonite #
- Kaolinite

Shiny

- Limonite #
- Biotite

Albite # Labradorite # Apatite

# indicates that some varieties can have a different luster.

Minerals Arranged by Streak

(Always use the streak test for minerals having a metallic, shiny, or earthy luster.)

Gray to Black	Reddish–Brown	Yellowish–Brown	Yellow
Galena Magnetite Pyrite	Hematite	Limonite	Sphalerite (when streak is fresh it gives off an odor of sulfur)

Table 3.4 (cont'd) – Mineral Properties Chart

Minerals Arranged by Method of Breaking

Fracture (no cleavage)

- Quartz
- Garnet
- Olivine
- Pyrite
- Kaolinite
- Hematite
- Limonite
- Apatite*

*apatite has one cleavage direction but it is not usually visible.

Cleavage

One (well expressed)

- Muscovite
- Biotite
- Chlorite

One (poorly expressed)

- Talc
- Graphite

Two (right angles)

- Albite
- Labradorite
- K–feldspar
- Augite

Two (not at right angles)

- Hornblende
- Gypsum *

* gypsum's cleavage is often poorly expressed.

Three or more

- Galena
- Halite
- Calcite
- Dolomite
- Fluorite
- Sphalerite

Minerals Arranged by Special Features

Striae

On cleavage faces

- Albite
- Labradorite

On crystal faces

- Pyrite
- Hornblende

Taste

Halite (salty)

Smell

Kaolinite (smells like clay when wet)

Magnetism

Magnetite

Reaction to Dilute Acid

Calcite (effervesces readily)
Dolomite (effervesces only weakly when first scratched or powdered)

High Specific Gravity

(minerals that "feel heavy")

- Galena
- Pyrite
- Magnetite
- Sphalerite
- Hematite

Laboratory Data Sheet 4.1

Mineral Identification

#	Luster / Color	Hardness	Cleavage	Streak	Other	Mineral Name
1						
2						
3						
4						
5						
6						
7						
8						
9						
10						

Laboratory Data Sheet 4.1
Mineral Identification

#	Luster / Color	Hardness	Cleavage	Streak	Other	Mineral Name
11						
12						
13						
14						
15						
16						
17						
18						
19						
20						

Lab 4 - Mineral Practicum and Introduction to the Three Rock Types

PURPOSE

The purpose of today's laboratory meeting is to test your proficiency in identifying unknown mineral specimens as provided by your lab professor. You will have ample time to use your test kits, record data, and name your mineral buddies. A sample practicum report sheet is provided, in reduced form, below. After reviewing your practicum, your lab professor will begin discussing the identification of rocks. To assist you in this task, starting on the next page, we have prepared two sections entitled "**INTRODUCTION TO ROCKS** and **HINTS FOR IDENTIFYING ROCKS IN HAND SPECIMENS**". Be sure to read these before you enter lab next week.

Geology Mineral Practicum Name: ________

Sample	Luster	Hardness	Cleavage (yes / no) type?	Color, streak or other	Mineral Name
1					
2					
3					
4					
5					
6					
7					
8					
9					
10					
11					
12					

Introduction to the Three Rock Types

You have already learned that geologists classify rocks into three major groups: (1) igneous; (2) sedimentary; and (3) metamorphic. But once you have learned the names of these three rock groups, so what? Why have geologists organized rocks into these three groups and on what basis did they do so? If a rock doesn't come with a label on it, how can one look at it and decide to which group it belongs? This division of rocks into three categories is of fundamental importance to the science of geology and we will spend the next four laboratories learning to recognize and interpret the different kinds of rocks.

In classifying rocks, geologists use two major factors: (1) **process(es)** and (2) **place(s)**, meaning at the Earth's surface or inside the Earth. As a first approximation, within each major rock group, one of these factors prevails throughout, and thus becomes a **unity factor**. The other factor typically is variable, and thus becomes a **diversity factor**. Among igneous rocks, process (cooling from a molten state) is the unity factor and place (where cooling takes place) is the diversity factor. Among sedimentary rocks, it is necessary to consider these two factors in two stages: (a) stage 1 is the the formation of the sediments, and (b) stage 2 is the conversion of the sediments into sedimentary rocks. Among the sediments of stage 1, place (the surface of the Earth) is the unity factor and process is the diversity factor. By contrast, in stage 2, the conversion of sediments into sedimentary rocks, place is once again the unity factor (but not the same place as with sediments; the place where lithification happens typically is inside the Earth rather than at the Earth's surface) and various processes become the diversity factor.

Among most metamorphic rocks, place (inside the Earth) is the sole unity factor, but two diversity factors are involved: (a) pre-existing rock (known as the **parent rock** or **protolith**) and (b) various processes. (Impact metamorphism associated with the arrival of a meteorite is an exception to place as a unity factor in metamorphic rocks; impact metamorphism happens only at the Earth's surface.) In the following paragraphs, we summarize how these factors are expressed in rock classification.

Igneous Rocks: One Process Operates in Two Places

All igneous rocks result from one process: the cooling and solidification of a molten liquid. But geologists have established two major categories of igneous rocks depending on the place (or location) where the cooling took place. **Extrusive igneous rocks** form at the Earth's surface by the cooling and solidification of a molten liquid known as **lava**. By contrast, **intrusive igneous rocks** form within the Earth by the cooling and solidification of a mixture of molten liquid + dissolved gases known as **magma**. Three important reasons can be cited for explaining why geologists make a major distinction among igneous rocks based on the place where the cooling happens.

(1) Beneath the Earth's surface, the pressure enables the gases to remain dissolved within the molten liquid (magma). As soon as the magma reaches the Earth's surface where it can be seen, geologists change its name from magma to lava. But,

something else happens, too. The dissolved gases escape. In its simplest form, then, we can consider lava as being degassed magma.

(2) Place generally affects the **rate of cooling**. As we shall see in detail when we begin studying igneous rocks (Lab 6), rate of cooling determines the sizes of the growing crystals. Therefore, any factor that affects the rate of cooling has to be considered a first-order factor in understanding igneous rocks. As a general rule, cooling at the Earth's surface is a fast process (meaning days or weeks), whereas cooling that takes place inside the Earth is a slow process (meaning thousands or millions of years depending on volume of magma, depth, and other factors).

(3) Place determines how the heat escaping from the molten liquid will affect its surroundings. A body of lava cooling at the Earth's surface heats only the materials over which it flows. A body of body of magma cooling underground heats its surroundings in all directions: top, bottom, and sides.

Sediments and Sedimentary Rocks: Many Processes Operate in One Place

Sedimentary rocks result from the lithification of sediments. To understand sedimentary rocks, then, we must unravel two stages: in the first stage, we must find out about sediments and their potential source areas; in the second stage, we study how the sediments become lithified to form sedimentary rocks. Here, we discuss the first stage of sediments at greater length than the second stage, the processes of converting sediments to sedimentary rocks.

The unity factor among sediments is place: they all form at the Earth's surface, albeit in vastly different environments. The diversity factor is process: many processes are involved in the origin of sediments. In the lithification of sediments, the unity factor is generally place (but not the same place as with sediments, but *within* the Earth; however this place is not a hard-and-fast generalization because some sediments are lithified at the Earth's surface); many processes constitute the diversity factor.

Metamorphic Rocks: Unity of Place but Diversity of Protoliths and Processes

In most metamorphic rocks, the unity factor is place (within the Earth); but not one but two diversity factors must be considered: (a) pre-existing rock (the **parent rock** or **protolith**) and (b) metamorphic process(es). Some major metamorphic processes include heat acting alone, deformation by pressure and shearing, and deformation plus heat acting in combination. Rearrangement of minerals during deformation gives rise to distinct **rock fabrics** in which silicate minerals such as micas and amphiboles become preferentially oriented.

So much for a general introduction to the classification of rocks into the three major groups. Remember that rocks are composed of aggregates of one or more minerals with which you are already familiar. But in studying rocks, the individual crystal lattices are small and typically show compromise boundaries. This means you will have to learn to make tests for diagnostic mineral properties on tiny constituents. Despite this size limitation, the same key properties by which you have learned to identify large individual mineral specimens are useful. As described in Lab 8 (take some time to look ahead), these include hardness, luster, color, method of breaking [fracture vs. cleavage(s)], striae, and reaction with dilute hydrochloric acid.

Lab 5 - Igneous Rocks

PURPOSE

Following on the heels of last week's introduction to rocks, the purpose of today's laboratory session is to acquaint you with some common igneous rocks. Today, we will start by assigning few interesting exercises (Exercises 5.1 through 5.3) to be completed at home. Next we'll examine the large display specimens on the side bench of the lab which will exemplify distinctive features of igneous rocks. Finally, we'll examine and identify unknown igneous rock samples. Be sure to read the discussion of Igneous-rock Classification, Rock-Forming Minerals, and Bowen's Reaction Series below and in your textbook to understand better the constituents and genesis of igneous rocks.

Exercise 5.1

A representative listing of 100 sites of recent volcanism is provided in Table 5.1 (modified from MacDonald, 1972). Using the world map sheet (Figure 5.1) and the latitude/longitude data provided in Table 5.1, plot the positions of the 100 centers of volcanicity. Round off the location data to the nearest degree and plot them, at home, on the world map using small circles a few millimeters in diameter. After you have finished this map-data exercise, evaluate the data and be prepared in next week's lab, to comment on the following:

Questions:

5.1) Are volcanoes distributed in any pattern on the Earth or do they occur randomly?

5.2) Since 1700 A.D., what region on Earth has experienced the greatest volcanic activity?

5.3) Are the continental volcanoes located in the interiors or on the edges of the continents? Why?

Table 5.1 - Geographic Distribution of Volcanic Activity Since 1700 A.D. (modified from MacDonald, G. A., 1972, p. 429-450).

Number	**Name**	**Lat.**	**Long.**	**Number of Eruptions**
1	Barren I.	12˚15'N	93˚50'E	4
2	Peuetsagoe	4˚55'N	96˚20'E	1
3	Marapi	0˚23'N	100˚28'E	36
4	Sumbing	2˚25'S	101˚44'E	2
5	Krakatoa	6˚06'S	105˚25'E	2
6	Galunggung	7˚15'S	108˚03'E	3
7	Slamet	7˚14'S	109˚12'E	19
8	Kelut	7˚56'S	112˚18'E	18
9	Lamongan	7˚59'S	113˚20'E	37
10	Batur	8˚14'S	115˚22'E	15
11	Tambora	8˚15'S	118˚00'E	3
12	Amburombu	8˚48'S	121˚11'E	5
13	Keli Mutu	8˚45'S	121˚50'E	2
14	Lewotobi Perampuan	8˚34'S	122˚47'E	2
15	Sirung	8˚30'S	124˚09'E	5
16	Nieuwerkerk	6˚36'S	124˚43'E	3
17	Serua	6˚18'S	130˚00'E	7
18	Soputan	1˚06'N	124˚43'E	18
19	Dokono	1˚42'N	127˚52'E	3
20	Bam	3˚36'S	144˚51'E	10
21	Victory	9˚11'S	149˚04'E	2
22	Pago	5˚35'S	150˚31'E	1
23	Gaua	14˚15'S	167˚30'E	2
24	Yasour	19˚31'S	169˚25'E	2
25	Matavanu	13˚31'S	172˚22'W	1
26	Home Reef	18˚59'S	174˚46'W	1
27	(Unnamed)	20˚51'S	175˚32'W	2
28	Tarawera	38˚16'S	176˚34'E	1
29	Ruapehu	39˚16'S	175˚35'E	13
30	Ragang	7˚40'N	124˚31'E	9
31	Bulusan	12˚46'N	124˚03'E	10
32	Smith	19˚32'N	121˚56'E	8
33	(Unnamed)	26˚11'N	122˚27'E	1
34	Tokara-Iwo-Zima	30˚47'N	130˚17'E	2
35	South Pagan	18˚06'N	145˚47'E	1
36	Uracas	20˚38'N	144˚54'E	10
37	(Unnamed)	26˚00'N	140˚46'E	0
38	Hatizyo-sima Nishi-yama	33˚08'N	139˚46'E	1
39	O-sima	34˚44'N	139˚23'E	18

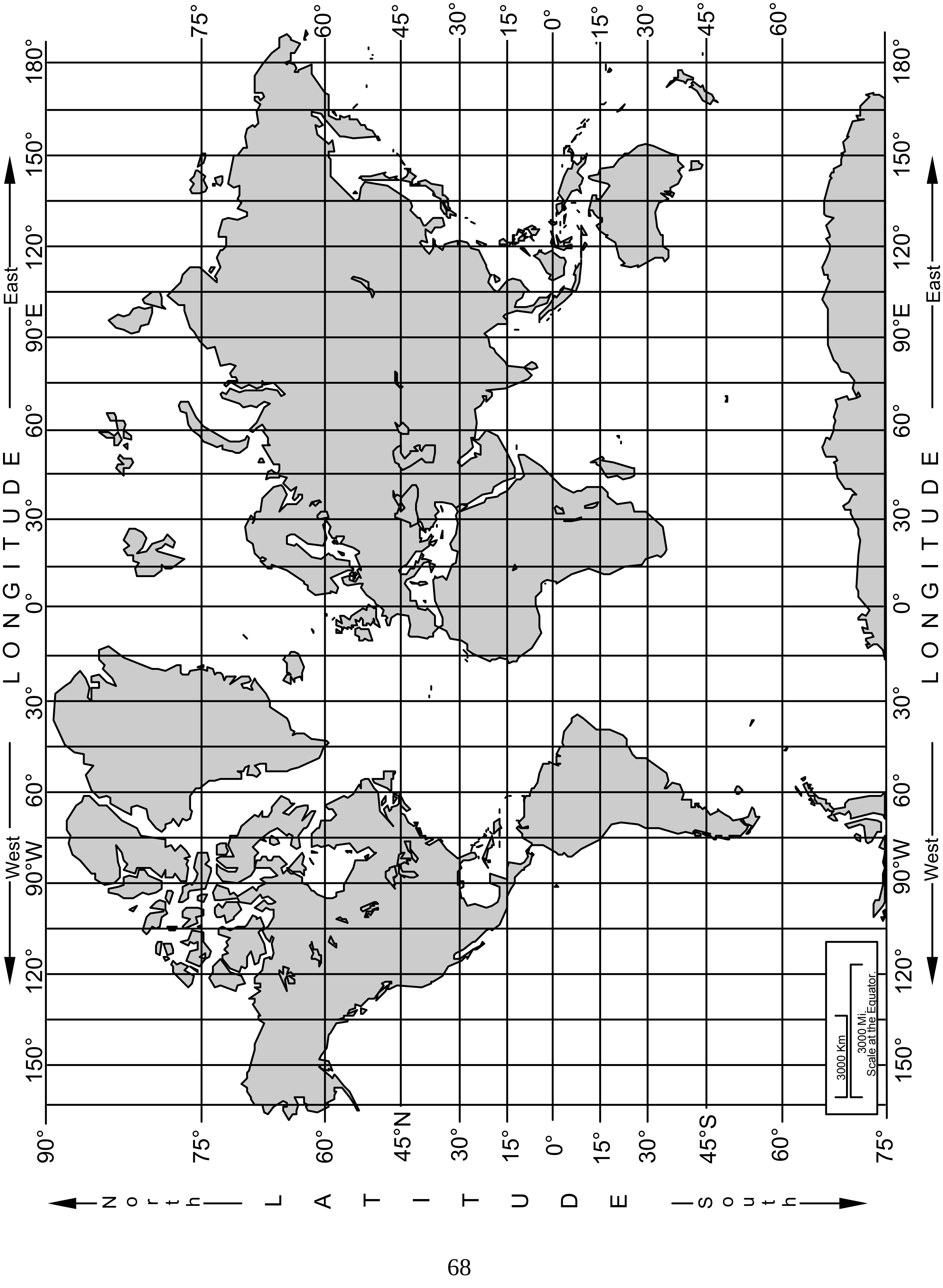
Table 5.1 (cont'd)

Number	Name	Lat.	Long.	Number of Eruptions
40	Kirishima	31˚53'N	130˚55'E	19
41	Hakusan	36˚09'N	136˚47'E	0
42	Nikko-sirane	36˚48'N	139˚23'E	4
43	Iwate	39˚51'N	141˚00'E	3
44	Tarumai	42˚41'N	141˚23'E	11
45	Mendeleev	43˚59'N	145˚42'E	2
46	Snow	46˚31'N	150˚52'E	4
47	Sarychev	48˚05'N	153˚12'E	8
48	Nemo	49˚34'N	154˚48'E	2
49	Ksudach	51˚49'N	157˚32'E	1
50	Koriaksky	53˚19'N	158˚41'E	2
51	Plosky Tolbachik	55˚49'N	160˚22'E	18
52	Anjuisky	67˚10'N	165˚23'E	0
53	Kiska	52˚06'N	177˚36'E	5
54	Gareloi	51˚48'N	178˚48'W	8
55	Sarichef	52˚19'N	174˚03'W	1
56	Okmok	53˚25'N	168˚03'W	10
57	Chiginagak	57˚08'N	157˚00'W	2
58	Mt. Rainier	46˚52'N	121˚45'W	6
59	Lassen Peak	40˚29'N	121˚30'W	4
60	Paricutin	19˚29'N	102˚15'W	1
61	Izalco	13˚49'N	89˚39'W	53
62	San Miguel	13˚26'N	88˚16'W	24
63	Las Pilas	12˚29'N	86˚41'W	2
64	Arenal	10˚29'N	84˚43'W	2
65	Tolima	4˚39'N	75˚22'W	4
66	Cumbal	0˚59'N	77˚53'W	2
67	Cotopaxi	0˚50'S	78˚26'W	50
68	Isluga	19˚09'S	68˚50'W	5
69	Peteroa	35˚15'S	70˚34'W	7
70	Los Copahues	37˚51'S	71˚10'W	?
71	Huequi	42˚20'S	72˚40'W	5
72	(Unnamed)	56˚15'S	72˚10'W	1
73	(Unnamed)	55˚54'S	27˚54'W	1
74	Mt. Darnley	59˚03'S	26˚30'W	5
75	Erebus	77˚35'S	167˚10'E	7
76	Buckle I.	66˚48'S	163˚15'E	2
77	Sierra Negra	0˚50'S	91˚10'W	3
78	Kilauea	19˚25'N	155˚17'W	47
79	(Unnamed)	23˚35'N	163˚50'W	1
80	Macdonald	29˚01'S	140˚17'W	1
81	Mt. Misery	17˚22'N	62˚48'W	2
82	Mt. Pelee'	14˚49'N	61˚10'W	4

Table 5.1 (cont'd)

Number	Name	Lat.	Long.	Number of Eruptions
83	(Unnamed)	49˚00'N	34˚30'W	1
84	San Jorge I.	38˚39'N	28˚05'W	2
85	Sete Cidades	37˚52'N	25˚47'W	4
86	Tenerife I.	28˚18'N	16˚38'W	5
87	(Unnamed)	7˚00'N	21˚50'W	1
88	(Unnamed)	3˚30'S	24˚30'W	1
89	Meitill	64˚00'N	21˚24'W	0
90	Eldgja	63˚58'N	18˚36'W	0
91	Krafla	65˚42'N	16˚44'W	1
92	Beerenberg	71˚05'N	7˚50'W	2
93	Vesuvius	40˚49'N	14˚26'E	Many
94	Etna	37˚44'N	15˚00'E	75
95	Santorin (Thera)	36˚24'N	25˚24'E	6
96	Dubbi	13˚42'N	41˚35'E	2
97	Longonot	0˚55'S	36˚27'E	1
98	Nyiragongo	1˚29'S	29˚14'E	13
99	Karthala	11˚45'S	41˚03'E	21
100	St. Paul Island	38˚43'S	77˚31'E	1

Figure 5.1 - Base map for plotting the volcanoes listed in Table 5.1 (above) and as explained in Exercise 5.1.



Exercise 5.2

A block diagram illustrating the various forms of bodies of igneous rock (igneous features) is provided in Figure 5.2 for this exercise. Using your textbook and other sources, define the igneous features listed and match them to the numbered features on the block diagram (Nos. 1 through 13). Fill in the table below indicating on your list with a "V" or a "P" where volcanic textures or plutonic textures would occur and, if you can, what typical igneous rocks might be found at each locality. Be prepared to discuss your work in lab next week.

#	Igneous Feature	Volcanic - Plutonic	Rock Types
1			
2			
3			
4			
5			
6			
7			
8			
9			
10			
11			
12			

Dike
Sill
Caldera
Laccolith
Laccolith exposed by erosion
Batholith
Radial dike swarm
Volcanic neck (pipe)
Volcanic vent
Lava flow
Xenoliths (inclusions)
Flood basalt

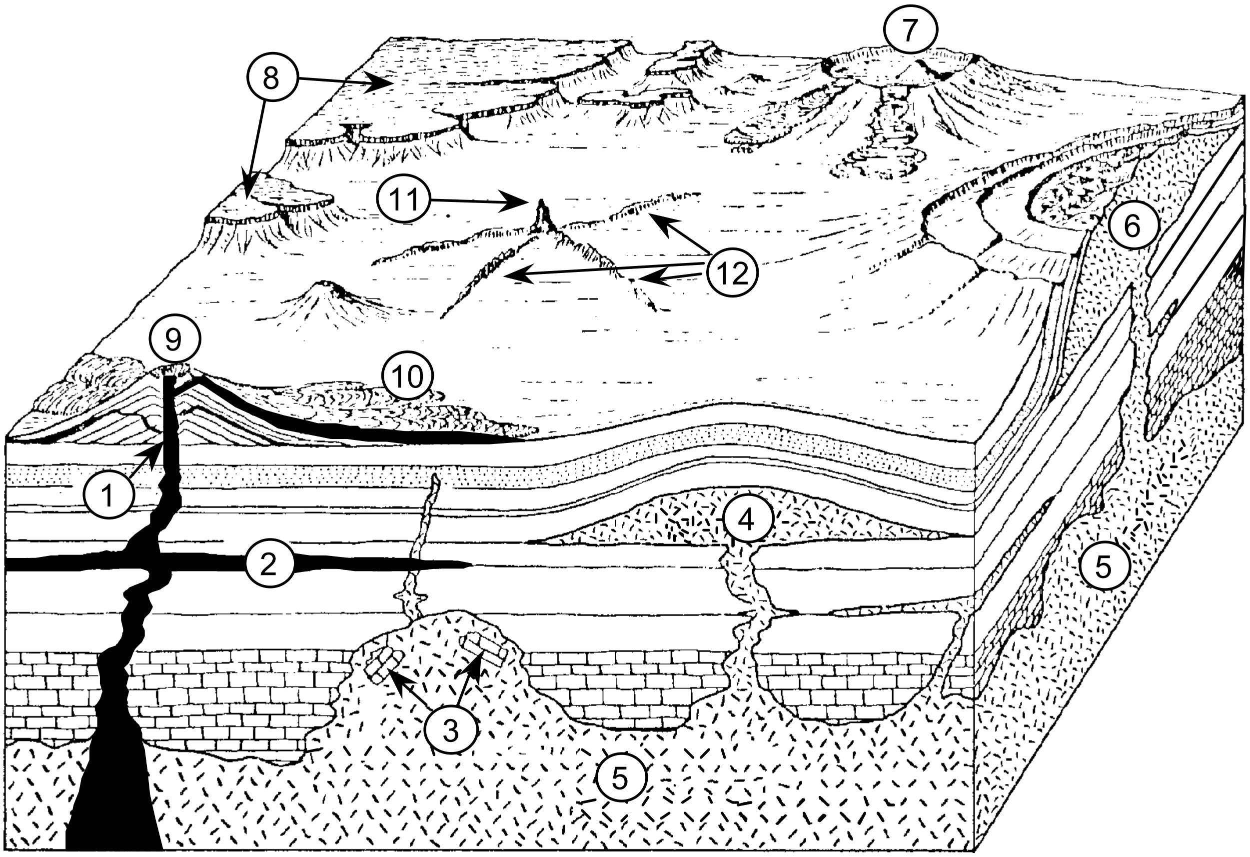


Figure 5.2 - Block diagram illustrating the common occurrences of volcanic- and plutonic rocks. See instructions for Exercise 5.2.

Exercise 5.3

Discuss the following questions in lab:

5.4) What geologic events are implied if a body of granite is found exposed at the Earth's surface?

5.5) A granite may be pink, red, or various shades of gray and may be coarse-, medium-, or fine textured. What characteristics are found in all granites which serve to distinguish them from other rock types?

5.6) What is the origin of vesicles in a basalt?

5.7) What rock type is most abundant in the crust of the ocean floors? What is the origin of magma in these areas?

5.8) What is the difference between granite and diorite?

5.9) What is the difference between basalt and andesite?

CLASSIFICATION OF IGNEOUS ROCKS

Recall from last week's introduction to the three major rock groups that all igneous rocks form by the cooling and solidification of a hot liquid consisting of molten rock and gases. **Volcanic** (or **extrusive**) igneous rocks form by the solidification of **lava** at the Earth's surface. **Plutonic** (or **intrusive**) igneous rocks form from solidification of **magma** within the Earth.

Two fundamental factors affect the development of igneous rocks: (1) the cooling history, and (2) the chemical composition of the melt. Cooling history determines the **texture** of an igneous rock, and the chemical composition determines the kind(s) of mineral constituents. Chemically, igneous rocks are divided into the light-colored **felsic group** (dominated by K-feldspar, albite, and quartz), a dark-colored **mafic group** (dominated by the ferromagnesian silicate minerals olivine, pyroxenes, amphiboles, and biotite plus labradorite), and a group of intermediate composition.

In the following paragraphs, we explain these two factors and show how they are used to classify and name hand specimens of igneous rocks. Using these two features of texture and mineral composition, you can classify and name common igneous rocks (all rocks, for that matter!). However, small specimens alone may not always provide enough information to determine unambiguously whether the igneous rock is intrusive or extrusive.

Cooling History of Igneous Rocks:
Fast or Slow? Uniform or Nonuniform?
Textures of Igneous Rocks

The fundamental principle governing all solidificaiton of solids from a melt is that **the cooling history determines the outcome**. Cooling may be fast or slow, uniform or nonuniform. Whatever it was, the cooling history of an igneous rock governs the sizes, uniformity of sizes, and mutual arrangements of the mineral crystals which collectively determine the **texture** of the rock. Texture is one of the fundamental properties that geologists use to classify all rocks. Starting right now, pay attention to the texture of any rock specimen you study.

Fast- vs. Slow Cooling

As noted above, one of the important parts of the cooling history of an igneous rock is the rate of cooling. Was cooling fast (days to weeks) or slow (thousands to millions of years)? During slow cooling, scattered large crystal lattices can grow. During rapid cooling, many small lattices can grow. During extreme chilling or quenching, no lattices at all form; the result is a noncrystalline solid known as a **natural glass**. This principle that the **rate of cooling determines the outcome** is valid no matter where the cooling takes place.

Geologists use the sizes and uniformity of sizes of the mineral crystals in an igneous rock as first-order factors in classification. We begin with examples where the crystals are all about the same size. Igneous rocks having such uniform textures are designated as **equigranular igneous rocks**.

Among equigranular igneous rocks, four textural categories based on the size of mineral crystals are:

1 - **glassy** (no crystal lattices and occur in massive or frothy [porous] varieties depending upon chemical- and environmental conditions).

2 - **aphanitic** (crystal lattices present, but they are not large enough to be identified using only a hand lens or even a binocular-stereomicroscope, thus their lengths are less than 0.05 mm).
3 - **phaneritic** (individual mineral crystals large enough to be clearly visible, even without using a hand lens or a stereomicroscope, but less than 10 mm in length).

Three grain-size classes are:
coarse - 10 to 5 mm
medium - 5 to 1 mm
fine - 1 to 0.05 mm

4 - **pegmatitic** (individual mineral crystals larger than 10 mm; a kind of texture that is not easily shown by small specimens; refer to the large display specimens).

Thus, the genetic relationship between crystal size and origin of igneous rocks cooling under ideal equilibrium conditions is quite simple. Slowly cooled plutonic rocks exhibit pegmatitic- or phaneritic textures, whereas rapidly cooled volcanic rocks exhibit aphanitic- or glassy textures. Thus, a simple crystal-size measurement illuminates much about the history of that rock!

Uniform- vs. Nonuniform Cooling

In addition to the rate, a second aspect of the cooling history affects the texture of igneous rocks. This second aspect is whether the cooling was uniform or nonuniform. If the cooling history was uniform, then the texture of the igneous rock will be uniform. That is, the crystal lattices in the rock will all be of about the same size (the texture will be equigranular, as defined above). If the cooling history was not uniform, then the texture of the igneous rock will display nonuniform textures (somemineral crystals distinctly larger than the others). Such nonuniform textures in igneous rocks are named **porphyritic textures**. In an igneous rock having a porphyritic texture, the distinctly larger crystals are named **phenocrysts** and the smaller crystals are collectively designated as the **groundmass**. The key point is the distinctly contrasting sizes. The phenocrysts can be any size; the texture of the groundmass is usually aphanitic, but may be phaneritic or even glassy.

In naming igneous rocks having porphyritic texture, it is necessary to estimate the proportion of the phenocrysts. Refer to Figure 8.1 for graphic assistance in estimating the abundances of the phenocrysts (or other constituents). If the phenocrysts constitute more than 25% of the volume of the rock, then the rock is named a **porphyry**. If the phenocrysts constitute less than 25% of the volume of the rock, then whatever rock name is determined by the minerals is modified by inserting ahead of the rock name the adjective **porphyritic**.

Special Features of Volcanic Rocks Formed During Cooling

Cooling not only controls the textures just described, but while cooling is in progress, gases dissolved in the lava may be rapidly escaping. The rapid escape of gases from lava may create a spongy-looking rock full of small more-or-less spherical cavities named **vesicles**. The resulting igneous rock is referred to as a **vesicular igneous rock**. Continuous trails of vesicles are collectively called **pipe vesicles**. Many vesicles remain as empty spaces but others become filled with minerals to form tiny cavity-filling mineral deposits named **amygdales**.

Vesicular igneous rocks. A frothy felsic rock having tiny (<1 mm) cavities is **pumice**. A frothy mafic rock having large (>l mm) cavities is **scoria** or if the vesicles are not abundant enough to qualify the rocks as being "frothy," the term **vesicular** may be appropriate.

Amygdaloidal igneous rocks. If vesicles have been filled by minerals of later origin (such as calcite, quartz, chlorite, and zeolite minerals), the result is amygdales. An igneous rock containing amygdales is described as being **amygdaloidal**. Can you distinguish amygdales from phenocrysts? (Hint: think about shapes.)

ROCK-FORMING MINERALS AND BOWEN'S REACTION SERIES

Silicate minerals are the keys to understanding igneous rocks, which form as the result of changes that take place in molten liquids (named **magma** if underground with gases held in, or **lava** if on the Earth's surface where the gases can separate).

Our understanding of the systematic behavior of various silicate minerals in magmas derives largely from the work of N. L. Bowen, who used empirical data and a series of experiments to understand how growing, solid silicate minerals interact with molten silicate melts. Bowen found two contrasting mineralogical behaviors: (a) *continuous reaction*, and (b) *discontinuous reaction*.

In a **continuous reaction**, the growing lattices interact with the melt by exchanging ions. Within a given temperature range, lattices of a mineral persist but as the temperature drops, ions are continuously interchanged between the melt and the already formed mineral lattices. This interchange changes the composition of the lattices. For example, the predominant ion in high-temperature olivine is magnesium (Mg^{2+}). But, the predominant ion in lower-temperature olivine is ferrous iron (Fe^{2+}). Therefore, as long as the temperature remains within the range of olivine stability, olivine lattices undergo continuous reaction with the melt: magnesium ions leave the olivine lattices and go back into the melt and iron ions from the melt enter the ever-changing olivine lattices. Notice that the electrical charges of magnesium and ferrous iron ions is the same, 2+. Therefore, an exchange between these two ions does not alter the electrical balance and any composition between the Mg^{2+}-rich and Fe^{2+}-rich olivine end members is possible. Mineralogists refer to this process as **solid solution**.

In high-temperature plagioclase (feldspar), calcium (Ca^{2+}) is the dominant ion. As temperature drops, calcium ions leave the plagioclase lattices and re-enter the melt: sodium ions (Na^{+}) from the melt enter the plagioclase lattices. Notice that every exchange involves a loss of one positive charge, from +2 of the calcium to +1 of the sodium. In order to keep the feldspar lattice electrically balanced, each exchange between a Ca^{2+} from a plagioclase lattice position and an Na^{+} from the melt requires a corresponding substitution in which either the positive electrical charge drops by one or the overall negative charge is also reduced by 1. The first option, that of reducing the overall positive charge by 1 and not changing the negative ions, can be fulfilled if an Al^{3+} ion from the melt substitutes for an Si^{4+} ion in one of the tetrahedra. Such a change that takes place simultaneously between the ions occupying positions outside the silica tetrahedra (the Ca^{2+} and the Na^{+}) and those within the tetrahedra (the Al^{3+} and the Si^{4+}) is known as a **coupled substitution**.

In a **discontinuous reaction**, the lattices of the first mineral, which grow and are stable only while the cooling melt remains within a given temperature range, exhibit a different behavior if the temperature of the melt drops below that range. When such a temperature change takes place, the lattices of the first mineral break down and lattices of a different mineral start to grow. For example, as the temperature drops below the olivine range, olivine crystals dissolve back into the melt and pyroxene lattices form.

The crystallization of the dark-colored silicate minerals rich in iron and magnesium (the *ferromagnesian silicate minerals*: olivine, pyroxenes, hornblendes, and biotite), proceeds by alternating continuous- and discontinuous steps during cooling. The initial mineral olivine exchanges its Mg^{2+} for Fe^{2+} for a time. Eventually, this continuous reaction stops and a discontinuous step begins--the olivine crystals are dissolved and crystals of pyroxene form. Pyroxene exchanges ions with the melt (magnesium at higher temperatures, iron at lower temperatures). When the temperature drops below the lower limit for pyroxene, then another discontinuous reaction begins: the pyroxene crystals are dissolved and those of amphibole (hornblende) appear. The crystals of hornblende react continuously with the melt (the usual magnesium at the high end of the temperature range and iron at the lower end) but eventually the temperature drops below the lower limit of amphibole stability. Now, yet-another discontinuous reaction begins: the hornblende lattices are dissolved and lattices of biotite start to grow. Notice that the progression of the arrangement of the silica tetrahedra in the minerals that form as the temperature drops from higher to lower is in the order: isolated tetrahedra (olivine), single chains (pyroxene), double chains (amphibole), and sheets (biotite).

The contrasting group of light-colored silicate minerals is dominated by feldspars. Among the feldspars, plagioclase exhibits a continuous-reaction relationship by substitution of Na^{+} for Ca^{2+} with the cooling melt, as explained above.

A **solid solution** (gradational ionic substitution) is not unusual in the silicate minerals. In fact, it is the normal situation. Variability in the chemical composition of silicate minerals is a normal consequence of these reactions. Minerologists use these chemical variations as fingerprints which yield clues as to the temperature and pressure (depth in the Earth) of crystallization of igneous rocks. In this way, minerals provide information that allows geologists to reconstruct the geologic history of a region.

Bowen's Reaction Series is illustrated graphically below in Figure 6.3.

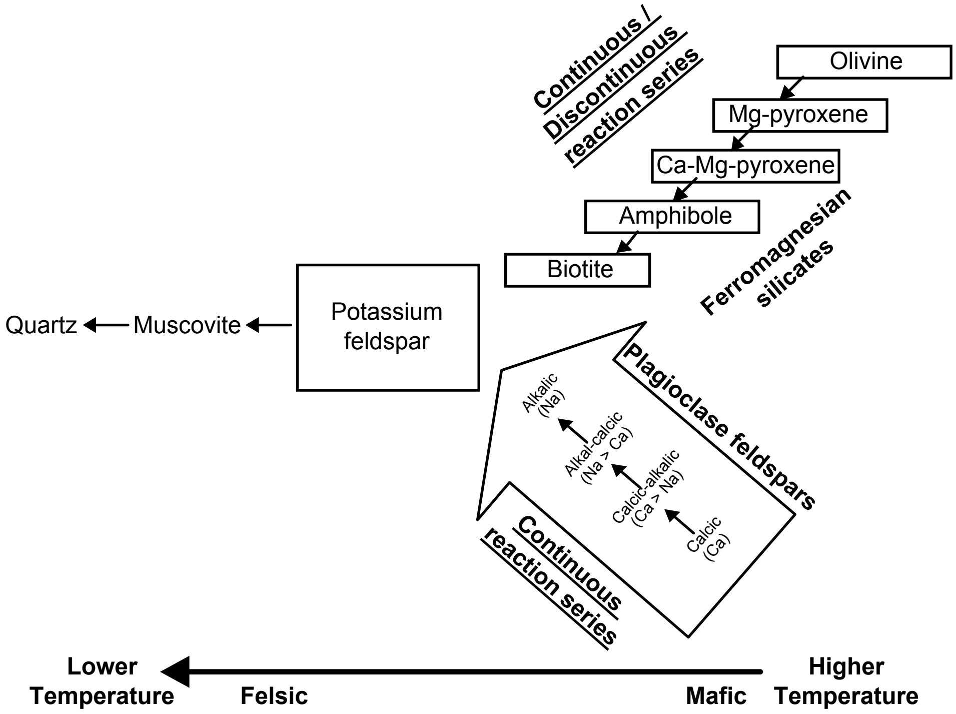


Figure 5.3 - Bowen's Reaction Series.

IDENTIFICATION OF IGNEOUS ROCKS

Now that the preliminaries are out of the way, it is time to think about the process of identifying igneous rocks. Igneous rocks are classified on the basis of composition and texture.

Start by checking the color of the rock. Light-colored igneous rocks generally are **felsic** (contain abundant potassium feldspar, albite, and quartz). Dark-colored igneous rocks generally are **mafic** (contain abundant pyroxene or other ferromagnesian minerals, such as amphibole, biotite, or olivine and plagioclase). Here are some points about individual minerals.

Minerals of Igneous Rocks

Quartz: Any present? if so, how much. No igneous rock contains more than about 35 per cent quartz. Some igneous rocks (**syenites**) contain less than 5% quartz. Quartz tends to be clear to gray in igneous rocks. Commonly quartz looks like broken-up glass - vitreous - in igneous rocks, but without distinct cleavage surfaces.

Feldspars: These are the key minerals for igneous rocks. In the felsic rocks, K-feldspar and albite predominate; in mafic rocks, plagioclase dominates. Some igneous rocks (**anorthosites**) contain nearly 100% feldspar(s). Although some feldspars are light in color and can look like quartz in igneous rocks, they are distinguished by the presence of reflective cleavage surfaces.

> ***K-feldspar:*** low-temperature variety including microcline (greenish or pinkish), orthoclase (cream-colored to pinkish), and sanidine (clear to whitish; restricted to volcanic rocks). Stubby crystals are typical.
>
> ***Albite:*** a variety of plagioclase containing mostly sodium that occurs with low-temperature K-feldspars; the color generally milky white (but, being a plagioclase, can sometimes display twinning striae).
>
> ***Plagioclase (other than albite):*** the calcium-rich varieties of plagioclase predominate among diorites, basalts, dolerites, gabbros, and other mafic- and ultramafic rocks. The common high-calcium plagioclase is ***Labradorite***, smoky-colored to deep blue with "play of colors" (iridescence). Some varieties of plagioclase are clear and glassy (i.e., their luster is vitreous as in quartz); despite being glassy, however, they display the striae, cleavage, and hardness of plagioclase. Plagioclase typically forms elongate, lath-like crystals.

Ferromagnesian silicates: the dark-colored minerals in igneous rocks (exception: magnetite). The common ferromagnesian minerals listed in order of decreasing temperature of formation are:

> ***Olivine*** - very difficult to observe in plutonic rocks but commonly forms transparent green phenocrysts in basalts; vitreous in appearance with no cleavage.
>
> ***Pyroxene*** - dark-green color, 6-sided crystals that are tabular to prismatic; two good cleavages at right angles. ***Augite*** is the most common pyroxene.
>
> ***Amphibole*** - generally black, shiny, lath-like crystals; may be striated; two good cleavages not at right angles. ***Hornblende*** is the most common amphibole.

Biotite - generally black, shiny, flat crystals with one perfect cleavage; softer than hornblende; no striae; is part of the mica group.

Igneous-Rock-Classification Table

Table 5.3 lists the names of common igneous rocks based on their texture (vertical axis) and mineral composition (horizontal axis). By estimating the overall mineral composition of an igneous rock sample (represented by the proportion of a vertical line that falls within each mineral field of the diagram) one chooses a family of igneous rock types. Identifying the texture of the sample then allows one specific type of rock to be chosen from the family. Example A on Table 5.3 and Data Sheet 5.1 illustrates how to use this chart. In this case, felsic minerals predominate (50% K-feldspar, 20% quartz, 20% albite), the mafic minerals (hornblende and biotite) constitute only 10% of the rock. This identifies the specimen as belonging to the rhyolite - granite family. Because the texture is one of large, visible crystals (phaneritic) the specimen is identified as a granite.

Table 5.3 - Principle igneous rocks classified according to mineral composition and texture.

Exercise 5.4

Identifying Igneous Rocks

The objective of today's lab is to learn to identify the most common igneous rock types by learning to distinguish their different textures and mineral compositions.

Your principal guide in identifying igneous rock specimens is the igneous-rock-classification chart (Table 5.3) on the preceding page. Information on specific igneous rock types can be found in the general description of igneous rocks in this chapter and in chapter 8. For estimating the percentages of different minerals in the specimens you will find Figure 8.1 very useful.

What to do:

Eleven igneous rock specimens are located in the large sample box at each lab bench. For each group of samples given below, identify the numbered specimens by determining their textures and mineral compositions and match the specimen number to the rock name given. Use Figure 8.1 and Table 5.3 to identify these rocks. For the phaneritic specimens try to estimate the percentages of different minerals in the rock. For aphanitic specimens simply note whether the specimen is felsic (light in color), mafic (dark in color) or ultramafic (black in color). A worksheet (Laboratory Data Sheet 5.1) is included at the end of this exercise for tabulating your identifications.

1. Begin by trying to identify the following specimens, which represent a typical range of igneous textures and compositions.

Specimens # 1, 2, 3, 4, 5, 6, 7

Basalt
Rhyolite
Andesite porphyry
Gabbro
Granite
Diorite
Peridotite

2. Having successfully identified the above specimens, identify the following **volcanic rocks**:

Specimens #8, 9, 10, 11

Scoria
Obsidian
Pumice
Amygdaloidal basalt

Descriptions of igneous rocks not shown in Table 5.3

<u>Volcanic rocks</u>

Scoria	Rapidly cooled, frothed basaltic rock, often very jagged with large gas-bubble cavities (vesicles).
Amygdaloidal basalt	A vesicular basalt in which the vesicles have been infilled with mineral crystals.

Laboratory Questions 5.4

1. Look at the igneous-rock classification table (Table 5.3) in your lab book. Notice that the rocks to the right of the center of the chart are called mafic and ultramafic rocks (label on top of chart) and the rocks to left of center are called felsic rocks.

   What is the most apparent visible difference between felsic- and mafic / ultramafic rocks?

2. Compare the following specimen pairs: granite / rhyolite, andesite / diorite, and basalt / gabbro.

   a. Are there differences in composition between the rocks in each pair (consult table 5.3)?

   b. What is the approximate mineral composition for each pair?

   c. What is the difference between the rocks in each pair?

   d. What causes this difference?

3. What is the cause of the abundant **vesicles** present in the scoria specimen? How are the amygdules in the amygdaloidal basalt related to vesicles?

Laboratory Data Sheet 5.1
Lab 5 - Igneous Rocks

#	Texture	**Mineral Composition** Minerals present and their approximate percentages in the rock*		Rock Name
		Felsic	**Mafic**	
example	phaneritic	K-feldspar 50% quartz 20% Na plagioclase 20%	hornblende 7% biotite 3%	granite
example	aphanitic	light-colored, pink		rhyolite

* Use general color for aphanitic specimens without visible mineral grains.

Lab 6 - Sediments and Sedimentary Rocks

PURPOSE

In Lab 6 you will examine and describe samples of nonlithified sediments using the binocular stereomicroscopes on the benches and/or your hand lenses. After you have done that, you will classify hand specimens of common sedimentary rocks. Read through the introductory paragraphs first to be sure that you understand the terms to be used and look ahead at the appropriate section in Lab 8. Finally, fill out the worksheet for the sediments and sedimentary rocks.

INTRODUCTION

Most of the rocks exposed at the Earth's surface are sedimentary rocks that formed by lithification of a pre-existing **sediment** (such as sand, clay, or lime mud). Three important geologic aspects of sediments are: (1) their appearance; (2) their origin; and (3) the environment (lagoons, deltas, rivers, dunes, shallow marine coasts, coral reefs, deep-sea floor, etc.) in which they were originally deposited. In today's exercise, we will concentrate on the appearances of sediments as a basis for classifying sediments and relating sediments to sedimentary rocks.

CLASSIFICATION OF SEDIMENTS

Sediments are classified according to the origin of their constituents. Four major categories are:

(1) **Clastic Particles** (or clasts), derived from the breakdown of some pre-existing rock and the physical transport, as individual solid objects, of the breakdown products to the site where they are deposited. [Many names are in use for sediments composed of such breakdown products that were transported as individual solids. Among these are **terrigenous sediment**, **extrabasinal sediment**, **detritus** (or **detrital sediment**), **clastic sediment**, and **siliciclastic sediment**. Each geologist displays a marked preference for using one of these several terms. No consensus has ever existed nor is ever likely to.]

(2) **Biogenic Particles** secreted by organisms using ions dissolved in the water. Some geologists use the term **skeletal debris** for solid materials secreted by organisms.

(3) **Crystals** precipitated chemically from ions dissolved in the water, usually as a result of evaporation of the water. (Accordingly, all geologists refer to such crystals as **evaporites**).

(4) **Ooids**, distinctive spherical objects that result from the combination of crystal growth by evaporation and agitation of the water in which such growth takes place.

A common feature of all clastic particles is evidence of abrasion during their physical transport as individual solid objects. Any initial sharp edges become rounded. Ultimately, no matter what they looked like when they were first broken loose from the bedrock, sedimentary particles become increasingly smooth and spherical the more they are transported.

In addition, after they have formed, biogenic particles secreted by organisms, crystals precipitated by evaporation, and ooids formed by combined evaporation and agitation, are also subject to being transported physically as individual solid objects. Accordingly, these objects can also display evidence of breakage and abrasion.

CLASSIFICATION OF SEDIMENTARY ROCKS

Geologists classify sedimentary rocks, using the same two attributes of texture and mineral composition used to classify igneous rocks. **Texture** refers to the sizes and uniformity of sizes of the constituent particles of sediments and sedimentary rocks, and includes the attribute of **shape**. In sediments and sedimentary rocks, shapes of particles are important because during many processes of sediment transport, initial angular shapes become rounded as a result of abrasion. **Composition** refers to the mineral composition of the rock. An important aspect of mineral composition in clastic sediments is **compositional maturity**, which refers to the proportion of easily decomposed minerals (such as feldspars and mafic minerals) and non-weathered fragments of still-recognizable rocks (**rock fragments**) to weathering-resistant minerals (such as quartz and clay). A **compositionally immature** sediment is one that contains feldspars and/or rock fragments, components that generally do not survive intense chemical weathering. A **compositionally mature** sediment contains only resistant minerals such as quartz, garnet, and iron oxides.

A second aspect of mineral composition in sediments and sedimentary rocks centers around the nonsilicate minerals, such as calcite, dolomite, gypsum, or halite. Such minerals were not eroded directly from bedrock as solids but rather came from the bedrock in a dissolved state and were **precipitated** into solids within the water of the basin of deposition, either by evaporation or the activities of organisms. Only rarely is calcareous rock eroded and re-deposited as calcareous clastic particles.

Classification of Sedimentary Rocks by Texture

Fragmental texture: The fragmental texture (also known as mechanical or detrital) is the texture that forms when solid objects (particles or clasts) are transported by physical means. Three subvarieties are clastic, bioclastic, and pyroclastic.

Clastic texture: The clastic subvariety is characterized by particles that were broken. Such broken particles include older rocks, re-cycled sediments, or skeletal debris.

Bioclastic texture: If the broken particles consist of skeletal debris, then the texture is refered to as bioclastic.

Pyroclastic texture: The pyroclastic subvariety refers to rocks composed of tephra, volcanic materials broken up and transported by violent eruptions.

Crystalline texture: Among sedimentary rocks cyrstalline texture is confined to the so-called chemical rocks that are composed of soft minerals (calcite, halite, gypsum). The crystalline texture resulted from growth of minerals that were being precipitated chemically out of the water of the basin of deposition. Typical examples include the evaporites, rocks composed of minerals formed by evaporation of brines. These include halite rock, anhydrite rock, and gypsum rock.

Organic texture: This kind of texture is formed when organisms secrete massive frameworks as parts of their skeletons or colonies and these frameworks survive in the geologic record. Typically such framework rocks (called framestones) are composed of calcium carbonate. A typical example is the coral-reef rock found along many tropical shores.

Oolitic texture: Calcium carbonate is sometimes precipitated out of the waters of the basin of deposition in such a way that it accretes concentrically to form a coating around a nucleus particle (quartz, skeletal particle, clay particle etc.).

Such coated particles are named ooids and their appearance is very distinctive. A sediment or sedimentary rocks composed of ooids is said to possess **oolitic texture**.

Classification of Sedimentary Particles by Size

Geologists have adopted a standard scale of classifying sediments by size. Each sediment size is given a name that encompases a range of diameters, as shown in Table 7.1.

Table 7.1 - Sediment sizes, their bounding sizes (mm), and some common reference objects.

Particle Class Name	Range (mm)	Reference object
Boulder	>256	Volley ball
Cobble	256 >64	Tennis ball
Pebble	64 >4	Pea
Granule	4 >2	Head of wooden match (limit of visibility)
Sand	2 >1/16	
Very coarse	2 >1	
Coarse	1 >1/2	
Medium	1/2 >1/4	
Fine	1/4 >1/8	
Very fine	1/8 >1/16	
Silt	1/16 >1/256	
Clay*	<1/256	

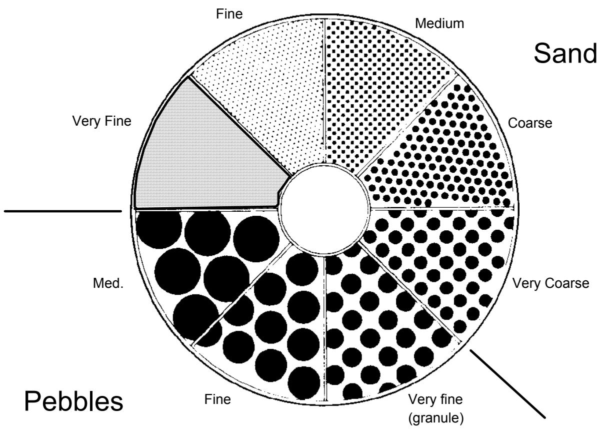


Figure 7.1 Diagram for estimating sediment size.

(*The term **clay** is used in geology in two ways: (1) as a mineral name in the sense of clay minerals and (2) as a size class for the smallest particles. Most clay minerals occur in the clay sizes).

Aggregates (populations) of sediments

As long as all the sediment particles fall into only one of these size classes, no problem exists for giving it a name: it is a fine sand, coarse sand, silt, or whatever. In the real world, however, a typical sediment consists of an aggregate (or a population) of particles from more than one of these size classes. In that case, the name of the predominant size class in the population is used as the sediment name; the names of the less-abundant size classes are placed ahead as modifiers, starting with the least abundant and working upward. Thus, a pebbly-sandy silt could consist of some pebbles, more sand, and mostly silt.

The term **sorting** refers to the range of sizes within a sediment. In a well-sorted sediment, only a few size classes are represented. In a poorly sorted sediment, many sizes classes are present. For example, a typical beach sand is well sorted, but a mixture of sand, pebbles, and mud is poorly sorted. Sorting is a function of the range of sizes available and the kind of selectivity on the basis of size exerted by the processes of transport and -deposition. As a first approximation, try to determine the extent to which the finest sizes (silt and clay) have been segregated from the sand-size- and coarser particles.

In laboratory-based textural analysis of sediments, sieves of standard mesh openings are used. The sediment is passed through the sieves and the amount trapped on each is weighed. Another method makes use of settling speeds in a column of water. Today, we will fall back on the old "eyeball" method and estimate the proportions of the different classes by comparing them against reference sets of sieved fractions.

IDENTIFICATION OF SEDIMENTARY ROCKS

Minerals of Sediments / Sedimentary Rocks

Sedimentary rocks are characterized by abundant quartz (can be up to 100%), clay, and/or the "soft" minerals (calcite, dolomite, gypsum, halite). Thus, as a first approximation, we can organize sedimentary deposits into two large categories based on mineral composition: (1) those containing silicate minerals and/or recognizable rock fragments (**clastic group**); and (2) those lacking silicate minerals and containing the "soft" minerals such as calcite, dolomite, gypsum, and halite (non-clastic- or **chemical group**). Those of category (1), the clastic group, include sediments reworked from older bedrock or from volcanic rocks (a special subcategory called **pyroclastic**) that were transported physically as particles to the basin of deposition. By contrast, those of category (2), the chemical group, consist of materials that were transported to the basin of deposition dissolved in the water and later precipitated as as shell by an organism or as an evaporite. After precipitation, however, these chemical sediments can be transported physically as particles.

Silicate minerals. The silicate minerals in sedimentary rocks are dominated by **quartz** and the **clay minerals**. The feldspars, so predominant in igneous rocks, belong in the second rank among sedimentary minerals because feldspar usually does not survive to be present in sedimentary rocks. However, under some conditions of climate and/or erosion, it does survive- or escape weathering. Other minor silicate constituents of sediments include the co-called "heavy" minerals (specific gravity >2.8) such as magnetite, ilmenite, hornblende, biotite, and various minerals found in metamorphic rocks such as garnet.

Nonsilicate minerals. The nonsilicate minerals of sedimentary deposits are dominated by the minerals **calcite** and **dolomite** (collectively called **carbonates**). Calcite generally is secreted as the shells of invertebrate organisms (such as clams, snails, corals, bryozoa, echinoderms, and tiny single-celled organisms such as foraminifera) or as plates inside the tissue of marine algae. Common evaporite minerals include **halite** (rock salt), **gypsum**, and **anhydrite**.

Using the Classification Chart

Table 6.2 shows a simple classification of sedimentary rocks. Two major groups are: (1) clastic (for the silicate-mineral group) and (2) chemical (for the nonsilicates).

Clastic rocks. The key point in recognizing clastic sedimentary rocks is that they all display clastic or detrital texture. Notice that in the clastic group, the names of the main groups of sedimentary rocks correspond with sediments based on size: coarse sediment (particles predominantly larger than 2 mm and called **boulder, cobble, pebble or granule**) form rocks known as **breccia** and **conglomerate** in Table 7.2. These coarse-textured rocks are subdivided according to the shapes of particles. If the coarse particles have been rounded, the rock is a **conglomerate**. If the coarse particles are angular, then the rock is a **sedimentary breccia**. Medium-size sediment (**sand**) when lithified becomes **sandstone**; **silt** becomes **siltstone**; and **clay** becomes **claystone**. The term **shale** refers to a fine-textured sedimentary rock (particles in silt- and/or clay-size ranges) that is **fissile**, meaning that it will split readily into thin slabs. Two subvarieties of sandstone are recognized on the basis of their immaturity: (a) **arkose** (if the content of feldspar >25%); and (b) **graywacke** or **lithic sandstone** (quartz sand with a high concentration of dark rock fragments and some clay).

Sandstones. Clastic rocks with a sandstone texture can easily be recognized by feel. When rubbed between the fingers, quartz sandstones, arkoses, and graywackes will feel abrasive (like sandpaper).

Shale: composed of clay minerals, quartz, and possibly feldspar; may include calcite. Hard shales break parallel to the stratification; this may be expressed by laminae that are along the sides of the specimen, but parallel to the top and bottom. Especially tough, fine-textured sedimentary rocks that are too hard to be called claystone are designated **argillite** (from the French word argile, which means clay).

Chemical rocks. The chemical- or nonclastic sedimentary rocks are easy to recognize because they consist almost exclusively of "soft" minerals (Mohs hardness of 3 or less). The nonclastic rocks are subdivided into evaporite- and carbonate subdivisions and several textural varieties are represented. Among rocks composed of calcite (=limestones), two common textures are **bioclastic** and **oolitic**. Some limestones and nearly all the rocks composed of evaporite minerals display crystalline textures that are subdivided according to size into fine, medium, and coarse, as in phaneritic igneous rocks.

Carbonate rocks: composed of calcium-carbonate minerals (ex: calcite or dolomite. Can be recognized by hardness of 3 and reaction with dilute hydrochloric acid; look with hand lens to see where the rock is effervescing--it might be a calcite-cemented sandstone or -siltstone). Diagnostic features include skeletal debris, organic textures, or oolitic texture. A few coarsely crystalline limestones and dolostones resemble calcite- or dolomite marbles. (See next section for further discussion.)

It is important to distinguish limestones from **dolostones**, nonclastic sedimentary rocks composed chiefly of the mineral dolomite. Limestones consist of calcite, which effervesces freely in cold dilute hydrochloric acid. The mineral dolomite displays many of the same properties as calcite (Mohs hardness of 3; three cleavages not at right angles), but reacts with dilute hydrochloric acid only slowly and only with a little help. The "help" consists of powdering a small portion of the specimen. Make sure that you are able to recognize a dolostone by the weak effervescence of the mineral powder (use a nail or the point of your knife from your geology test kit to scratch the specimen where the acid is; check for the bubbles by looking through a binocular stereomicroscope or with your hand lens).

Evaporites: Include one-mineral rocks, such as gypsum rock, halite rock, and so forth. These have a crystalline texture but are composed of soft minerals that do not react with dilute acid. Halite rock tastes salty.

Two other nonclastic sedimentary rocks you should be able to recognize are **chert** and **coal**. Chert is a hard, very-fine-textured rock that consists of microcrystalline quartz. Chert breaks with conchoidal fracture like window glass; its luster is usually waxy, and its Mohs hardness is 7. **Coal** is combustible solid organic matter consisting of the carbonized remains of land plants. Various grades of coal begin with **peat** and include **lignite** (brown coal) and **bituminous coal**. You can recognize bituminous coal because of its shiny luster, blocky fracture, and because it will leave a black, dusty residue on your hands when you touch it. Because coal is made of carbon, it feels very light relative to other rocks.

Table 6.2 - Classification Chart for the Common Sedimentary Rocks.

Laboratory Exercise 6.1
Identifying Sedimentary Rocks

Identifying Sedimentary Rocks

The objective of today's lab is to learn to identify the most-common sedimentary rock types by learning to distinguish their different textures and mineral compositions.

Your principal guide in identifying sedimentary rock specimens is the sedimentary- rock-classification chart (Table 6.1). Information on specific sedimentary rock types can be found in the general description of sedimentary rocks in this chapter. More information useful for identifying sedimentary rocks can be found in the introduction to Laboratory 8.

What to do:

Eight sedimentary rock samples are in the large brown wooden drawers marked in yellow. Use Table 6.1 to learn to identify these rocks. A worksheet (Laboratory Data Sheet 6.1) is included at the end of this exercise. Write down the as much information as you can glean from each specimen on the worksheet. This will help you to identify the specimens successfully. The **environment of deposition** (the type of place where sediments that make up a particular kind of sedimentary rock accumulate) for each rock is also given in Table 6.1.

Specimens # 1, 2, 3, 4, 5, 6, 7, 8, 9, 10, 11

Shale
Quartz sandstone
Arkose sandstone
Rock salt
Chert
Quartz-pebble conglomerate
Dolostone breccia
Coarse-grained limestone
Coal
Coquina
Greywacke

Laboratory Questions 6.1:

1. What are the characteristics of a **detrital** or **clastic** rock? Which specimens are detrital?

2. What are the characteristics of a **carbonate** rock? Which specimens are carbonates?

3. What are the characteristics of an **evaporite** rock? Which specimens are evaporites?

4. What substance is **coal** made out of and how does it form?

LABORATORY DATA SHEET 6.1

#	Sediment type	Grain size	Rock texture and Mineral composition	Environment of Deposition	Rock Name
example	clastic	sand	well sorted quartz	beach, desert, river bed	quartz sandstone

Lab 7 - Metamorphic Rocks

PURPOSE

Now that you have mastered the identification of igneous- and sedimentary rocks in the past two laboratory sessions, you should be ready for today's study of metamorphic rocks. We will analyze their unique textures and mineral compositions, and discuss the importance of regional- and contact metamorphism.

The ultimate goal for today is for you to appreciate, and to interpret the history of metamorphic rocks via identification. Because metamorphic rocks occupy regions of the continental crust that were affected by lithospheric-plate interactions in the geologic past, integrated studies of metamorphic terrains can teach geologists about the plate-tectonic history of the Earth. Thus, the study of metamorphic rocks constitutes a complexly interwoven subdiscipline of geology that utilizes field work, stratigraphic- and structural analysis, as well as petrology and geochemistry, to map, study, and interpret the results of the process of metamorphism and the structural development of mountain belts. Together with information acquired from detailed study of adjoining igneous- and sedimentary regions, geologists have been able to develop a history of the Earth dating back roughly 4.5 billion years (Ga).

INTRODUCTION

Recall from our general introduction to rock classification that among most metamorphic rocks, the unity factor is place (inside the Earth) and not one, but two diversity factors must be included. These are **parent rock(s)** and **metamorphic process(es)**. As we shall see, in some cases, the resulting metamorphic rock is determined largely by the kind of parent rock. In other cases, the same parent rock, for example shale, as it is subjected to progressively more-intense ("higher-grade") metamorphism, can yield a variety of different metamorphic rocks.

Metamorphic rocks form in response to four major types of metamorphism:

1) Regional-
2) Contact-
3) Dynamic-, and
4) Shock metamorphism

In terms of the amount of metamorphic rock in the Earth, categories 1 and 2 are the most important. In brief, **regional metamorphism** causes changes in the parent rock as a result of elevated heat and pressure, often associated with collisions between lithosphere plates and the formation of mountain belts. Regional metamorphism affects rock throughout very large areas. **Contact metamorphism** is metamorphism resulting from the intrusion of a hot igneous mass (a **pluton**) into the surrounding country rock or from the flow of lava over rock at the Earth's surface. The heat from the pluton or lava flow literally 'bakes' the country rock around it, causing a variety of textural and compositional changes. Contact metamorphism is usually limited to a relatively small area surrounding the igneous rock. Consult your geology textbook for a more detailed explanation of metamorphic processes.

CLASSIFICATION OF METAMORPHIC ROCKS

As with igneous rocks and sedimentary rocks, geologists classify metamorphic rocks on the basis of textures and mineral compositions. A further feature, **fabric**, is an important- and distinctive attribute of metamorphic rocks.

Textures- and Fabrics of Metamorphic Rocks

Textures. Most metamorphic rocks display crystalline textures that have resulted from the growth of minerals during metamorphism. As with the textures of igneous rocks, those of metamorphic rocks may be uniform or nonuniform. Metamorphic rocks having uniform texture are described as **granoblastic**. The sizes of the constituents in metamorphic rocks can range from fine to coarse. Three basic categories exist here (also refer to Lab 8 for a discussion of uniform vs. nonuniform textures):

Very fine - minerals not visible to the unaided eye.
Fine - minerals barely visible to the unaided eye.
Medium to coarse - minerals easily visible.

If the sizes of the constituents are not uniform, the rock texture appears similar to a porphyritic igneous rock. The textures of metamorphic rocks whose constituents are not of uniform sizes but contain some crystals that are distinctly larger than the rest are said to be **porphyroblastic** (compares with porphyritic texture of some igneous rocks). In such metamorphic rocks having nonuniform textures, the larger crystals are named **porphyroblasts** (compare with igneous phenocrysts).

Despite some similarities with the crystalline textures of igneous rocks, however, no simple relationship exists in metamorphic rocks between sizes of constituents and rate of cooling. In fact, porphyroblasts are younger than the surrounding matrix (in contrast to the early formed phenocrysts of igneous rocks).

Another kind of texture in metamorphic rocks is not the result of crystal growth, but just the opposite. As a result of pervasive- and intense shearing, the constituent minerals have been mechanically broken. The distinctive textures resulting from such breakage are named **cataclastic**. In metamorphic rocks, cataclastic textures are formed when rock is crushed, sheared, and generally chewed up. This kind of thing usually accompanies conditions where the rocks were brittle. Such conditions are present near large faults in the upper 10 kilometers of the Earth’s crust.

Fabrics. The term **fabric** refers to the orientation of nonspherical mineral grains in rocks. Two major categories are: (1) **random** and (2) **oriented**.

In random fabric, there is no distinct orientation to the nonspherical grains. Random fabrics are typical of most igneous rocks and of many sedimentary rocks. Random fabrics can form in metamorphic rocks, but only if: (1) the fabric of the parent rock was random, and (2) the heat responsible for metamorphism was not accompanied by pressure and deformation (in other words, if the metamorphism was **static** as contrasted with **dynamic**).

The fabric is said to be oriented if elongate- or flattish mineral grains are parallel to a preferred direction that was determined by the conditions of metamorphism. Oriented fabrics are typical of metamorphic rocks, but are known also in some igneous- and sedimentary rocks. Preferred orientation of platy minerals in metamorphic rocks yields a layered- or parting property named **foliation**. We discuss foliation further in a following section devoted to progressive metamorphism. Suffice to say that metamorphic rocks are either **foliated** (consisting of aligned flaky-, tabular-, or prismatic minerals) or **non-foliated** (minerals not visibly orientated). Non-foliated rocks with granular, equidimensional mineral grains are termed granoblastic.

Mineral composition of metamorphic rocks

The minerals of metamorphic rocks include many of the same silicate minerals that form igneous rocks (feldspars, quartz, pyroxenes, amphiboles, biotite, muscovite) and some of the distinctive nonsilicate minerals of sedimentary rocks (notably calcite and dolomite). The essential minerals of metamorphic rocks include your notorious mineral buddies, quartz, feldspar, and the micas. In addition, metamorphic rocks may include garnet, staurolite, talc, and some distinctive green minerals (such as chlorite, epidote, and those of the serpentine group) that we have not encountered in the other rock groups, as well as some silicate minerals in which aluminum is notably abundant (andalusite, kyanite, and sillimanite).

During metamorphism, the minerals in the parent rock are exposed to conditions that tend to re-build the rock-forming silicates that were created in igneous rocks and weathered into quartz and clay to make sedimentary rocks. However, once the separation of minerals during a sedimentary cycle has created a pure sandstone (100% quartz) or pure carbonate rock (100% calcite or dolomite), not much mineralogic change can be caused by metamorphism. The very-reactive minerals during metamorphism are the clay minerals of shales; these minerals change during metamorphism into micas (muscovite, chlorite, biotite), feldspar, and quartz. The contrasting paths of metamorphism can be seen in two examples, one where the parent rock exerts the predominant influence and another where progressive metamorphism of the same parent rock yields different metamorphic rocks.

Parent Rock Determines Metamorphic Product

In the following four examples, the composition of the parent rock determines the composition of the metamorphic rock:

Parent rock	**Mineral composition**	**Metamorphic rock**
Sandstone	Quartz	Quartzite
Limestone	Calcite	Calcite marble
Dolostone	Dolomite	Dolomite marble
Granite	Quartz, feldspar(s) biotite	Granite gneiss

Notice that in these examples, the mineral composition of the parent rock is the same as the mineral composition of the metamorphic rock. During metamorphism, the texture has been changed so that the resulting metamorphic rock is clearly crystalline and, in the case of granite-gneiss, an oriented fabric has been formed from the random fabric of the parent rock granite. These similarities of mineral composition raise the question of how it is possible to distinguish between the parent rock and the metamorphic rock. The following remarks may be helpful as you search for answers in each case.

Quartzite vs. sandstone: Both the sandstone parent rock and the metamorphic rock quartzite may consist of 100% quartz. Thus, the rock does not react at all with dilute hydrochloric acid and is very hard. Your knife will not scratch the rock; indeed, the rock will scratch your knife blade. Quartzite is distinguished from sandstone by the smoothness of broken surfaces. In quartzite, the rock breaks **across** the individual quartz grains. In the sandstone parent rock, a broken surface feels "rough" (like sandpaper) because the rock breaks **around** the quartz particles, between the clasts and the cement. You will have to use a hand lens or one of the binocular stereomicroscopes to convince yourself of these distinctive ways of breaking with respect to the quartz constituents.

Calcite marble vs. limestone: Both the limestone parent rock and the calcite marble consist of calcite and thus can be scratched with a knife blade and will effervesce immediately with dilute hydrochloric acid. The limestone parent rock may display fossils, show bioclastic- or oolitic texture, and be gray to dark gray. In calcite marble, glistening crystalline texture is typical and rhombohedral cleavage is present in coarse-grained specimens; traces of fossils rarely remain and the color is light. Distinguish calcite marble from dolomite marble by the acid / scratch test.

Dolomite marble vs. dolostone: Distinguishing between these two rocks may be the most difficult challenge you will face. The difficulty arises because the texture of both rocks is crystalline and they both consist of the mineral dolomite. Let the coarseness of the crystalline texture be your guide: if the dolomitic rock is coarsely crystalline, call it dolomite marble. Otherwise, identify it as dolostone. If you can see traces of oolitic texture, then you are looking at a dolostone, not a dolomite marble. The key points about both rocks are their softness (can be easily scratched with a knife) and acid reaction that requires assistance (make some rock powder by scratching or heat the acid; only after such assistance will the rock effervesce--and then only slowly, never vigorously as in calcite-rich rocks).

Granite-gneiss vs. granite: The fabric of the parent rock granite is random. The fabric of the granite-gneiss is oriented. This oriented fabric is expressed by a well-developed gneissic layering or foliation. The mineral layers may appear streaky, but are clearly visible by the alternation of light-colored minerals (feldspars and quartz) with dark-colored biotite or amphibole.

Progressive Metamorphism of Same Parent rock

In contrast to the previous examples in which the parent rock determines the metamorphic rock, we now consider two cases in which the same parent rock, shale or basalt, yields a different metamorphic rock as a function of **degree** (how hot and at what depth) of metamorphism. The fact that a given parent rock yields different metamorphic products as metamorphic temperatures increase is the basis for recognizing **metamorphic facies**.

Shale Parent rock: Progressive Metamorphism to Slate, Phyllite, Mica Schist, and Gneiss. The growth of new mica in shale marks the onset of metamorphism and its transformation into slate. As such, a **slate** differs from a shale in that tiny crystals of mica (you may need your hand lens to determine this) adorn the broken rock surface. These give the slate a very subtle sheen in reflected light. The metamorphism has oriented the mica crystals and hardened the rock resulting in a slaty cleavage or sheet-like structure to the specimen that may or may not be parallel to the original stratification of the shale. Slate roofs, billiard tables, and blackboards owe their extreme flatness to slaty cleavage. The potential for nonparallelism between cleavage and bedding results in layers that are not parallel with the flat sides of the specimen (with breakage along slaty cleavage not parallel to original bedding). Because of the effects of recrystallization inherent in rocks with slaty cleavage, they also exhibit a metallic ring when tapped with a nail.

Progressively higher temperatures and pressures yield **phyllite**, **schist**, and ultimately, **gneiss**. In phyllites (a favored decorating stone for Greek coffee shops on the upper east side of Manhattan) the growth of micas imparts a shine to the altered slaty cleavage that is unmistakable when the specimen is rotated. In schists, the micas are quite coarse and compose more than 50% of the mineral content with the development of an obvious foliation. Destruction of original bedding and fossils is complete. In gneissic rocks, some degree of internal melting has taken place and the rock has segregated into mica-rich- and mica-poor layers that together define the gneissic layering. The destruction of mica occurs at the highest grades with the crystallization of feldspar(s). Gneissic rocks with granitic (igneous) zones caused by partial melting are called **migmatites**.

Basalt Parent rock: Progressive Metamorphism to Chlorite-rich Slate, -Phyllite, -Schist, Amphibolite, and Amphibole Gneiss. The growth of new mica also marks the onset of metamorphism in basaltic rocks. As such, ferromagnesian mica (chlorite) and various members of the amphibole group (including hornblende, tremolite, and actinolite), take center stage with the formation of slate, phyllite, and schistose rocks rich in these constituents. Serpentine-group minerals (including talc and asbestiform varieties) are common in metamorphosed basalts. At progressively higher metamorphic grades, first amphibolites (foliated plagioclase + amphibole rocks) then amphibole gneisses predominate.

Helpful Hints for Identifying Common Metamorphic Rocks

The following is a short list of comments on some of the metamorphic rocks that may help you in your identifications. Use this in conjunction with Table 7.1, the metamorphic-rock-classification chart.

Quartzite: may be 100% quartz; distinguished from sandstone by breakage of rock across the particles. Grains appear indistinct. Lacks the sandpaper feel of sandstone.

Marble: may contain 100% carbonate minerals (relatively soft- calcite or dolomite); distinguished by coarsely crystalline texture and generally light color; fossils rarely preserved.

Slate: distinguish from shale by metallic ring when tapped with nail (shales give dull thud); also, laminae that are not parallel with the flat sides of the specimen (breakage along slaty cleavage not parallel to bedding).

Phyllite, schist, and gneiss: these rocks all are distinctly foliated in hand specimen. Look for a distinct shine to recognize phyllite, large, oriented flakes of mica in schists, and distinct layering of minerals including quartz and feldspars in gneiss.

Table 7.1 - Metamorphic Rock Classification

Laboratory Exercise 7.1
Identifying Metamorphic Rocks

OBJECTIVE

The objective of today's lab is to learn to identify the most-common metamorphic rock types by learning to distinguish their different structures, textures, and mineral compositions.

Your principal guide in identifying metamorphic rock specimens is the metamorphic-rock-classification chart (Table 7.1). Information on specific metamorphic rock types can be found in the general description of metamorphic rocks in this chapter.

What to do:

Eight metamorphic rock samples are located in the large sample box at each lab bench. Use Table 7.1 to learn to identify these rocks. A worksheet (Laboratory Data Sheet 7.1) is included at the end of this exercise. Write down the as much information as you can glean from each specimen on the worksheet. This will help you to identify the specimens successfully. The **parent rock** (the type of igneous or sedimentary rock that becomes a particular metamorphic rock type) is also given in Table 7.1.

Specimens # 1, 2, 3, 4, 5, 6, 7, 8, 9, 10, 11

Calcitic marble
Quartzite
Felsic gneiss
Chlorite schist
Slate
Phyllite
Serpentinite
Porphyroblastic schist
Metaconglomerate
Mica schist
Amphibolite

Lab Questions:

1. What is the difference between a **foliated** and a **non-foliated** rock? Which specimens are foliated?

2. List the different metamorphic rocks that can be formed from each of the following parent rocks:

   a. Claystone
   b. Granite
   c. Sandstone
   d. Limestone
   e. Basalt

Laboratory Data Sheet 7.1

Lab 7 – Metamorphic Rocks

#	Texture / Description	Grain size	Mineral Composition	Parent Rock	Rock Name
example	Foliated- platy micas	Obvious	Muscovite mica some quartz	felsic, shale?	Schist

Lab 8 - Integrated Rock Identification

The objective of lab 8 is to review rock identification and to learn to distinguish among the three types of rock - igneous, sedimentary, and metamorphic. Your task is to examine each of 24 unknown specimens and attempt to determine rock type, texture, mineral composition and identification. For many people, the most challenging aspect of rock identification is determining rock type. For example, it is easy to identify a dull, black rock as a basalt if you already know you are dealing with an igneous rock. But there are also dull, black sedimentary rocks (mudstones) and metamorphic rocks (hornfels, slate). How do you determine to which category a rock belongs before you have identified it? This is the challenge of today's lab! In addition, the questions for **[R2]** will be assigned today and will be due next week, at the start of Lab 9, the day of the rock practicum.

HINTS FOR IDENTIFYING ROCKS IN HAND SPECIMENS

When you are given a hand specimen of a rock and asked to identify it, where do you start? At first glance all rocks may look alike--how is it possible to tell them apart? The secret of success in identifying rocks lies in establishing a set procedure to follow in each case. Compile a check list of simple tests to make and of questions to ask yourself. Then carry out the tests systematically and answer the questions.

A good place to begin this review is with the **definition of a rock**, namely an aggregate of minerals in which many (as opposed to only one) crystals are present of one or more than one mineral. Moreover, the processes that form rocks leave behind reliable clues in the form of the **texture**. We list some simple tests to apply as a starting point in trying to identify rocks and then include some hints for how to take a first close look at rocks. Then comes our list of questions to ask, arranged under four major headings.

SIMPLE TESTS TO PERFORM ON ROCKS

Three simple tests should be applied to any rock specimen: hardness; dilute hydrochloric acid; and density.

HARDNESS: Try to scratch the rock with a knife blade or a nail. If you can't scratch the rock, then you know right away that it probably consists of silicate minerals (the hardness of most silicates falls in the range of 5 to 7; exceptions are some layer-lattice silicates, such as minerals of the mica group and talc). Apart from these flaky "soft" silicates, the other soft minerals (hardness <5) are abundant only in some sedimentary rocks (such as limestone, dolostone, gypsum rock, halite rock, and shale) or metamorphic rocks derived from these (calcite marble, dolomite marble, etc.). If you can scratch the rock, be sure to apply the acid test.

REACTION WITH DILUTE ACID: If you detect a "soft" rock, then apply a drop of dilute hydrochloric acid and pay attention to what happens. Vigorous bubbling indicates calcite (the chief mineral in limestones and calcite marbles and possibly the cement in sandstones and siltstones and shales; use your hand lens to see where the tiny bubbles are coming from. If the calcite forms the cement of a sandstone, for example, the bubbles will be coming from the spaces between the quartz particles). If no bubbling takes place, do not end the test just yet. Powder a portion of the rock with a knife or nail and check to see if the powdered rock effervesces. If the powder reacts faintly, then the mineral dolomite is present and the rock may be a dolostone or a dolomite marble. Silicate minerals do not react with dilute acid.

DENSITY: The density of a rock is a function of the density(ies) of the constituent mineral(s) and of how much air-filled pore space is present. Dense rocks are composed of dense minerals closely arranged with minimum pore space. From the first laboratory session on the size- and shape of the Earth, you are already familiar with the densities (Table 1.1) of a few minerals (quartz and feldspar) and of some rocks (granite-gneiss, basalt, sandstone, limestone). An example of a dense rock is the igneous rock dunite (density 3.3), composed of olivine (density in range of 3.2 to 4.3). Two low-density rocks are obsidian (density 2.29) and halite rocks (density 2.0).

After you have completed these three tests, take a first close look with your hand lens (or use one of the binocular-stereo microscopes on the lab table).

EYEBALL TEST, THEN LOOK USING 10X MAGNIFICATION: One of the first things that may strike you about a rock is its color. Look for the presence or absence of quartz. Check for cleavage in the mineral grains by determining the way minerals have broken.

Minerals. What minerals are present? Can you identify the minerals? What proportion of the volume of the rock does each mineral occupy? (Use visual charts [Figure 8.1] to practice estimating abundances.) What do the minerals indicate about the history of the rock?

Rocks are composed of minerals and the number of minerals in each rock is small: one, two, or three, rarely more. Many rocks consist of silicate minerals and of these, the feldspars are the most abundant, especially in igneous- and many metamorphic rocks. The other abundant minerals are the ferromagnesian silicates (olivine, pyroxene, hornblende, and biotite) and quartz. Details of the minerals present in rocks are found under the appropriate labs (Labs 5, 6, and 7) describing the classification of the various rock types.

Color. As with minerals, color can be helpful in arranging rocks into certain general categories, but it is not in itself a diagnostic property. An important objective of your first look at the rock under low-power magnification is to find out about the color. Is it from the color of the main minerals (such as a green rock composed of olivine, chlorite, or green hornblende; or a black rock composed of biotite or black pyroxene, for example) or is the color coming from what is known as a **pigment**, that is a colored mineral powder, such as hematite, which coats noncolored particles such as quartz in some sandstones? Among igneous rocks color is a reliable expression of the general composition of the rock. Light-colored igneous rocks are **felsic** (rich is potassium feldspar, albite, and quartz), whereas dark-colored igneous rocks are **mafic** (rich in ferromagnesian silicates and calcium-rich plagioclase). The color of most obsidian is black as is that of basalt glass. The difference between them is than obsidian is translucent on thin edges whereas basalt glass is everywhere opaque.

Quartz. If quartz is the only mineral present (key features: hard, vitreous luster, and conchoidal fracture, but particles may have been abraded and rounded), then the rock may be a sandstone (sedimentary) or a quartzite (metamorphic). To tell these two rocks apart, you must proceede to the next test of breakage style. The important point associated with breakage style is to determine if the rock has broken AROUND the individual quartz particles, as in sandstone (sedimentary), or ACROSS the quartz particles, as in quartzite (metamorphic).

SOME QUESTIONS TO ASK

Here are some important questions arranged under four major headings:

(1) What is the **TEXTURE**? What size(s) are the constitents? Are the individual constituents all of about the same size, or are some distinctly larger than the others?

(2) What is the **FABRIC**? Are constituents elongate or flattened? If so, are they oriented at random, or are their long axes or planes of flattening aligned parallel to a preferred direction?

(3) Are any **LAYERS** present?

In rocks, layers result from differences in composition or other properties. As a first approximation, sedimentary rocks and layers are synonymous. But, this does not mean that layers are found only in sedimentary rocks. Layers can be present in any kind of rock. We review some examples of layers in each of the three rock families.

Layers in igneous rocks

Two processes can lead to layers in igneous rocks: (1) during cooling of cumulate-type igneous rocks, early-formed crystals that are denser than the magma may settle to the bottom of the magma chamber and there be spread out in horizontal layers, one layer at a time, with the oldest at the bottom and successively younger upward, as in the characteristic bedding of sedimentary rocks; (2) the spreading out on Earth's surface of volcanic materials.

Layers in sedimentary rocks

Layers, formed by processes of transport and deposition, are typical of sedimentary rocks. Layers at all thicknesses, from paper thin on up to many meters, can be present. The term **stratum** (pl. **strata**) designates layers spread out one at a time, the oldest at the bottom, as with sediments or lava flows).

Layers in metamorphic rocks

Layers in metamorphic rocks, named **foliae** (plural; singular is folia), result from carrying over of sedimentary- or volcanic strata during metamorphism, or are products of extreme shearing during metamorphism. Thus, in metamorphic rocks several kinds of layers can be present.

After preliminary study enables you to answer these three questions you should be able to decide if the rock is igneous, sedimentary, or metamorphic using Table 8.1. After you have assigned the rock to one of these three families, the next step is to give the rock a name. In order to do that, you must proceed to the last group of questions under the fourth major heading:

Using this general philosophical approach you should be able to narrow down the possible choices and name the rock type with a high probability of getting the correct group. Table 8.1 is a flow chart you can use to help you determine which of the three rock groups a specimen belongs to.

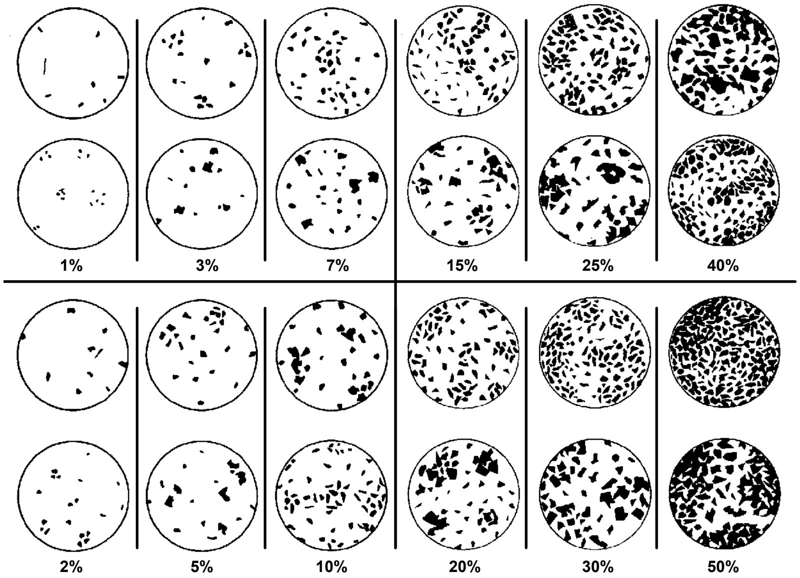


Figure 8.1 - Chart for visually estimating percentages.

Table 8.1 - Rock type determination chart

Laboratory Exercise 8.1
Identifying All Rock Types

OBJECTIVE

In today's lab you must pull all of your rock identification skills together to distinguish between specimens drawn from all three rock types. Examples of each kind of rock we have seen during the last three weeks are mixed together. Decide which specimens belong in each rock category and then identify them.

What to do:

Twenty four mixed rock specimens are numbered in white and located in the large sample box at each lab bench. Use Table 8.1 to help identify these rocks. Begin by attempting to separate the specimens into three groups representing igneous, sedimentary, and metamorphic rocks. After you have done this, try to identify each individual specimen. Record your identifications Laboratory Data Sheet 8.1.

Specimens # 1 - 24

Igneous
Basalt
Rhyolite
Andesite porphyry
Gabbro
Granite
Diorite
Pumice
Obsidian

Sedimentary
Shale
Arkose
Chert
Quartz sandstone
Greywacke
Conglomerate
Dolostone
Limestone

Metamorphic
Calcitic marble
Gneiss
Slate
Quartzite
Chlorite schist
Phyllite
Amphibolite
Mica schist

Laboratory Questions

1. Black shales, black fine-grained limestones, and basalt all look the same - they are dull, black, and fine-grained. How can you tell them apart?

2. What features would help you distinguish a crystalline metamorphic rock from a crystalline igneous rock?

3. What features would help you distinguish a sedimentary rock from a metamorphic rock?

LABORATORY DATA SHEET 8.1

Sample	Textures and / or other features	Igneous Metamorphic Sedimentary	Minerals Present (one or two)	Rock Name
1				
2				
3				
4				
5				
6				
7				
8				
9				
10				
11				
12				
13				
14				
15				
16				
17				
18				
19				
20				
21				
22				
23				
24				

Questions for Report 2 [R2] on Igneous-, Sedimentary-, and Metamorphic Rocks.

Aside from the appropriate chapters in this lab manual (Lab Chapters 2-5), look in the glossary and index of your textbook for a good start.

GENERAL

1. Define the terms **bedrock** and **regolith**. Discuss the main differences between them.

2. Describe the kinds of **layering** found in the igneous-, sedimentary-, and metamorphic rock groups. Which rock group is typified by its layering?

3. Define and provide a diagram of **The Rock Cycle**. Give examples of the processes that transform rock from one type to another through the cycle. Include a short discussion of the three rock types produced by the cycle.

4. List the **silicate minerals** studied earlier this semester and specify which are commonly found in igneous, sedimentary, and metamorphic rocks. Designate which are common in more than one rock type.

IGNEOUS ROCKS

5. Discuss the similarities and differences between **magma** and **lava**. How are they related to the major kinds of igneous rocks (intrusive vs. extrusive)?

6. Explain how cooling history and other factors create the common **textures** found in igneous rocks (aphanitic, phaneritic, porphyritic, vesicular, and glassy).

7. Using textural- and/or field relationships as evidence, how can one determine whether an exposure of rock was a former **buried sheet** of <u>extrusive</u> rock or a **sill** of <u>intrusive</u> rock?

8. Why do shield volcanoes and stratovolcanoes have different eruption styles? How does this relate to the different kinds of igneous rocks commonly produced by lava from the two types of volcanoes?

9. Define **phenocryst**. How can phenocrysts in a **porphyritic** rock be used to indicate flow direction in a magma or lava? What minerals are likely to form phenocrysts in:
   a. rhyolite
   b. andesite
   c. basalt

10. Using **Bowen's Reaction Series**, describe the process by which a magma having the original chemical composition of a **gabbro** (mafic rock) can ultimately crystallize into a **diorite** (intermediate rock).

11. Go to the library and find descriptions of two different **volcanic eruptions** (these can be found in encyclopedias, geology textbooks, and books about volcanoes or natural disasters). Write a short description of each eruption and then compare and contrast the two in terms of their style of eruption, type of lava and pyroclastics generated, destructiveness, etc.

12. How can geologists use the mineral composition of an igneous rock to provide rough estimates of the cooling history and the pressure (depth) and temperature profile of the area under examination?

13. Using the common igneous rocks found in your lab sets as reference points, describe the **textures** and **mineral composition** of the following:
    a. pumice and obsidian
    b. gabbro and basalt
    c. granite and rhyolite
    d. diorite and andesite

14. Examine the **Classification Table** for igneous rocks. Aside from silica (SiO_2) variation, describe the differences in elemental composition that exist between the felsic and ultramafic end members of igneous rock composition.

SEDIMENTARY ROCKS

15. Describe the processes of **weathering**, **erosion**, **transport**, **deposition**, **compaction** and **cementation** and how they combine to produce sediments and sedimentary rocks.

16. What are the main compositional differences between **clastic**, **chemical**, and **organic** sedimentary rocks? How do they differ in the source of their sediments?

17. Describe the differences in texture between a **breccia** and a **conglomerate**. Relate these textural differences to difference in the depositional history of the two rocks.

18. Define **sorting** of clastic sedimentary particles. Describe how and why sediments transported in a stream differ in their degree of sorting from sediments transported by a glacier.

19. A clean, well-sorted quartz sandstone usually contains well-rounded particles as well. Describe the process of **rounding** and explain why poorly sorted sediments are not typically well rounded.

20. Describe the process of **lithification** of sedimentary rocks including deposition, compaction, and cementation. What are the common cements produced in sedimentary rocks?

21. Relate the formation of the three types of **sandstones** (quartz, arkose, and graywacke) to their geologic environments of deposition. Describe the main differences between their respective mineral compositions and depositional histories.

22. Define and describe the formation of the following primary and secondary sedimentary rock features:
    a. bedding planes
    b. graded beds
    c. mud cracks
    d. ripple marks
    e. fossils

23. How can you distinguish a black limestone from a basalt? A black dolostone from a basalt? Discuss two methods for each pair.

24. Describe the wide range of important economic resources that are derived from sedimentary rocks. Include in your discussion sedimentary sources of both raw materials and fossil fuels.

25. Explain the role of sediments and sedimentary rocks in the **Rock Cycle** - use a diagram to explain your ideas.

METAMORPHIC ROCKS

26. Write a well-explained definition of **metamorphism**. Discuss three of the major types of changes that occur in rocks that are being metamorphosed.

27. What is meant by **parent rock** or **protolith**? How can **quartzite** be distinguished from sandstone? **Metaconglomerate** from conglomerate? **Granite gneiss** from granite?

28. Discuss the similarities and differences between the terms **porphyroblast** and **phenocryst**.

29. Describe the main characteristics by which **slate** can be distinguished from **shale**.

30. How can one distinguish a **dolomitic marbl**e from a **calcite marble**?

31. Discuss what the **parent rocks** of the following metamorphic rocks might have been:
    a. quartzite
    b. quartz-feldspar gneiss
    c. muscovite schist
    d. graphitic phyllite
    e. amphibolite
    f. serpentinite
    g. marble

32. How can you explain finding **foliated** and **non-foliated** metamorphic rocks in the same area? How can they both be the products of regional metamorphism?

33. Explain the role of metamorphic rocks in the **Rock Cycle** - use a diagram to explain your ideas.

34. Discuss the relationship between mountain formation and metamorphism. What do geologists learn about the Earth from studying metamorphic rocks?

35. How are metamorphic rocks economically important? Describe some valuable resources that are derived from metamorphic rocks.

Lab 9 - Rock Practicum and Introduction to Maps

PURPOSE

The purpose of today's laboratory meeting is to test your proficiency in identifying unknown rock specimens as provided by your lab professor. You will have ample time to use your geology test kits, record data, and identify the rock specimens. A sample practicum report sheet is provided, in reduced form, below. After reviewing your practicum, your lab professor will begin discussing maps and you will embark on a new series of adventures. We will start out simple with two quiz-like geography exercises on the United States and countries in the world.

Geology 1c Rock Practicum Name ________

Sample	Igneous Metamorphic Sedimentary	Textures and/or other features	Minerals Present (one or two)	Rock Name
1				
2				
3				
4				
5				
6				
7				
8				
9				
10				
11				
12				

Figure 9.1 - Outline map of the United States for use in Exercise 9.1.

Figure 9.2 - New York Times article on failure of the superpowers in geography and map for use in exercise 9.2.

Figure 9.3 - Correct answers for the two geography exercises.

Lab 10 - Introduction to Topographic Maps

PURPOSE

This week we begin a two-session study of topographic maps. The objective of this lab is to learn how modern topographic maps are used and what the "mess of numbers and symbols" on the map sheet mean. We will discuss a variety of map-use skills, including how to read map grid systems, how to estimate distances using map scales, how to estimate bearings, and how to interpret map symbols. Read these pages carefully after discussion. Exercises are included in this lab to give you practice with latitude, longitude, the universal metric grid system (UTM), map scale, and measuring bearings.

INFORMATION FOUND ON MAPS

A **map** is a scaled two-dimensional (that is flat) representation of some part of the Earth's curved surface. Maps are designed to give the map user an accurate picture of the real world that in most cases emphasizes certain information of interest. However, no matter what features a map displays, to use a map to its fullest, all users need to be able to answer the following questions about the map:

1. What location on the surface of the Earth is represented on the map? (**Longitude** and **Latitude**)
2. How can a point on the map be specified? (**Grid Systems**)
3. How can a compass direction from one point to another be specified? (**Bearing**)
4. How are distances estimated from the map? (**Scale**)
5. What do colors, symbols, and miscellaneous information on the map mean?

LONGITUDE AND LATITUDE

The first important question a user of a map must answer is: "What part of the Earth's surface is portrayed?" In order to answer this question, one must be be able to specify location on the surface of the Earth. The location of points or areas on the surface of the Earth can be shown by means of two groups of intersecting circles known as **latitude** and **longitude** (Figure 10.1). Both latitude and longitude lines represent subdivisions of a circle and are therefore measured in **degrees**, **minutes**, and **seconds**. Remember that there are 60 minutes in a degree and 60 seconds in a minute.

Latitude lines are lines that encircle the Earth in east-west-parallel planes perpendicular to the Earth’s axis (Figure 10.1). Latitude increases in either the **north** or **south** direction moving away from the zero degree line around the middle of the Earth (the **Equator**). Thus, latitude lines increase in value north and south from **0°** at the Equator to **90°** at the Earth’s poles.

Longitude lines are lines that encircle the Earth from pole to pole in north-south-parallel planes parallel to the Earth’s axis. Longitude increases in either the **east** or

west direction away from the zero degree line called the **Prime Meridian**. Because there is no natural ‘middle’ to the Earth in a vertical-, axis-parallel orientation, the Prime Meridian is defined as the N-S circle that passes through the town of Greenwich, England (the reason for this is historical: Greenwich was the site of the British Royal Observatory and of the telescope used to make the astronomical observations on which the longitude system was originally based). Longitude lines increase in value from **0°** at the Prime Meridian to **180°** at the **International Date Line**.

Figure 10.1 - Longitude and latitude lines on the Earth.

Using longitude and latitude, any point on the surface of the Earth can be assigned a unique coordinate. For example, New York City is located approximately 41 degrees north of the equator and 74 degrees west of the Prime Meridian. This would be written:

New York City: 41°N, 74°W

However, we can be much more accurate than this using degree subdivisions of minutes and seconds. For example, the American Museum of Natural History in New York City is located at these coordinates:

A.M.N.H.: 40° 47 min. 00 sec. N, 73° 57 min. 50 sec. W

GRID SYSTEMS

Grid systems allow the map user to locate- or report on a specific point on the map. For example, longitude and latitude lines on a Mercator-projection map form a rectangular grid system that can be used to identify locations. In the United States, four kinds of grid system are common found on maps: (a) map quadrangles based on longitude and latitude; (b) the universal metric grid system or UTM; (c) the United States Land Survey; and (d) the 10,000-foot grid system. Types (a) and (b) are used extensively in the United States and we will learn how to use them in this lab. Land-survey maps (c) are used extensively in the western states and the 10,000-foot grid system (d) is used by the the New York State Department of Transportation and formerly by the U. S. Army Corps of Engineers.

LONGITUDE AND LATITUDE MAP-QUADRANGLE GRID

In the United States, four sizes of quadrangle-map grids are used. These are based on longitude and latitude measured in minutes or in degrees. The standard map-quadrangle sizes are: 7 l/2-minutes; 15 minutes; 30 minutes; and 60 minutes = 1 degree. Each map size covers an area bounded by an equal number of minutes of longitude and latitude. On a standard U. S. Geological Survey topographic quadrangle map, the information about methods of preparing the map, the map projection used (and its datum), and grid systems used appear in the lower left-hand corner of the map.

Reading Longitude and Latitude on a Topographic Map

The maps that we will be working with in lab are United States Geological Survey (USGS) 7 l/2-minute topographic quadrangle maps. On these maps the longitude and latitude coordinates are given at each corner of the map, and in thirds along the sides of the map at 2 minute, 30 second intervals (2’ 30”). Figure 10.2 shows the longitude and latitude grid common to all 7 l/2-minute topographic maps. Complete longitude and latitude coordinates are shown at the corners of the map. Coordinates are also shown at two locations, spaced 2.5 minutes apart, along each side of the map. Notice that degree values are not shown if they remain the same between corners and that seconds are usually not shown if they are 00. Notice also that each edge of the map covers 7-l/2 minutes of longitude or latitude.

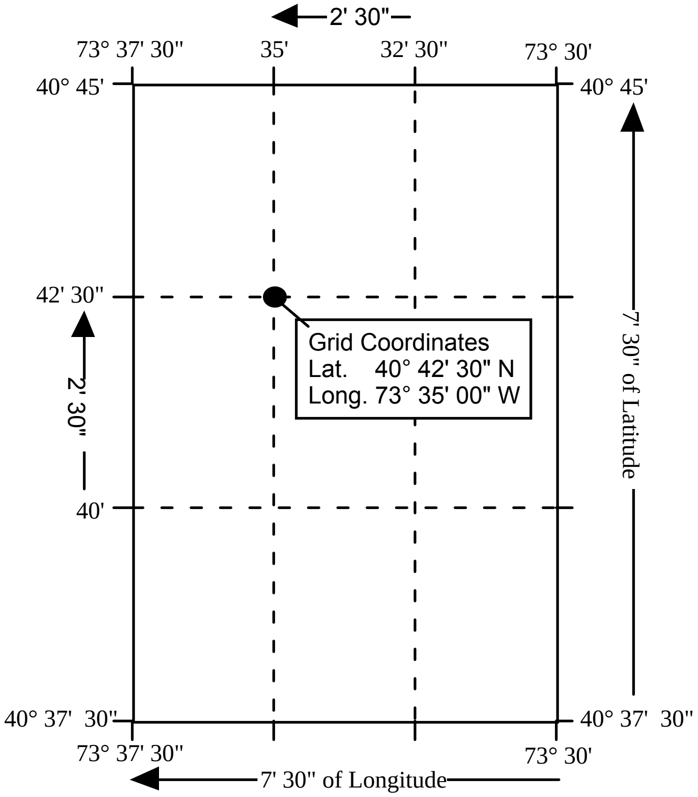


Figure 10.2 Longitude and latitude grid found on all USGS 7 l/2-minute topographic maps.

(Coordinates for Freeport, NY quadrangle shown above.)

Remember, because the United States is west of the prime meridian and north of the equator, all longitude coordinates are west (W) and all latitude coordinates are north (N). Notice that the longitude numbers increase from right to left and that the latitude numbers increase from bottom to top.

The longitude and latitude grid is very useful for locating a map on the globe and for designating quadrants on a map. However, for specifying the location of a particular point on a topographic quadrangle map, longitude and latitude can be cumbersome because the 7 l/2-minute map shows only four marked lines each of longitude and latitude. Furthermore, it is time-consuming and inaccurate to try to interpolate points that lie between the marked lines.

UNIVERSAL METRIC GRID (UTM)

An easier-to-use grid system for specifying a point on a topographic quadrangle map is the **Universal Metric Grid** or **UTM**. This grid system subdivides the map region into one-kilometer squares. Each marked UTM line on the map is exactly 1000 meters (1 kilometer) to the north or east of the last UTM line. Each UTM line has a number designation based on its distance from a reference point. One does not need to know where these reference points are to use the UTM grid. It is sufficient to specify the name of the map being used and the UTM coordinates read from the map to locate a particular point. Points that fall between the marked UTM grid lines can be accurately located by using the 1000-meter scale bar found at the bottom of the map.

The UTM grid system

As with longitude and latitude, the UTM grid system can be used to locate a map on the globe. In this respect, UTM is somewhat more complicated than longitude and latitude. If you are interested in how the UTM grid is constructed, read on. Otherwise, skip down to learn about how to read UTM coordinates on a topographic quadrangle map.

The UTM grid system divides the world into 60 grid zones that span 6˚ of longitude. Each zone that is 6˚ wide is numbered consecutively from 1 to 60, starting with Long. 180˚ W and proceeding counterclockwise, as viewed from above the Equator. From S to N, the latitudes are divided into 20 rows, each spanning 8˚. These rows are lettered, starting at 80˚S Lat. and proceeding northward. The Universal metric grid system is based on the Universal Transverse Mercator projection (hence the name 'UTM') between 80˚ N and 80˚ S, and on the Universal Polar Stereoscopic projection between 80˚ and each pole.

Within each quadrilateral, smaller areas are further desigated using a double-letter system of squares measuring 100 km on a side. Each 6˚ by 8˚ quadrilateral area contains 54 of the 100-km squares and 18 narrow marginal rectangles that measure 100 km in a N-S direction, but much less in the E-W direction, The N-S columns of squares are designated by capital letters, starting with A at Long. 180˚ W and proceeding alphabetically eastward. This sequence of letters designating large squares is repeated every 3 grid zones (every 18˚ of longitude). The E-W rows are also lettered, but in contrasting pairs of sequences. The odd-numbered grid zones begin with the letter A and proceed northward alphabetically, repeating every 20 rows. The even-numbered rows start at the Equator with the letter F and repeat after reaching V.

Further identification is numerical, based on a metric grid. The southwest corner of the C row of grid quadrilaterals (i, e,, bounded on the south by 80˚ S. longitude) is the origin for the Southern Hemisphere; and that of the N row (i. e., bounded on the S by the Equator), for the Northern Hemisphere. The Equator is assigned a value of 10,000 000 m N in the Southern Hemisphere, and zero in the Northern Hemisphere. The central meridian of each grid zone is assigned a value of 500,000 m E. Because of this assignment, the smaller values lie in the SW corner of each grid quadrilateral and the numbers increase both eastward and northward. One identifies any individual grid square by citing the coordinates of its SW corner. In such citations, one need list only four digits, two digits from the E distance, followed by two more digits for the N distance. For example, the grid square whose SW corner is 8**67**,000 m E and 3,**80**0,000 m N can be designated simply by the four digits **6780**.

Reading UTM Coordinates on a Topographic Map

UTM lines are marked on the margins of USGS topographic quadrangle maps as small, blue tick marks with numbers beside them (Figure 10.3)

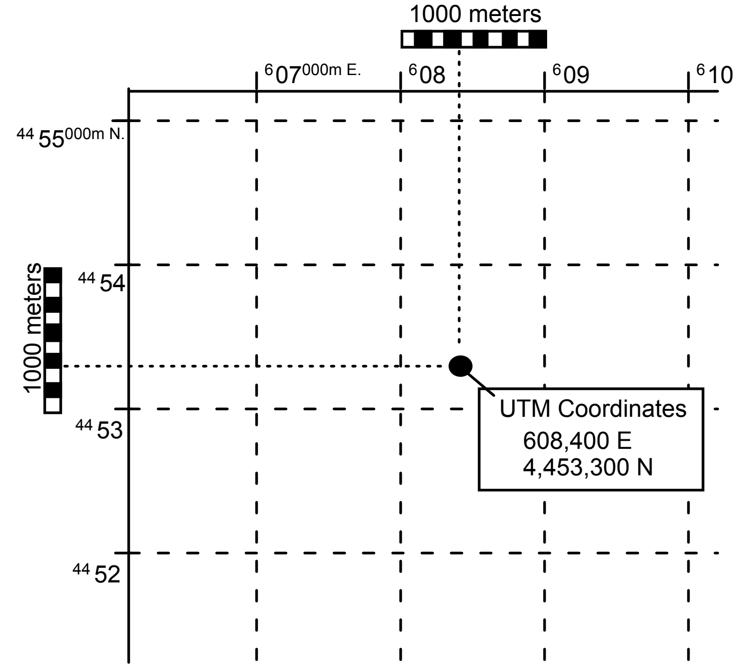


Figure 10.3 - The upper left corner of a topographic quadrangle map showing the UTM grid.

The upper-left and lower-right corners of the map show the full UTM coordinates. These numbers are read simply as numbers of meters east or north of a reference line. The digits are printed at different sizes to accentuate the thousands and ten thousands places, which change as each new grid line marks 1000 meters of distance. Notice that the ones, tenths, and hundredths places are left off of most of the UTM coordinates printed on the map.

Example: The upper left corner of Figure 10.3 is marked $^{6}07^{000m}$ E indicating that the tick mark associated with this number is 607,000 meters to the east of the reference line. Moving east, each UTM line increases by 1000 meters giving $^{6}08$, $^{6}09$, $^{6}10$, etc. The 100's of meters are left off because they are 000 at each tick. On the left side is a similar number $^{44}55^{000m}$ N indicating that this mark is 4,455,000 meters north of the reference line. These numbers count down by 1000 meters going south ($^{44}55$, $^{44}54$, $^{44}53$, etc.) indicating a decrease in distance of 1000 meters toward the reference line with each UTM mark.

The precise UTM coordinates of any point on a map can be found by noting the coordinates of the nearest intersecting blue tick marks and then by using the kilometer scale at the bottom of the map to measure the number of meters away to the east and north the point is from the nearest tick marks (Figure 10.3).

OTHER GRID SYSTEMS

U. S. Land Survey Grid System

Much of the land surveying in the central- and western states is based on a scheme established by Congress in 1785 when the Northwest Territories were settled. The fundamental grid unit is a square, 6 miles on a side, that has become known as the "Congressional township." These townships are parts of a larger system of surveyed lines. The N-S lines are the **principal meridians**; and the E-W lines, **base lines**. The first principal meridian is the Ohio-Pennsylvania border. Other principal meridians west of this one are numbered from 1 through 6. After 6, they are named by geographic features. The base lines are taken from given latitudes.

The 10,000-foot State Grid System

On many of the 7 I/2-minute quadrangle maps, a 10,000-foot grid system is marked. Each of these grids is based on an individual state. The grid is used just the opposite as in the universal metric system. Instead of "reading right (to the East) up (to the North)," one uses the order "northing" and "easting" (i. e., give the NS coordinate first and the EW coordinate second). This 10,000-foot grid system was used for many engineering purposes before the universal metric grid system had been widely surveyed. At present, the trend is to abandon the 10,000-foot grid in favor of the univeral metric grid.

Exercise 10.1 Map coordinate systems

BEARING AND MAGNETIC DECLINATION

A **bearing** is a compass direction from one point to another on a map or on the surface of the Earth. Bearings are defined by the angle between a line pointing north and the line connecting the two points. In the field, a bearing is measured using a compass, which is an instrument capable of measuring the angle between a sighted bearing (the direction the compass is pointed) and the direction to the Earth's magnetic pole. On a map, bearings are measured using a protractor, with north generally defined as the direction pointing straight to the top of the map.

The Compass Rose

The compass rose is the familiar north, east, south, west cross used to show direction on a map (Figure 10.4). There are two methods of designating a bearing, depending on which style of compass rose is used. The **Azimuth** method is based on a 360° circle. A bearing is reported as the angle between the bearing line and 0°, measured clockwise around the compass rose. The **Quadrant** method is based on a division of the compass rose into four quadrants. Bearings are read as the angle between north or south and the bearing line in either the east or west direction. For example, in Figure 10.4 a bearing line midway through the NW quadrant of the compass rose can be read as 315° (start at north, turn 315° clockwise - [azimuth method]) or as N 45° W (start at north, turn 45° to the west - [quadrant method].)

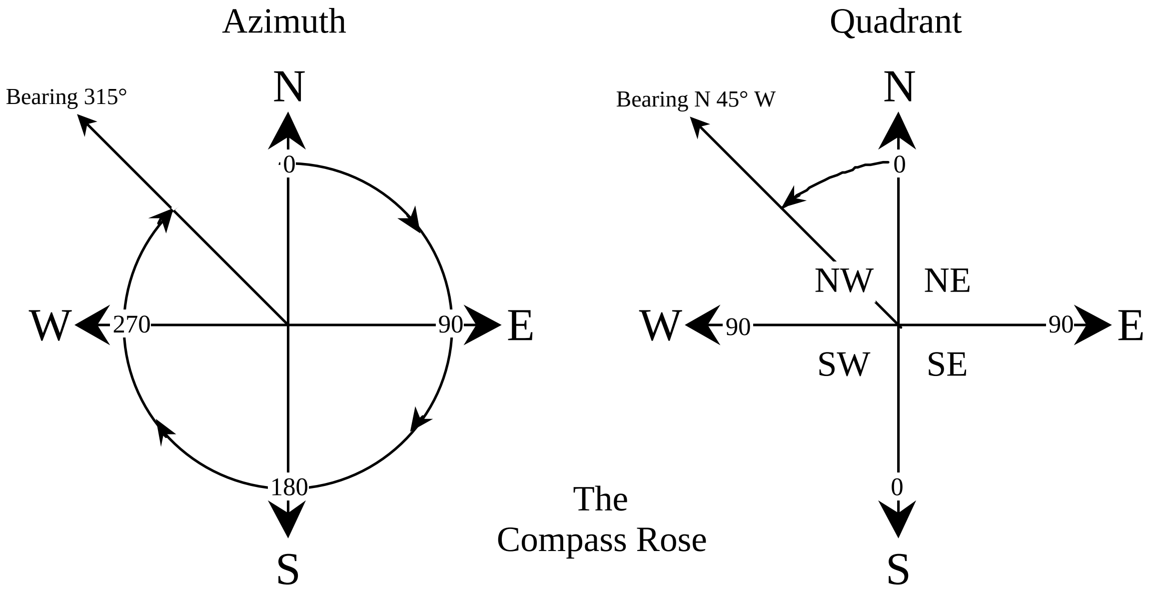


Figure 10.4 - The Compass Rose showing both azimuth and quadrant bearings.

In general, the quadrant method of reporting bearings is easiest to use. One advantage is seen in coverting from a bearing to its reverse in the opposite direction. For example, if a bearing from point A to point B is given as N 55° W, the reverse bearing from B back to A is S 55° E; one has only to reverse the compass directions while keeping the same angle. Azimuth bearings are advantageous if one needs to process them using a computer because each bearing can be represented by a single number.

Magnetic North and True North

North is shown along the lower margin of a standard topographic map by three arrows whose tips are marked MN, H, and GN (Figure 10.5). These refer to Magnetic North, True North, and Grid North, respectively. By convention, True North is toward the top of the map; it is defined by the meridians of longitude. Magnetic North is the direction toward which a compass needle points within the map area. The angle between Magnetic North and True North is known as the **magnetic declination**. Because the magnetic pole shifts westward with time, the declination needs to be monitored and updated for accurate navigation. Grid North shows the deviation between the rectangular grids overlayed on the map and true geographic north. This deviation occurs because a map is a flat representation of the Earth's curved surface.

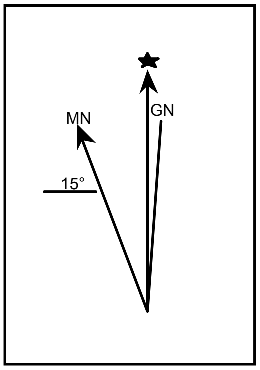


Figure 10.5

True North is different from Magnetic North because the Earth's geographic north pole and its north magnetic pole are not located at the same place (Figure 10.6). The apparent angle between them (magnetic declination) changes as you move to different places on the globe. All USGS topographic quadrangle maps show the amount and direction (east or west of north) of magnetic declination for a given map area. For compass bearings to agree with bearings measured from a map, the compass must be set to compensate for magnetic declination. Otherwise, map and compass bearings are being measured from different 0° (north) lines. One can convert from a True North map bearing to a magnetic north compass bearing, or vica-versa, by either adding or subtracting the magnetic declination value from the true north value (Figure 10.7).

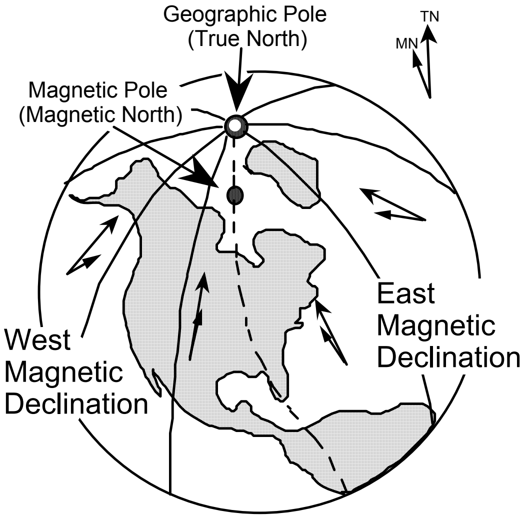


Figure 10.6

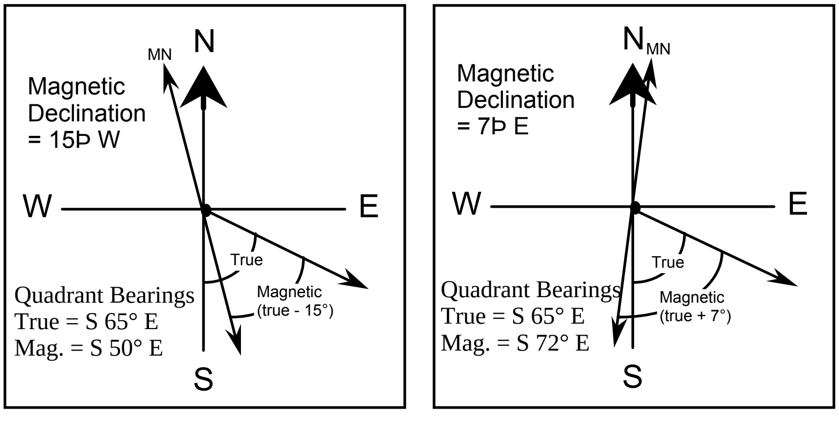
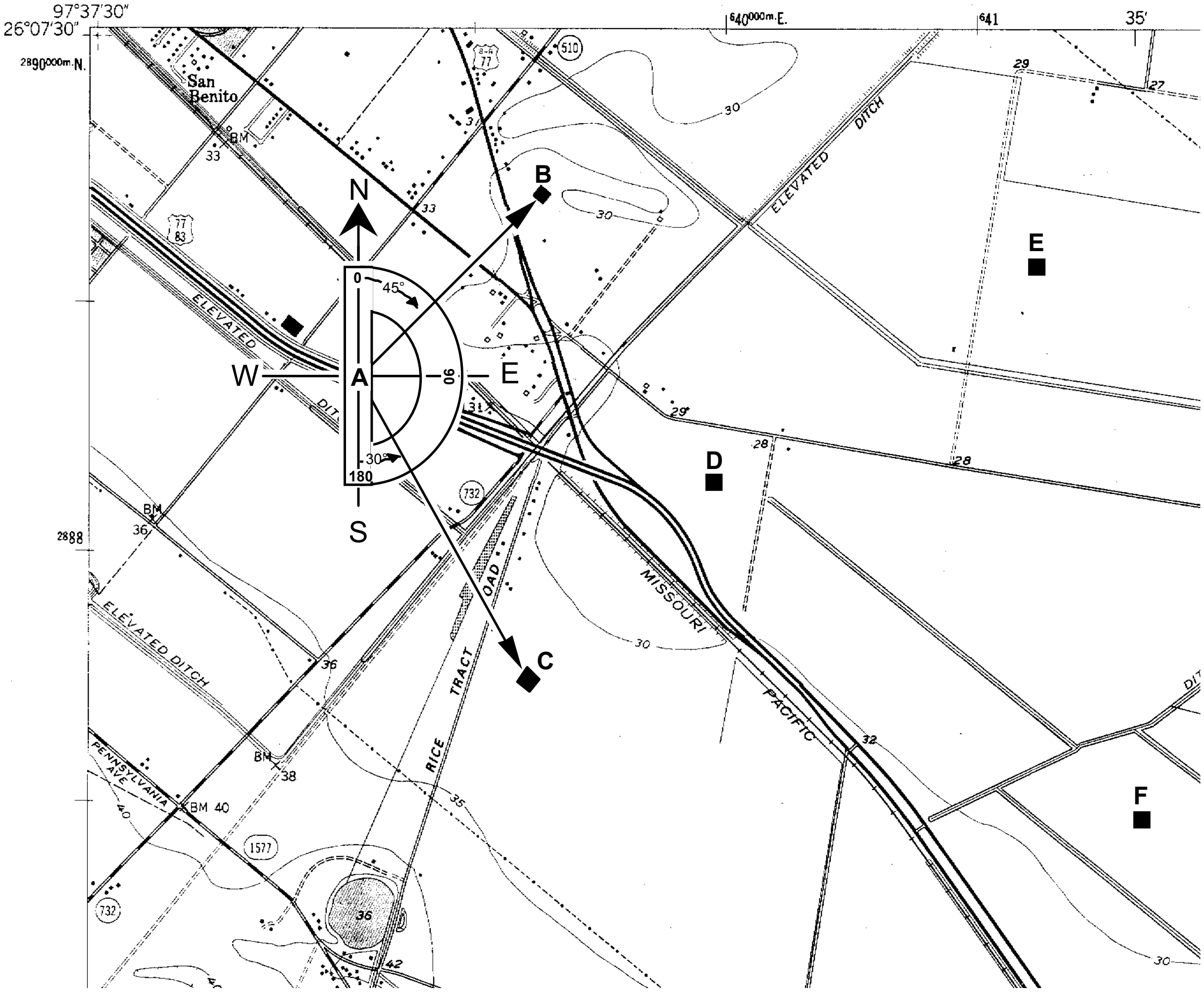


Figure 10.7 Converting between True North and Magnetic North.

Laboratory Exercise 10.2 - Measuring Bearings



Instructions: To measure a bearing between two points on a map, center the protractor on the map at the first point with the zero line (axis) of the protractor facing north (straight up the map) or south (straight down the map) . With the protractor correctly oriented, read the angle between north or south and the bearing line to the second point. It is usually better to ignore the numbers marked on the protractor and simply use the protractor scale to count degrees away from the axis toward the line you are measuring.

Example (shown above):

Bearing A - B : **N 45° E** (quadrant) or **045°** (azimuth)
Bearing A - C : **S 30° E** (quadrant) or **150°** (azimuth)

Draw and then measure the following bearings using a protractor:

D - E : ______________________
D - F : ______________________
C - D : ______________________

D - B: ______________________
D - C : ______________________
B - D : ______________________

MAP SCALES AND MEASUREMENT OF DISTANCE

A **map scale** is a **ratio** that expresses the relationship between distances on the map and corresponding distances in the real world, in the same units, whatever these may be. The standard map scale for topographic quadrangle maps in the United States is 1/24,000. This means that one unit on the map (centimeter, inch, shoe length, etc.) is equivalent to 24,000 of those units (centimeter, inch, shoe length, etc.) in the real world. It does not matter what units you use to make your measurement from the map, in the real world the equivalent distance will be 24,000 times the map distance on a standard topographic map.

If you measure a distance on a 1:24,000 scale map to be two inches, the corresponding actual distance is 48,000 inches. However, this is not a very useful way to report distance because very few people have any intuitive sense of how far 48,000 inches is. It is much more useful to convert 48,000 inches into the equivalent number of feet (48,000 inches / 12 inches per foot = 4000 feet) or miles (4000 feet / 5280 feet per mile = .76 miles). Below are the commonly used conversion factors for converting between inches and miles or between centimeters and kilometers. For practice making unit conversions and ratio scale calculations, see **Laboratory Exercise 10.3**.

Centimeters, meters, and kilometers:
100 centimeters = 1 meter
1,000 meters = 1 kilometer
100,000 centimeters = 1 kilometer

Inches and miles:
12 inches = 1 foot
5,280 feet = 1 mile
63,360 inches = 1 mile (5,280 feet x 12 inches/foot)

When reporting a scale, the <u>units of distance must be specified if they are not the same between the map and the real world</u>. If they are the same then the scale can be expressed as a unitless fraction.

For example:
If 1 inch on the map equals 2000 feet on the ground, the scale is **1 inch : 2000 feet** <u>or</u>, because 2000 feet x 12 inches = 24,000 inches, **1/24,000**. If 1 inch on the map equals 1 mile on the ground, the scale is 1 inch/5280 feet x 12 inches/foot or **1/63,360**.

The ratio expressing the scale of the map is named the **representative fraction**. This is sometimes abbreviated as **RF**.

Some common map scales are listed below:

U. S. Topographic map series	**RF**
7.5-minute	1/24,000
7.5-minute	1/31,680
15-minute	1/62,500

30-minute	1/125,000
l-degree	1/250,000
4-degree	1/1,000,000

On each map will be found at the bottom center, and possibly also along one of the sides, a **graphic scale**, in which actual distances along the map are shown with their equivalent real-world units, such as thousands of feet, miles, or kilometers. Such a graphic scale (also known as a **bar scale**) is very useful because it will change size if the map is photocopied and either reduced or enlarged. The RF is valid only on the size of the map as originally printed.

Miscellaneous information shown on maps

In the lower-right corner of a standard topographic map sheet below the actual map is an index map of the state showing the location of this quadrangle within it. In the extreme lower-right corner is found the name of the quadrangle. Beneath the name are the latitude of the S border (numbers of degrees and minutes run together and use decimal subdivisions of minutes of latitude) and the range of longitude from the E border to the W border. Below that is the date when the aerial photographs were flown that were used to compile the map (with dates of revision indicated in purple).

What the colors represent

On modern topographic maps, the 8 basic colors are:

A. White - most undeveloped areas lacking a tree cover (fields, parks, etc.)
B. Green - tree-covered areas (special patterns for orchards)
C. Blue - water features or marshes
D. Red - major roadways
E. Black - cultural features (schools, roads, railroads, place names
F. Pink - urbanized areas (high concentrations of homes and buildings)
G. Brown - contour lines; used to show relief (changes in elevation) to give you a feeling for "depth".
H. Purple - used to show revisions in cultural features or changes in contours after original map was made.

Map Symbols

The standard map symbols are shown on Figure 10.8.

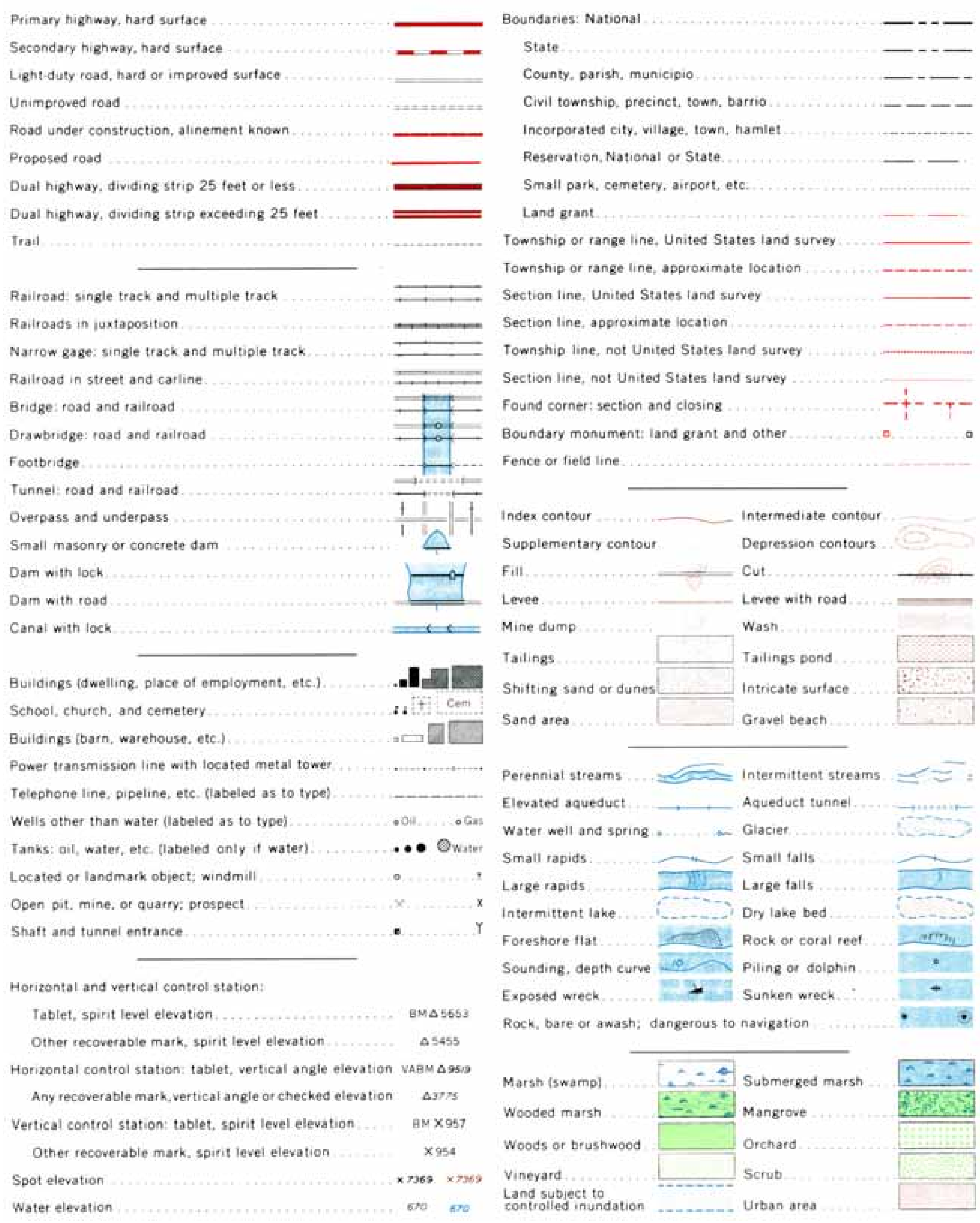


Figure 10.8 - Topographic map symbols. (United States Geological Survey.)

Laboratory Exercise 10.3
Calculating Map Distance Using a Ratio Scale

A. Unit Conversions (use a calculator, but write out all calculations neatly.)

1 mile (mi) = 5280 feet (ft)
1 foot (ft) = 12 inches (in)
1 kilometer (km) = 1000 meters (m)
1 meter (m) = 100 centimeters (cm)

1. How many inches are there in one mile?

2. How many centimeters are in a kilometer?

3. 278,000 in = how many ft?

4. 3,900,000 in = how many mi?

5. 4.2 mi = how many in?

6. 5.5 mi = how many ft?

7. 4,300,200 cm = how many km?

8. 42 km = how many cm?

B. Calculating distance using a Ratio Scale (use a calculator, but write out all calculations neatly.)

9. Given a scale of 1:24,000
How many mi are represented by 3.2 in on the map?

10. Given a scale of 1:62,500
How many mi are represented by 1.5 in on the map?

11. Given a scale of 1:120,000
How many km are represented by 13 cm on the map?

12. Given a scale of 1:62,500
How many km are represented by 20 cm on the map?

Laboratory Exercise 10.4

Reading Topographic Quadrangle Maps

Questions based on the USGS Freeport, N. Y.
7 1/2-minute quadrangle map

General questions:

1. What type of surface area on the map is indicated by pink? _____________

   by green? ________________ by white? ____________________

2. What is the most common type of building identified by name on this map?

   ____

   What symbol is used to indicate it? _______________

3. What other kinds of buildings are marked?

   ____

   Why are these buildings shown on the map?

   ____

4. A variety of information is found at the bottom of a topographic map. Can you find...

   Name of the quadrangle
   Date
   Date of photorevisions
   Location of the quadrangle in the state
   General road classification
   Type of map projection used
   Grid systems shown
   UTM Zone number

On the map itself, can you find...

Hofstra University
Gittleson Hall
Nassau Coliseum
Meadowbrook Parkway
Southern State Parkway
Sunrise Highway

Magnetic Declination and Bearings

5. What is the magnetic declination in the map area? _______________

6. Find the M8 interchange of the Meadowbrook Parkway near Freeport. What is the **true north bearing** from the **M8** interchange to the **W5** interchange to the east along the Sunrise Highway? _____________

7. Find the Nassau Coliseum. It is the large, oval building colored in purple and located to the northeast of Hofstra. What is the true north bearing from the Coliseum to Gittleson Hall?_______

8. What is the true north bearing from Gittleson Hall to the Coliseum? _______

Converting bearings from true to magnetic

(These are easier to figure out if you draw a small diagram for each showing the position of the magnetic north-south axis relative to the true north-south axis.)

9. If a magnetic bearing is N 67° E, what is the true bearing? _______

10. If a magnetic bearing is N 59° W, what is the true bearing? ______

11. If a true bearing is S 60° E, what is the magnetic bearing? _______

12. If a true bearing is S 25° W, what is the magnetic bearing? _______

13. What is the magnetic bearing from Gittleson Hall to the Nassau Coliseum? _______

Distance and Scale

14. How many feet are represented by one inch on the map? ________

15. How many meters are represented by one centimeter on the map? ________

16. How many inches on the map equal one mile? ________

17. How many centimeters on the map equal one kilometer? ________

18. What is the straight-line distance from Gittleson Hall to the Nassau Coliseum?

Miles ________ Kilometers ________

Using the ratio scale

19. Using the ratio scale, estimate the distance in both miles and kilometers from **Gittleson Hall** to the center of the **Roosevelt Raceway** track along the true bearing **N 5° E**. Show all calculations below.

Miles ________ Kilometers ________

20. Why is it easiest to use the metric system to make distance calculations?

Finding Location Using Coordinate Systems

21. How many degrees of latitude and and how many of longitude are encompassed by the map? ____

22. Why isn't this map square? (Try to answer this question on your own; if you are stumped, your lab professor will explain.)

____

____

____

____

23. Let’s call in an air strike on the sewage disposal plant south of Seaford, near the bottom right corner of the map. What **latitude** and **longitude** coordinates do we give the pilot? Estimate as accurately as you can.

Lat. ______________________ Long. ______________________

24. What is the name of the school located at UTM **623,000 E**., **4,508,000 N**. ?

____

25. As accurately as possible, determine the UTM coordinates for **Gittleson Hall** and **Axinn Library**.

Gittleson: E_____________N_____________

Axinn: E_____________N_____________

Lab 11 - Topographic Contour Maps and Profiles

PURPOSE

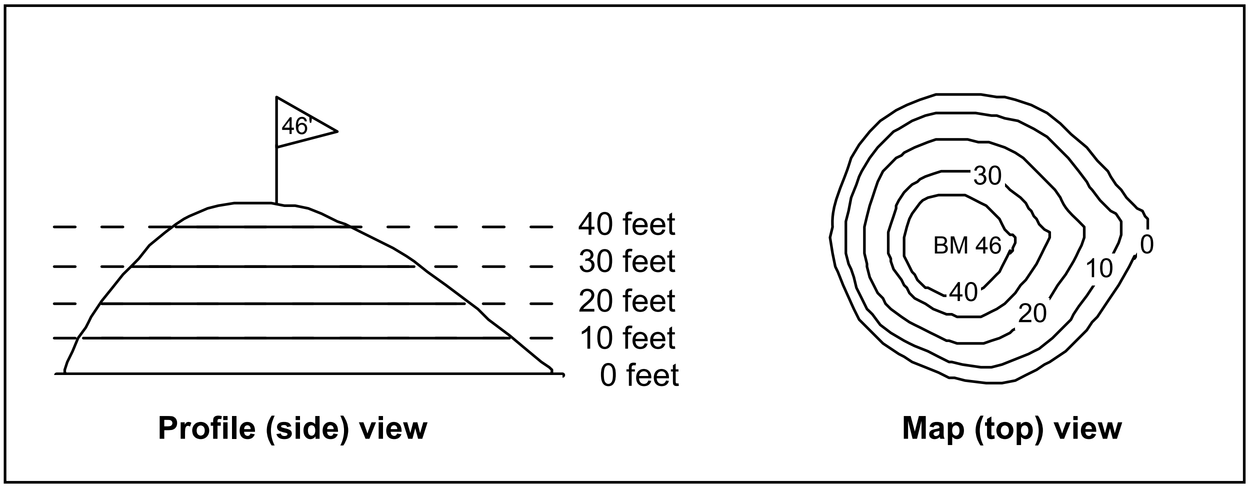
Today's laboratory is intended to acquaint you with topographic contour maps and topographic profiles. The basics of topographic maps, map scales, map grids, and symbols were covered in last week's lab. This week, we will focus on reading and interpreting these maps for information on **elevation**, **gradient**, and **landscape profile**. Remember that a **MAP PRACTICUM** will be given during the last lab meeting (Lab 13) of the semester. It will consist of questions similar to the exercises in Labs 10, and 11.

CONTOUR MAPS

Topographic contour maps are maps that show the changes in elevation throughout the map area using lines of constant elevation called **contour lines**. By means of contour lines the three-dimensional "lay of the land" can be illustrated in two dimensions on a printed map. Standard USGS Topographic Quadrangle Maps include contour lines.

Contour Lines

Imagine, if you will, a small hill in the middle of a field. If we could get our hands on one of those chalk carts that are used to put lines on athletic fields, then we could use the cart to draw contour lines on the hill. We would do this by starting at the base of the hill and pushing the cart around the base, following a level line, but staying with the edge of the slope at the base of the hill. Then we would measure 10 feet of vertical distance (altitude or **elevation)** and move the cart to a point on the hill ten feet above the level ground. Starting from this point we would push the cart around the hill, never moving up or down the hill, but always staying exactly ten vertical feet above the level ground. We would go around the hill and eventually come back to where we started, having drawn a 10-foot contour line. Now, moving another ten feet up, we would do this again. Another ten feet after that, and again. If we keep making lines around the hill, moving up ten feet every time, eventually we will reach the top of the hill. Chances are that the actual summit of the hill would be a little above the last line we made, but below the next ten-foot interval.



Profile (side) view

Map (top) view

Figure 11.1 - Contour lines drawn on an imaginary hill viewed from the side and the top.

If we had a sensitive altimeter, we could measure the height of the top of the hill. Let's say that we make four lines above the base of the hill (that's five lines total, counting the one around the base) and that we measure the highest point at the top of the hill to be 46 feet high, marking it with a flag.

Now, if we flew over the hill in a helicopter and photographed it from above, the lines we have marked on the hill (side view, Figure 11.1) would appear as a series of concentric circles (top view, Figure 11.1). We could then mark each circle in the photograph with the height that it represents, and also mark the summit of the hill with the height that we measured.

Having done this, we have constructed a **topographic contour map**. Each line is called a **contour line**, and the change in height from one line to the next (10 feet in this case) is called the **contour interval**. The measured height marked at the top of the hill we call a **bench mark (BM)**. The base of the hill where we began measuring elevation is our zero elevation point or **datum**. In most cases, datum on topographic maps is defined as **mean sea level**.

Contour lines also indicate the **slope** of the Earth's surface. Where contour lines are closely spaced, slopes are steep. Where the contour lines are spaced widely apart, slopes are gentle. All of the land on one side of a contour is higher than the land on the other side of the contour. Therefore, when you cross a contour, you either go uphill or downhill. The basic determination in reading any contour map is to figure out the direction of slope of the land. Careful examination of stream-flow directions and bench marks will give you a general feeling for the overall slope of the Earth's surface in any given map area. Some general rules for contour lines are given below:

The Rules of Contour Lines

When reading contour maps, or when trying to determine what the elevations of contour lines are, one must apply a few basic rules of contour lines. There are no exceptions to these rules!

1. Closed countour lines on a map indicate either a hill (peak, mountain, etc.) or a hole (depression, etc.). Closed contours that indicate that the land slopes down into a hole are marked by hachured lines to distinguish them from closed contours that indicate that the land slopes up over a hill (Figure 11.2).

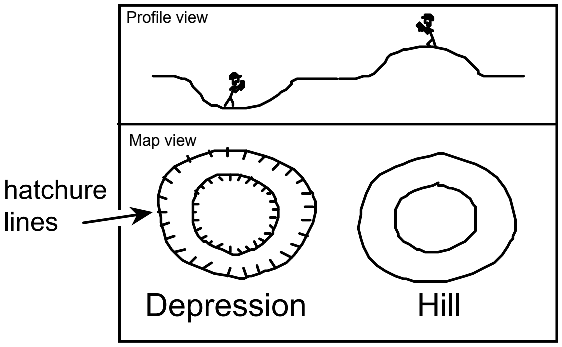


Figure 11.2

2. A **single contour line** represents a **single elevation** along its entire length. In other words, the elevations of all points along a contour line are the same.

3. Contour lines **<u>never</u>** split, cross, or intersect. At a vertical cliff they do, however, come together and touch (Figure 11.3).

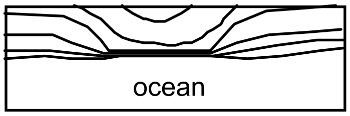


Figure 11.3

4. The elevation of a contour line is always a simple multiple of the contour interval. For ease of reading, by convention, each fifth consecutive contour line is an **index contour** (drawn as a thicker line than adjacent contours and also numbered somewhere along the trace of the contour line). Commonly used intervals are 5, 10, 20, 40, and 80 feet.

5. Widely spaced contour lines indicate a gentle slope. Closely spaced contours indicate a steep slope.

6. Every contour line eventually closes on itself. However, any one map will not be large enough to show the full extent of all contour lines, and some will simply end at the edge of the map. Where one closed contour line surrounds another, the inner contour line marks the higher elevation. If the contour lines are hachured, then the inner contour line marks the lower elevation.

7. Where a contour line crosses a stream or a valley, the contour bends to form a 'V' that points upstream or up the valley (Figure 11.4).

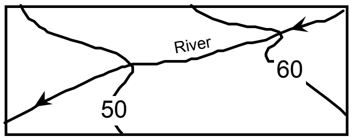


Figure 11.4

8. Where two adjacent closed contours indicate opposite slopes (hachured contour next to a plain contour) both are the same elevation (Figure 11.5).

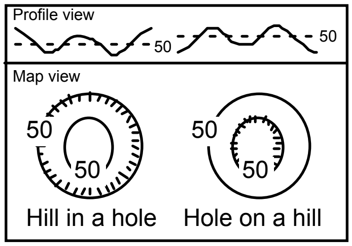


Figure 11.5

9. A hachured contour line, lying between two different contour lines, is the same elevation as the lower contour line (Figure 11.6).

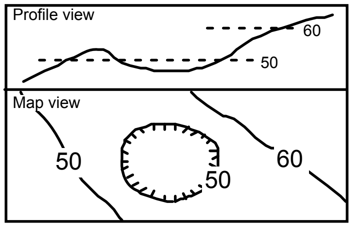


Figure 11.6

10. A closed contour line, lying between two different contour lines, is the same elevation as the higher contour line (Figure 11.7).

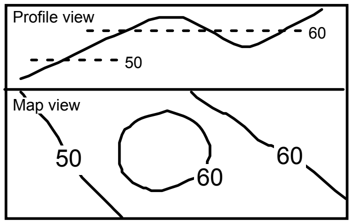
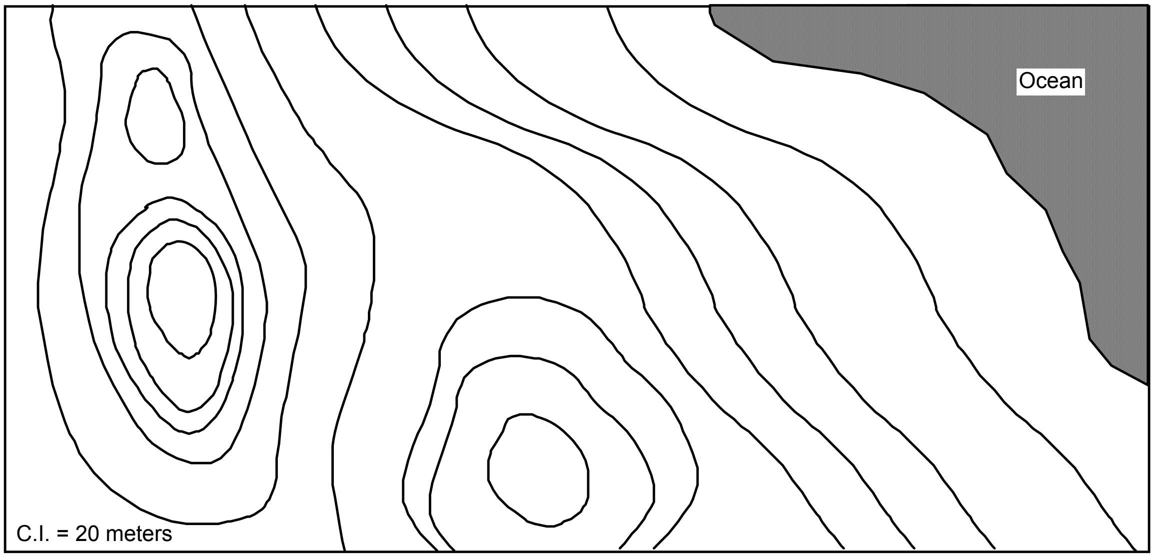


Figure 11.7

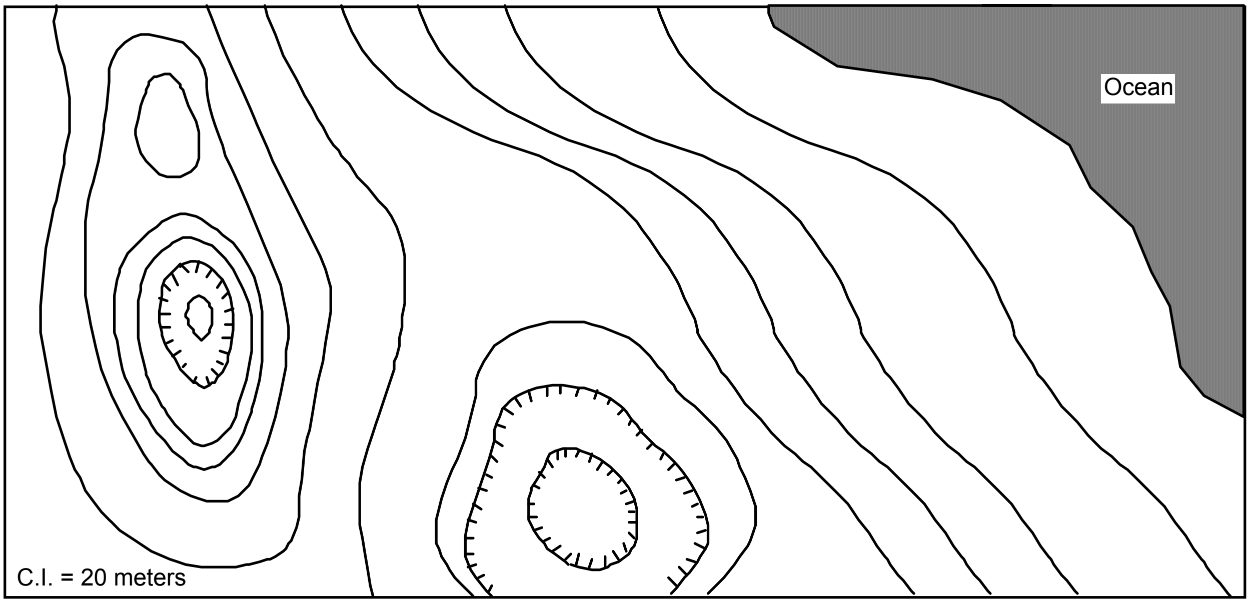
11. Finally, "Obey all the rules"!

LABORATORY EXERCISE 11.1
Determining topographic contour values

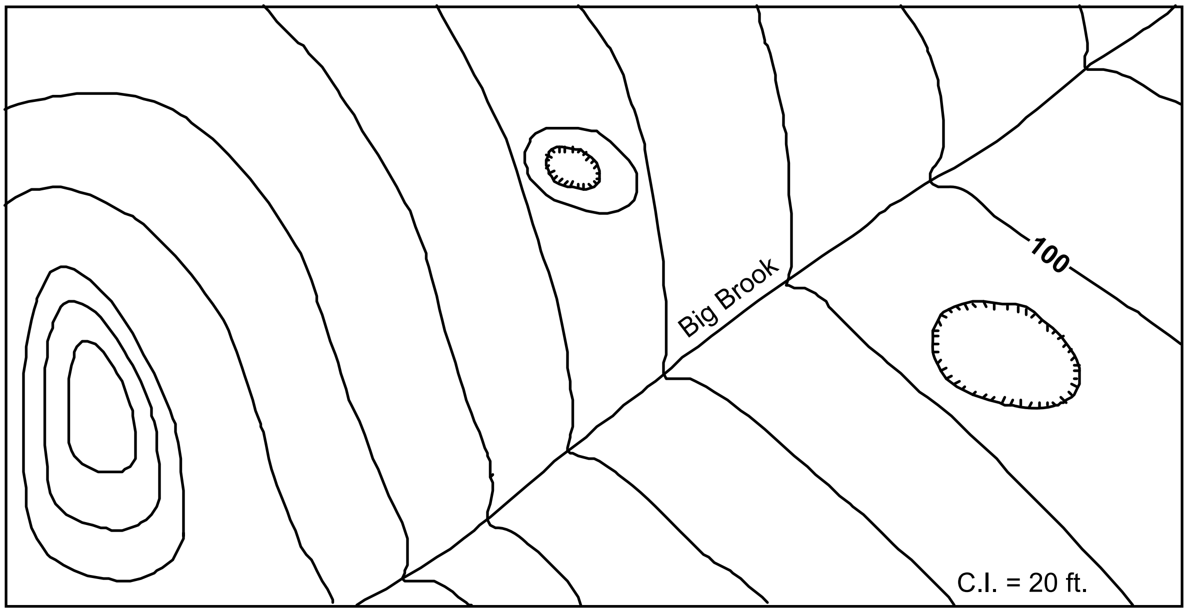
Using the rules of topographic contours listed above, label all of the topographic contour lines in the following maps (11.1a - 11.1d) with their correct elevations. Zero elevation is sea-level (shore line). Note the **contour interval** (C.I.) given on each map.



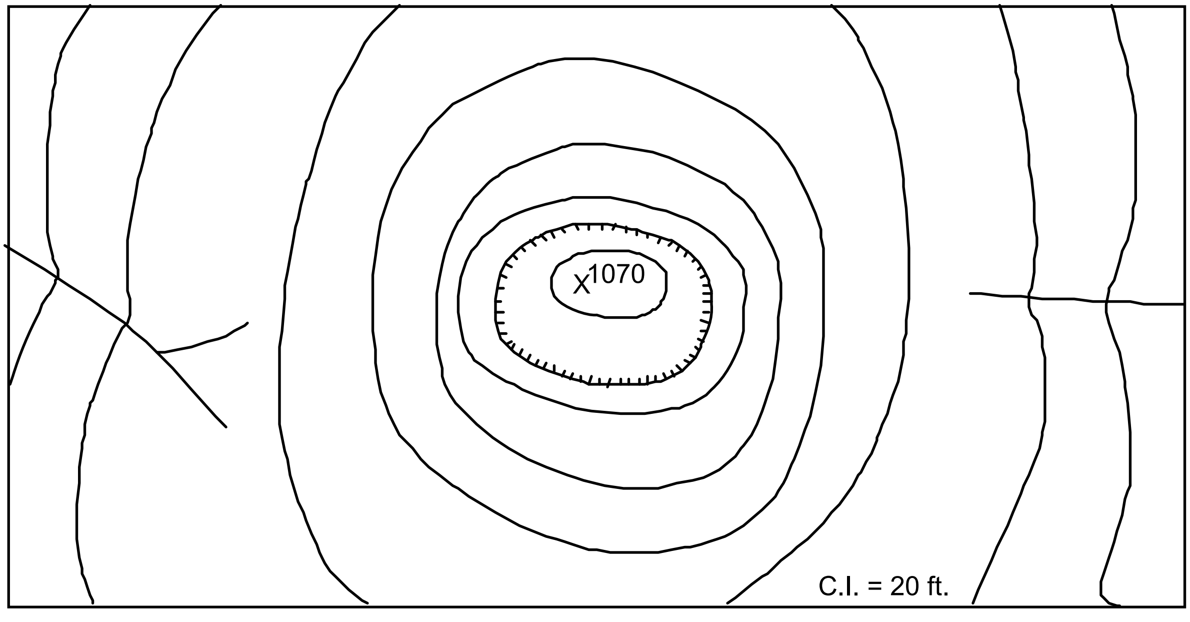
Exercise 11.1a



Exercise 11.1b

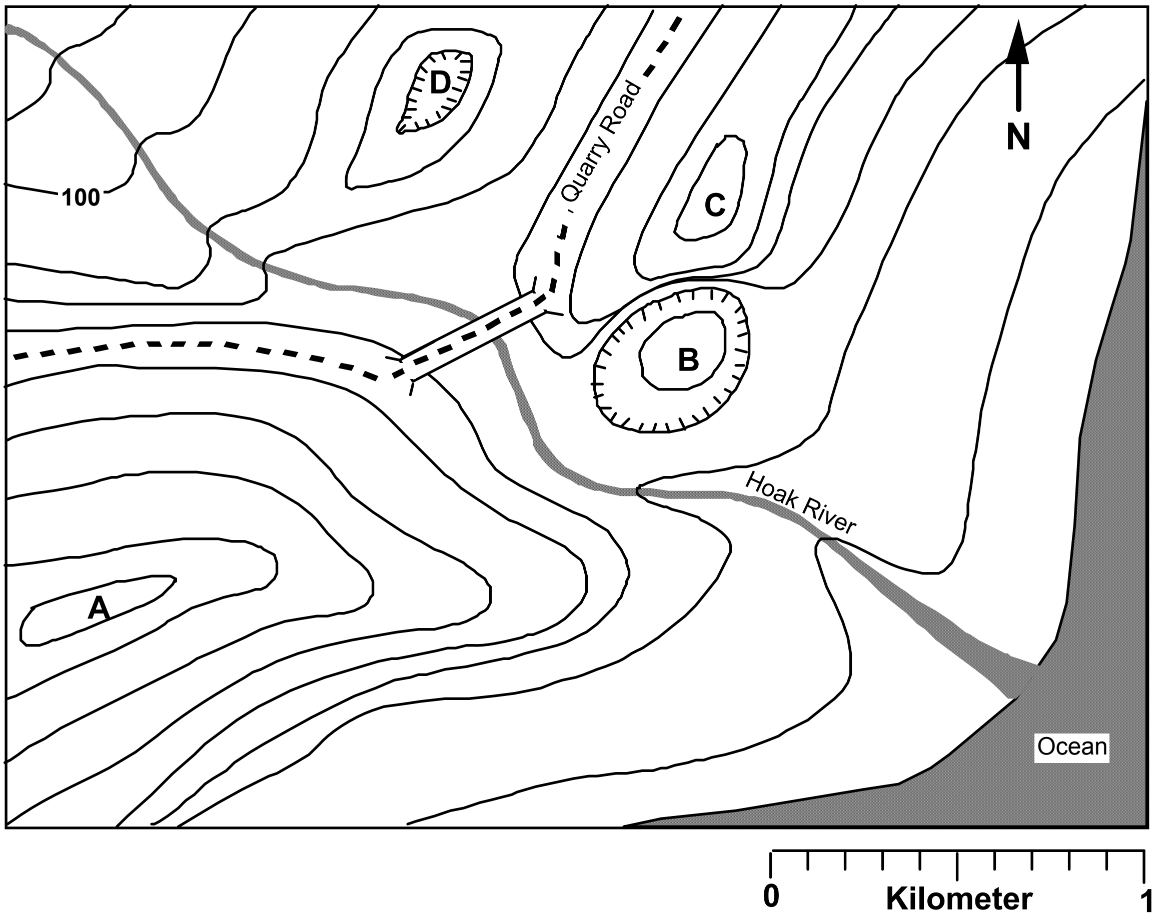


Exercise 11.1c



Exercise 11.1d

LABORATORY EXERCISE 11.2
Interpreting topographic contour maps



Using your knowledge of topographic contour maps, answer the questions based on the above map.

1. What is the contour interval?______________________ meters

2. What is the maximum elevation of point A?________________

3. What is the maximum elevation of point B?________________

4. What is the maximum elevation of point C?________________

5. What is the minimum elevation of point D?________________

6. Estimate the height of the Quarry Road Bridge above the Hoak River.
____

7. Which side of Ridge A is steeper?_______________

8. In what direction does the Hoak River flow?_____________

9. Explain three different methods that you can use to determine which way the Hoak River flows:

a.____

____

b.____

____

c.____

____

10. What is the total length of Quarry Road shown on the map?__________

11. Does Quarry Road go uphill, downhill, or remain level from west to east?

LABORATORY EXERCISE 11.3
Contouring a map from spot elevations

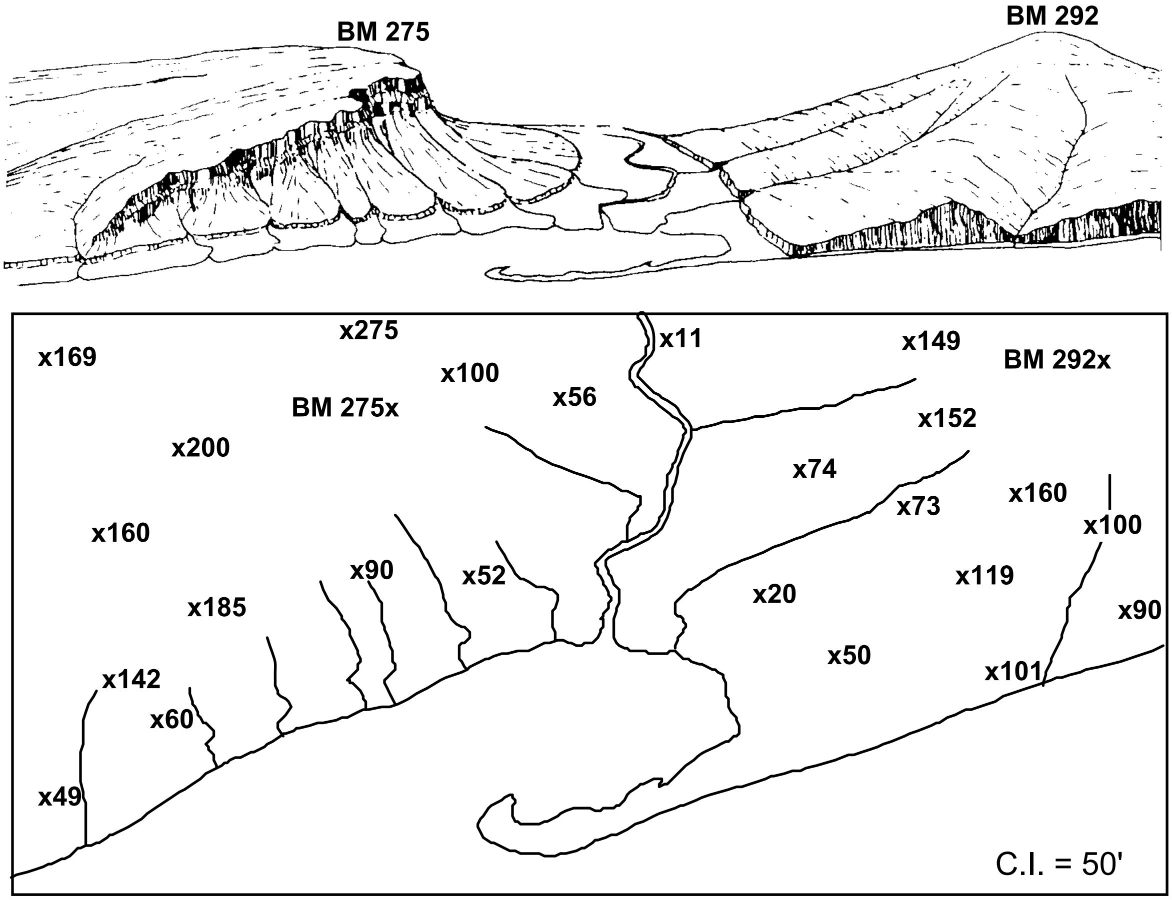
The diagram below includes an oblique view of a coastal terrain and a base map with spot elevations as measured by a surveying team. In order to complete the contour map, follow the simple directions below. Be sure to use a pencil with a new eraser and try not to get discouraged. Use the oblique view as a guide to drawing the contour lines.

1) For this map, use a contour interval of 50 feet.

2) Each contour line, in maintaining an equal elevation, **must** pass above points of lower elevation and below points of higher elevation. The spacing between the contour line and these points depends on their difference in elevation and is determined by the process of **interpolation**.

3) Make all lines light and round all curves (no sharp angles except around streams). Remember the rule of the "V's".

4) Draw the contour lines beginning at the lowest elevations (streams). Refer to rules above.



Exercise 11.3 - Spot elevation map and oblique sketch of a region made famous by the United States Geological Survey. You must create a contour map.

Topographic Profiles

A topographic profile is a vertical ‘slice’ through the landscape constructed along a straight **line of profile** drawn across a topographic map. A topographic profile shows changes in **relief** (change in elevation) in the vertical dimension as a silhouette. We construct topographic profiles to get a ground-level view of the lay of the land (Figure 11.8).

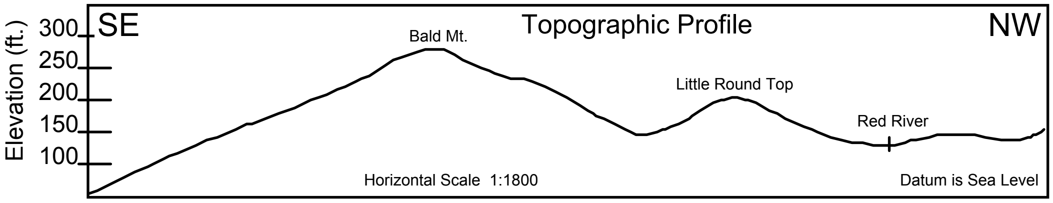


Figure 11.8 Example of a topographic profile.

A well-constructed topographic profile must also include the following information:

- A description of the line of profile, including the name of the map quadrangle from which the profile was derived and the compass directions at each end of the profile (i.e., SE and NW).
- Above the profile, letter in the names of prominent geographic features, such as rivers or mountains. The space below the profile line may be needed for showing geologic structure. Therefore, even though the profile line is intended to show the configuration of the land surface only, form the habit of keeping clear the space below the profile line.
- All profiles should be labeled with a horizontal scale (taken from the map), vertical scale (determined by the maker of the profile), a vertical exaggeration (see explanation to follow), and the datum used to control the vertical scale (usually sea level).

Instructions for drawing a topographic profile
(See figure 11.9)

A.. Select the line of profile on the map.

B. Place the edge of a blank strip of scrap paper along the line of profile. You may want to tape down the ends of this strip of paper so it will not move while you are working with it.

Using a sharp pencil point, make precise tick marks at places where contours and other features on the map (streams, tops of hills, etc.) intersect the edge of the paper strip. Label the tick marks with the elevation values of the contours.

C. Select a vertical scale. On a piece of graph paper, draw in a segment of the vertical scale having a range large enough to include the elevations all the points (= contours) to be plotted. Label the elevation of the datum (zero line) of the cross section (usually sea level).

Realign the edge of the strip that formerly lay along the line of profile so that it now lies along the bottom line of the graph paper.

Plot a point for each tick mark at the appropriate elevation directly above each contour tick mark. Connect the plotted points with a line that curves with the shape of the land (i. e., rounded hills, V-shaped stream valleys, etc.).

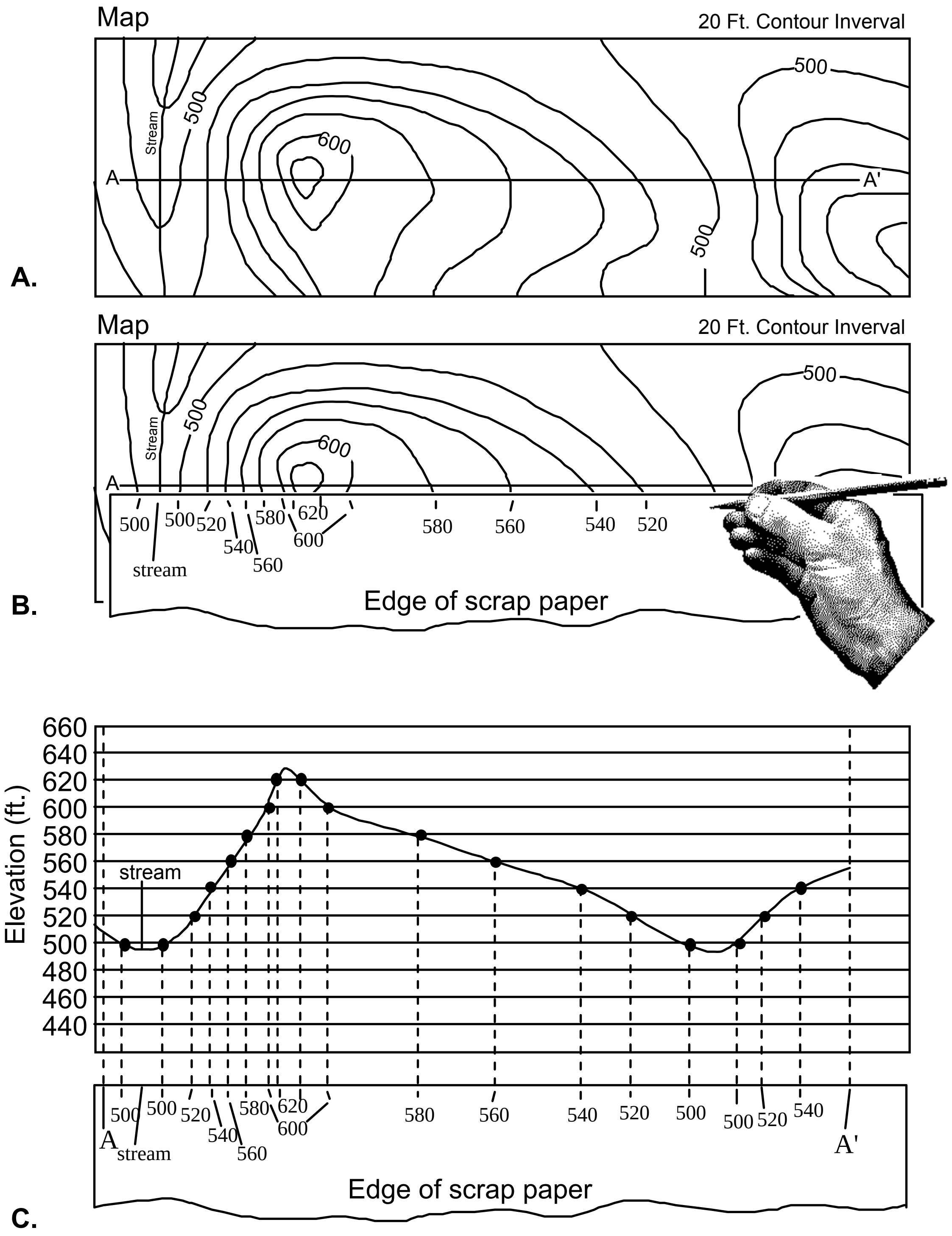


Figure 11.9 Diagrams illustrating the construction of a topographic profile.

Vertical Exaggeration (V.E.)

Vertical exaggeration is the ratio of the vertical scale of a topographic profile to the horizontal scale of the profile. One uses vertical exaggeration to show the shape of the land where the relief is so small that it does not stand out when the vertical scale equals the horizontal scale. In effect, what one does is to "stretch" the landscape according to the ratio (i. e. vertical exaggeration) selected. The choice of vertical exaggeration will vary depending on the relief of the area, the scale of the map, and the purpose of the profile.

If the vertical scale is larger than the horizontal scale (as is often the case), then the topographic profile will be ‘stretched’ in the vertical direction and steepnesses will be exaggerated. This has the effect of making peaks and valley seem larger and deeper than they really are.

A vertical exaggeration of 1 means that the horizontal scale and the vertical scale are equal so that the topographic profile is a realistic depiction of the actual shape of the features represented and, in effect, there is no vertical exaggeration.

To calculate V.E., first determine the fraction value of both the vertical- and horizontal scale (for example, 1:24,000 = 1/24,000). Dividing the vertical scale fraction by the horizontal scale fraction will give the vertical exaggeration.

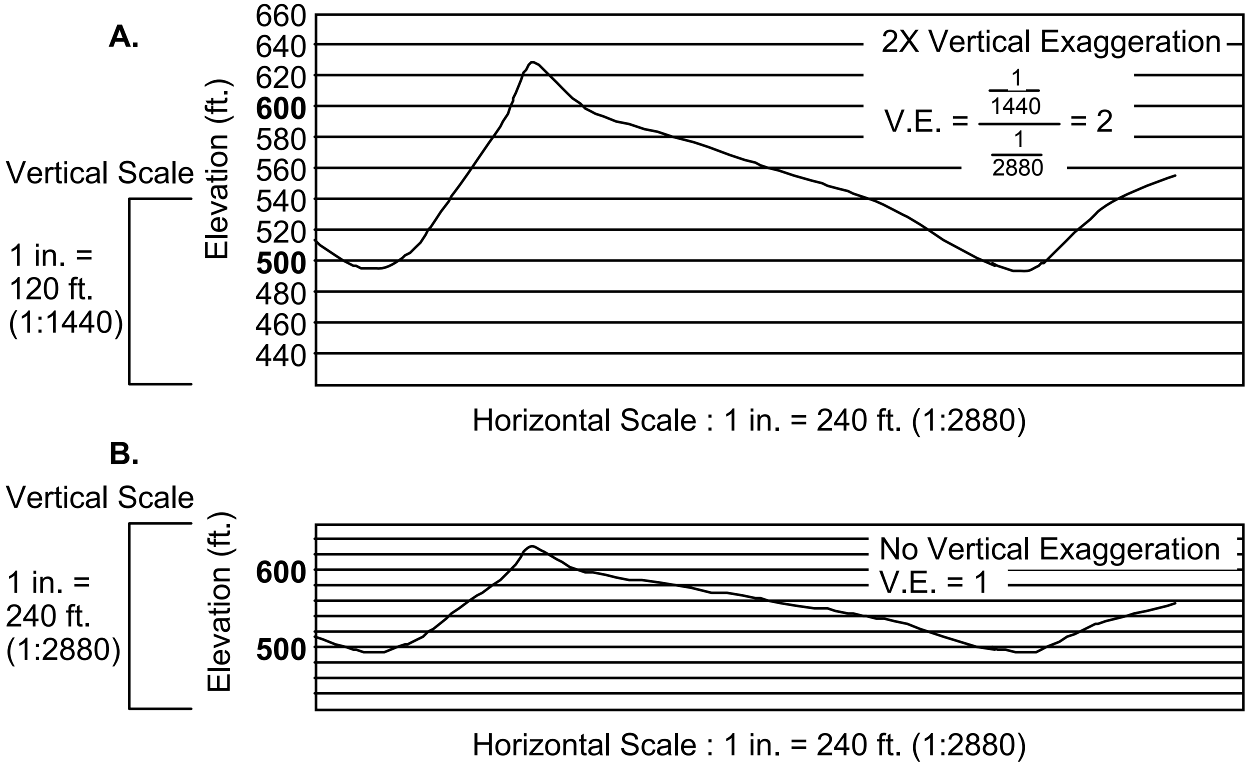


Figure 11.10 Vertical Exaggeration. A. Vertical scale is twice as large as horizontal scale (1/1440 = 2 x 1/2880) so that V.S. / H.S. = 2 = V.E. B. Vertical scale and horizontal scale are equal. V.E. = 1.

Topographic Gradients

A **gradient** is the ratio of vertical change in elevation to horizontal change in distance. Gradient can be thought of as 'steepness' and it answers the question "for every unit of horizontal distance I travel, how far vertically do I go up or down?"

A simple way to remember gradient is the think rise over run. A gradient can be expressed in whatever units are of interest. Typically we Americans express gradient in feet per mile. If the units of vertical and horizontal distance are the same, then gradient becomes a **percent ratio**.

For example:

The maximum steepness of a railroad track bed over any length of track is traditionally given as 2.5%, called 'railroad grade'. This means that, for every mile of track, the elevation change in the rail bed can be no more that 2.5% of a mile. 2.5% = .025 X 5280 ft/mile = 132 feet. Thus, a gradient of 132 feet per mile (=2.5%) is the maximum steepness used for railroad lines.

Estimating a gradient from a map

To estimate a gradient from a map, you must measure two quantities between two points on a map.

1. You must use the map scale to measure the **horizontal distance** (usually in miles or kilometers) between the two points.

2. You must use the contour lines to calculate the **change in elevation** (usually in feet or meters) from the first point to the last point.

   3**.** Dividing change in elevation (rise) by horizontal distance (run) will give the gradient between the two points. If the horizontal distance- and elevation units are different they must be included in the gradient (as in feet per mile or meters per kilometer). If the units are the same they can be reported as a percent.

LABORATORY EXERCISE 11.5
Working with Topographic Contour Maps

The following exercises are based on the **Kilauea Crater, Hawaii 7.5' topographic quadrangle map**. A copy of this map can be obtained from your lab instructor.

1. a. What is the C.I. on this map? __________

b. What is the horizontal scale on this map? __________

2. Locate the following features that will be referred to in the rest of this exercise:

________________ Crater UTM 2**63**,500 E 2,1**46**,200 N

________________ (building) UTM 2**61**,000 E 2,1**53**,700 N

________________ Crater 19° 25' 00" N 155° 17' 30" W

________________ Camp 19° 26' 15" N 155° 16' 15" W

Highway 11 (northwest of Kilauea Crater)

3. Estimate the width of Kilauea Crater along an east-west line running through the middle of the label "Kilauea Crater" on the map. Use a map measurement and the ratio scale to arrive at your estimate.

Distance in **inches** on the **map**: __________________

Actual distance in **feet**: __________________

Actual distance in **miles**: __________________

4. As accurately as possible, estimate the following elevations:

a. The elevation of the bottom of Puhimau Crater. Note the 3621 BM (bench mark) at the lip of the crater.

____

b. The elevation of the Kiluaea Military Camp. __________________

c. The elevation of the rain shed in the northeast quadrant of the map.

____

d. What is the **highest** elevation shown anywhere on this map?

____

e. What is the **lowest** elevation shown anywhere on this map?

____

5. What is the gradient in **feet / mile** of Highway 11, from the 3641' BM to the southern margin of the map?

Lab 12 - Earthquake Location and Isoseismal Maps

PURPOSE

No matter where in the world an earthquake happens, very quickly geologists are able to figure out three things: a. Where did the earthquake occur (where is its **epicenter**)? b. When did it occur? c. How big was it (what was its **magnitude**)? How do they determine these things? Yes, you guessed it, they use *science*! And so shall you... The goal of today's laboratory exercise is to show you how **seismologists** measure and interpret earthquakes. First, you will examine real data recorded around the world from two December 1961 earthquakes located in New Guinea and New Zealand. In doing so, you will come to understand travel-time curves and the **celerity** / time relationships between seismic waves. Next you will use what you have learned to locate the epicenter of an earthquake in the western United States and estimate the time that it occurred. In addition, you will use seismic data to help analyze the state of subcrustal material using an example from Tokyo, Japan (ancestral birthplace of Godzilla and Rodan). Finally, you will make comparisons between **Richter magnitudes** and **Mercalli intensity scales** and plot an **isoseismal map** for a fictitious future earthquake located in the New York City area. Students should take part in classroom discussion based on questions found throughout the exercise. Successful completion of this lab will turn you into an *earthquake expert*. So when the big one comes, while everyone is screaming and panicking, you will stand calmly and announce, “Those were P waves! Here come the S waves! Get ready for the surface waves!”

Travel-time curves

Travel time is the amount of time it takes for an earthquake wave to travel from the place of origin of the earthquake (**epicenter**) to a particular seismic recording station. For any one earthquake, travel times will be different for the different kinds of earthquake waves generated. In brief, **P waves** (primary) travel the fastest and have the shortest travel times. **S waves** (secondary) have longer travel times, and **surface waves** take the longest time to arrive. Consult your lecture textbook for more information on earthquake waves.

Tables 12.1 and 12.2 list the arrival times of earthquake waves at seismic recording stations scattered around the world. All times have been corrected to Greenwich Mean Time (GMT) from the various local times. Table 12.1 gives data on the most-prominent waves arriving at various stations from an earthquake located near the north side of New Guinea occurring on December 14, 1961. Table 12.2 gives similar data for the earthquake located near the northern island of New Zealand which occurred December 27, 1961.

Figure 12.1 is a graph on which travel times have been plotted against distance from the epicenter for the most-prominent waves for each station, except for Matsushiro, Japan (Table 12.1) and Fort Nelson, Australia (Table 12.2). Circles represent data from the New Guinea- and New Zealand earthquakes.

Laboratory Exercise 12.1
Earthquake Travel Times

1. Arrival times, travel times, and distances are given for recording stations listed in Tables 12.1 and 12.2. Calculate the travel times (Arrival time - Origin time) for the Matsushiro- and the Fort Nelson stations, then complete the missing travel-time data in the underlined areas of Tables 12.1 and 12.2. Finally, plot these data on the graph below (Figure 12.1).

2. Draw three lines or curves that give the best fit through the groups of points plotted on Figure 12.1. Based on the relationship between distance and time (= celerity or speed) label each curve as to the seismic wave it represents.

3. A seismograph station somewhere in the world determines that the difference in arrival times (time lag) of the P- and S waves for the New Guinea earthquake is exactly **10 minutes**. Use the graph below to determine the surface distance of this station to the epicenter.

Distance = ________________

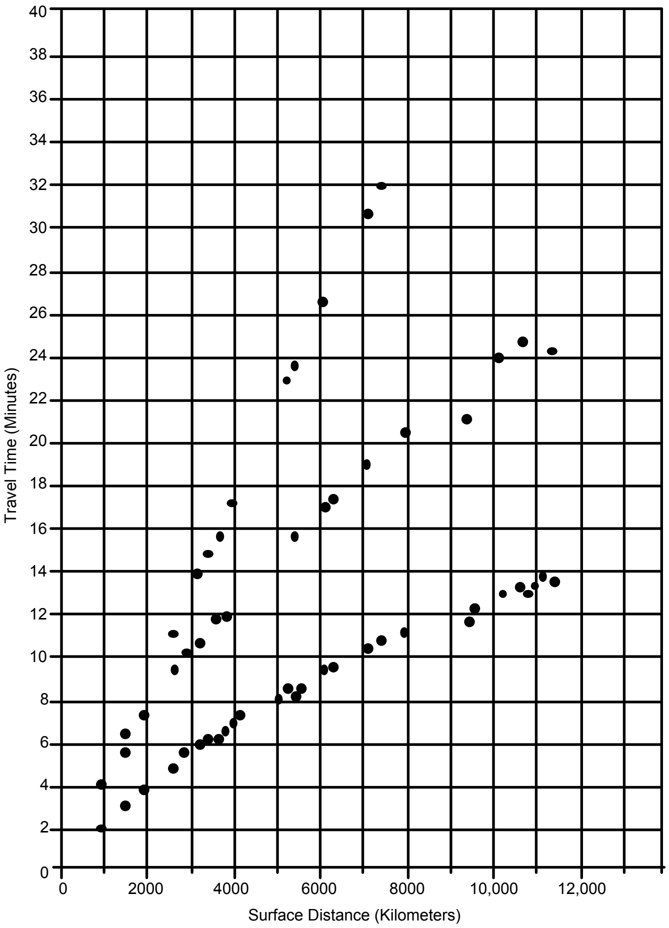


Figure 12.1 - Average travel-time curves for P-, S-, and Surface waves based on the data from Tables 12.1 and 12.2.

Laboratory Exercise 12.1
Additional questions

P and S waves differ from surface waves in that the former travel through the interior of the Earth but the later only travel across the surface. Furthermore, for P and S waves, the farther away the waves travel, the deeper they must penetrate into the interior of the Earth. With these facts in mind, answer the following questions:

4. Examine the shape of each curve in Figure 12.1.

a. How do the curves for the P- and S waves differ in shape from the curve for the surface waves?

b. Consider that in the case of a graph showing distance plotted against time, the slope of the curve defined by the points is a measure of distance / time = speed or celerity. As the slope of the curves for the P and S waves becomes flatter with increasing distance, how is celerity changing?

c. As the recording stations get farther and farther away from the epicenter (increasing surface distance in Figure 12.1), what is happening to the celerity of the P- and S waves?

5. Remembering that P- and S waves penetrate deeper into the Earth the farther they travel:

a. How is the celerity of the P- and S waves changing with depth?

b. Explain why the shape of the surface wave curve is different from the shapes of the P- and S wave curves.

c. Which wave type is most strongly affected by increasing depth?

d. What physical property of the Earth's interior that increases with depth might be causing the change in celerities?

6. Based on the data from these two earthquakes, what can you say about <u>lateral</u> homogeneity in the mantle? Is the mantle exactly the same from place to place? What is the basis for your answer?

Table 12.1 - Data on wave arrivals from the December 14, 1961 earthquake located near the north side of New Guinea.

Location: 3˚ 1' S., 140˚ 9' E. Depth of Focus = 44 km. Origin Time: **07h, 10m, 23s, GMT.**

Seismic Recording Station	**Arrival Times of Waves (GMT)**	**Travel Times**	**Surface Dist. (Km)**
Port Moresby,	07h 12m 32s	02m09s	950
New Guinea	07 14 30	04 07	
	07 14 33	04 10	
Darwin, Australia	07 13 36	03 13	1500
	07 16 08	05 45	
	07 16 58	06 35	
Guam	07 14 23	04 00	1950
	07 17 48	07 25	
Brisbane, Australia	07 16 03	05 40	2860
	07 20 39	10 16	
Noumea, New Caledonia	07 16 40	06 17	3370
	07 25 13	14 50	
Lembang, Java	07 17 01	06 48	3720
	07 22 20	11 57	
Nhatrang, Vietnam	07 17 20	06 57	3960
	07 27 45	17 22	
Mundaring, Australia	07 17 31	07 08	4050
Matsushiro, Japan	**07 17 54**	______	**4460**
	07 23 54	______	
	07 29 57	______	
Vladivostok, USSR	07 18 58	08 35	5230
	07 33 19	22 56	
Chittagong, Pakistan	07 19 51	09 28	6080
	07 27 23	17 00	
	07 37 03	26 40	
Shillong, India	07 19 57	09 34	6250
	07 27 39	17 16	
Irkoutsk, USSR	07 20 58	10 27	7050
	07 29 29	19 06	
	07 41 18	30 55	
Bombay, India	07 21 37	11 14	7890
	07 30 53	20 30	
Tiksi, USSR	07 22 08	11 45	9360
	07 31 42	21 19	
	07 51 26	41 03	
College, Alaska	07 22 56	12 33	9450
Tehran, Iran	07 23 30	13 07	10180
	07 34 32	24 09	
Tananarive,Madagascar	07 23 35	13 12	10200
Tibilisi, USSR	07 23 53	13 30	10730
Moskva, USSR	07 24 14	13 51	10710
Pasadena, California	07 24 20	13 57	11100

Table 12.2 - Data on wave arrivals from the December 27, 1961 earthquake located near the northern island of New Zealand.

Location: 41˚ 2' S., 175˚ 8' E. Depth of Focus = 40 Km. Origin Time: **23h, 48m, 02s, GMT**

Seismic Recording Station	**Arrival Times of Waves (GMT)**	**Travel Times**	**Surface Dist. (Km)**
Wellington, New Zealand	23h 48m 17s	00m 15s	80
Fort Nelson, Australia	**23 52 45**	______	**2330**
	23 56 51	______	
	23 58 15	______	
Brisbane, Australia	23 53 18	05 06	2560
	23 57 36	09 34	
	23 59 13	11 11	
Afiamalu, Samoa	23 54 10	06 08	3200
	23 58 48	10 46	
	24 02 04	14 02	
Charters Towers, Australia	23 54 34	06 32	3620
	23 59 54	11 52	
	24 03 55	15 53	
Byrd, Antarctica	23 56 17	08 15	5000
Mundaring, Australia	23 56 37	08 35	5350
	24 03 52	15 50	
South Pole, Antarctica	23 56 46	08 44	5420
	24 11 48	23 46	
Honolulu, Hawaii	23 58 58	10 56	7330
	24 20 11	32 09	
Pasadena, California	24 01 36	13 34	10670
	24 13 04	25 02	
La Paz, Bolivia	24 01 39	13 37	10800
Shillong, India	24 01 53	13 51	11330
	24 12 33	24 31	

Seismometers

Seismometers are devices that record the shaking of the Earth caused by an earthquake. Electrical signals generated by a seismometer are sent to a **seismograph** where they are recorded. Traditionally, the seismograph consisted of a rolling drum of paper on which a pen drew a continuous line in a spiral around the drum. Recorded earthquake waves caused the pen to jump from side to side, making the familiar ‘squiggles’ of a seismograph trace. The larger the earthquake vibration recorded, the greater is the jump of the pen away from the center line. Alas, we are in the digital age now and the traditional seismograph drum and pen are being steadily replaced by computer recorders and electronic displays.

Laboratory Exercise 12.2

Locating the epicenter of an earthquake

As we have just shown in the previous exercise, owing to the difference in their celerities, P- and S waves originating at the same focus do not arrive simultaneously at a particular receiving station on the Earth's surface. Instead, there is a time lag between arrival of the P- and S waves, and the greater the distance from the focus, the greater is the time lag. The time lag can be determined from seismograph recordings and can be used to compute the distance from the seismograph station to the epicenter, provided that the average celerity of each wave type is known.

7. Four partial records of the same earthquake (recorded at Los Angeles, San Francisco, Salt Lake City, and Albuquerque) are shown in Figure 12.2. The first major deflection, or "kick" (on the left), marks the first arrival of P waves; the second major "kick" marks the arrival of the first S waves. Determine the lag in arrival times at each station, and enter it to the right of the appropriate seismic traces found below in Figure 12.2. Note that times are local (PST and MST) and **not** GMT!

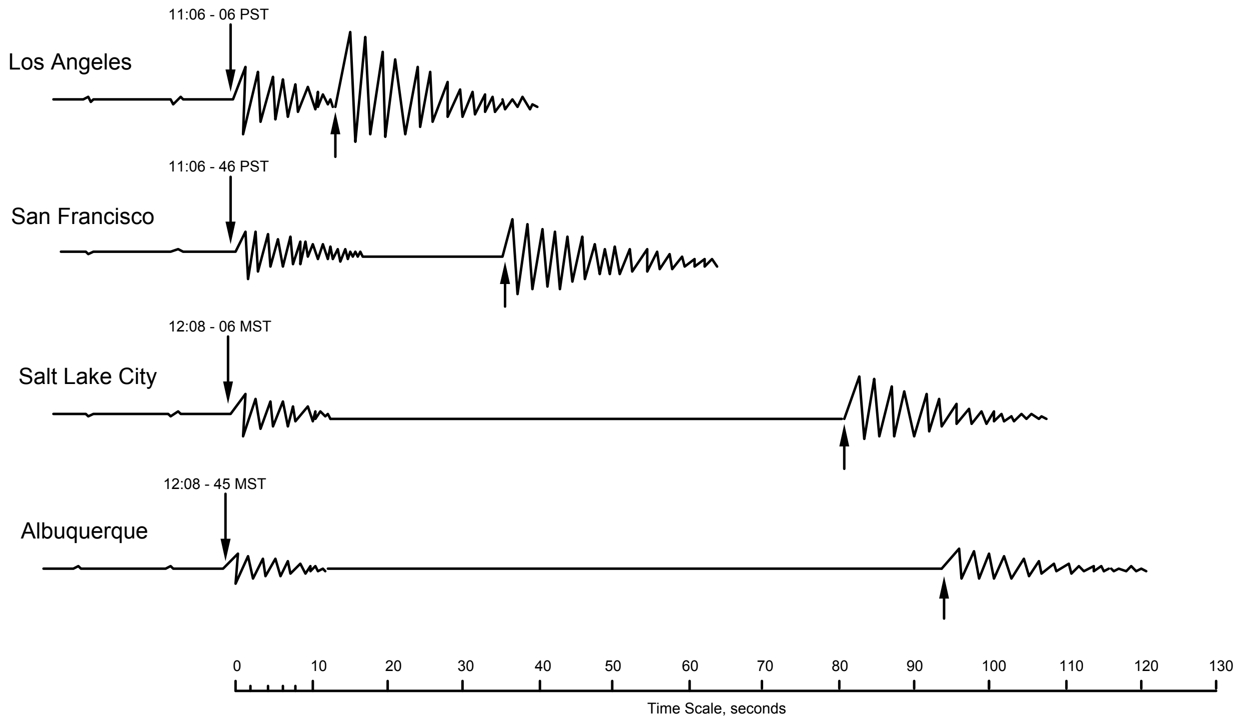


Figure 12.2 - Simplified seismograph records of an earthquake recorded at four stations.

Still More Questions (12.2 cont.):

8. The average celerity of earthquake waves is **3.80 miles/sec** for the **P-waves** and **2.54 miles/sec** for the **S-waves**

a. How long does it take for each type of wave to travel 100 miles? Show how you arrived at your answer.
Hint: Distance (mi) / Speed (mi/sec) = Time (sec)

Show computations here:

Answers:
P-waves_______________sec S-waves_______________sec

b. What is the S wave - P wave time lag at this distance (100 miles)?

Answer:____________sec

9. Determine the distance from each of the four seismograph stations to the epicenter of the earthquake. Set up the calculation for each station as a proportionality: 'Distance' (unknown) / 100 miles = 'time–lag at distance' (answers from 7) / 'time–lag at 100 miles' (answer from 8b).

Show computations here:

Record your results here:

Los Angeles __________miles Salt Lake City __________miles

Albuquerque __________miles San Francisco __________miles

10. Using the distances computed above, **triangulate** to find the location of the epicenter. Open a compass so that the spread from the point to the pencil equals one of the distances to a city listed above. Center the point on the city and draw an arc across the western U.S. After doing this for each city, the arcs should converge near a single point on the map where the epicenter is located. Near what city is the epicenter located?

Answer______________________________

Figure 12.3 - Base map of the western United States for plotting an earthquake epicenter as described in question 12.10.

11. Considering the cause of earthquakes discussed in class, what major geologic feature is located near the epicenter and thus probably related to this earthquake?

12. The time at which the first P waves arrived at each of the four stations is shown on the seismograph record. From this information you can determine when the earthquake actually occurred. The time of arrival of the earthquake waves minus the travel timel will give you the time the waves originated (when the earthquake happened).

To compute travel time use the same formula you used in question 12.8a.

When did the quake actually occur? Compute an origin time for P waves arriving at each of the four recording stations. The four origin times should be the same (or close). If some of your answers seem to be an hour different, think about the time difference between the Pacific and Mountain time zones.

Show origin time computations here:

Answer: Los Angeles ______________ Salt Lake City ______________

Albuquerque ______________ San Francisco ______________

Laboratory Exercise 12.3
The state of subcrustal matter

Seismologists learn a lot about the condition of materials deep in the Earth by studying the behavior of earthquake or other **seismic waves** (such as the waves generated by atomic tests) as they travel through the Earth. For example, we have already seen how it can be shown by the increasing celerity of seismic waves that the Earth increases in density with depth. We also know that, if an earthquake occurs at a particular depth, the state of the Earth at that depth must be solid and relatively brittle to allow the movement of rock that causes an earthquake.

13. Suppose that Tokyo were found to be the epicenter of an earthquake and that the Tokyo seismographs recorded an S-P time lag of 12 seconds. This time lag can be used to estimate the depth of the **focus** (the point in the Earth where the movement that causes an earthquake happens) below Tokyo. Calculate the depth of the focus, assuming an average celerity of **4.8 miles/sec for P waves** and **2.75 miles/sec for S waves** (these celerities are different from the ones given in question 12.8, which were average celerities for long distances travelled obliquely through the Earth).

Show computations here:

Hints: This is the same calculation as in question 12.8. You need to calculate:
Time to travel 100 miles for each wave.
Time lag at 100 miles.
The proportionality: distance/100 = time lag/time lag at 100.

Answer:____________________miles below Tokyo.

14. Considering the depth of the Tokyo earthquake and what you know about the cause of earthquakes, what can you conclude about the state of matter in the Earth below the crust? (The average thickness of the crust is about 22 miles or 35 km).

Earthquake magnitude and intensity

Earthquake magnitude is a measure of the energy released by an earthquake. The scale currently in use was devised by **Charles Richter** in 1935 and bears his name. Richter magnitude is determined by the **amplitude** (height) of the seismic waves recorded on a seismometer. The Richter scale runs from zero to ten, with each unit increase in magnitude equal to about a 30-fold increase in energy released. **Intensity** is a measure of the destructive power of an earthquake and is assessed by measuring how much damage the earthquake caused. After an earthquake the area affected is surveyed and an intensity map showing the pattern of destruction is produced.

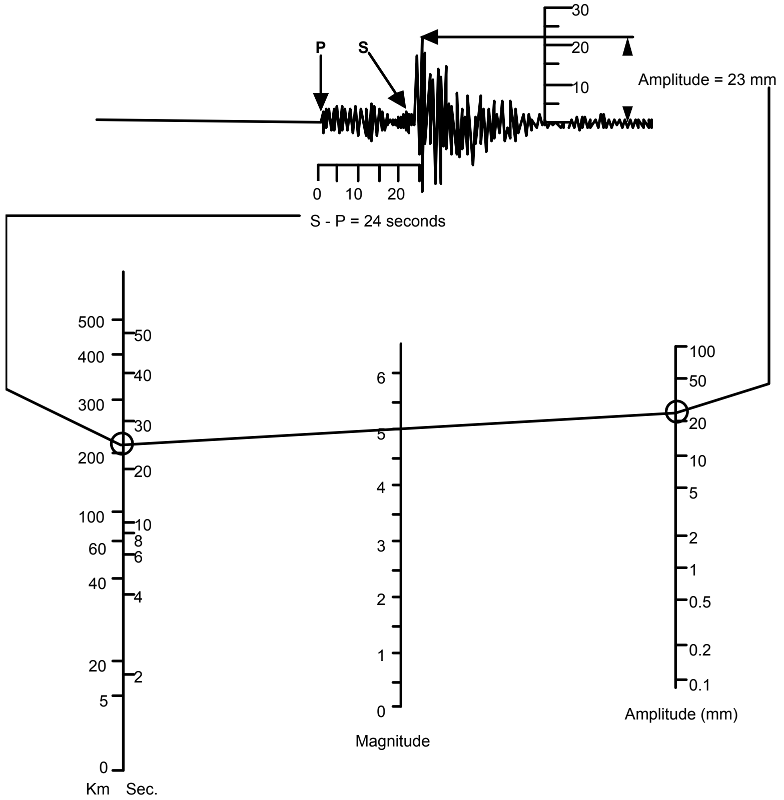


Figure 12.4 - Nomogram for determining Richter magnitude (ML) (from Earthquakes: A Primer, by Bruce A. Bolt. W. H. Freeman and Company).

Procedure for calculating the local Richter magnitude (ML)

A **nomogram** (Figure 12.4) is a graph that enables you by the aid of a straight edge to read off the value of a dependent variable when the values of two or more independent variables are given. To calculate local Richter magnitudes using the Richter Scale nomogram:

1) Calculate the distance from the seismic station to the focus (as described earlier) or the time lag between the P and S waves. Locate this distance on the left hand scale.

2) Measure the height of the maximum wave motion on the seismogram (in the example given = 23 millimeters). Locate this height on the right hand scale.

3) Place a straight edge between appropriate points on the nonogram distance scale (left) and amplitude scale (right) to obtain Richter magnitude (in this case ML = 5.0) on the center scale.

The Modified Mercalli Intensity Scale

The first scale to reflect earthquake intensities was developed by **de Rossi** of Italy, and **Forel** of Switzerland, in the 1880s. This scale, with values from I to X, was used for about two decades. A need for a more-refined scale increased with the advancement of the science of seismology, and in 1902, the Italian seismologist **Giuseppe Mercalli** (1850-1914) devised a new scale on a I to XII range. The Mercalli scale was modified in 1931 by American seismologists **Harry O. Wood** and **Frank Neumann** to take into account the strength of modern engineering structures.

The **Modified Mercalli intensity scale** (Table 12.3) expresses the intensity of an earthquake's effects in a given locality, and is perhaps much more meaningful to the layman because it is based on actual observations of earthquake effects at specific places. It should be noted that because the data used for assigning intensities can be obtained only from direct, first-hand reports, considerable time (weeks or months) is sometimes needed before an intensity map can be assembled for a particular earthquake. On the Modified Mercalli intensity scale, values range from I to XII. The most-commonly used adaptation covers the range of intensity from the conditions of "I - not felt except by very few, favorably situated," to "XII - damage total, lines of slight disturbed, objects thrown into the air." Whereas an earthquake possesses only one magnitude, it can display many intensities which decrease with distance from the epicenter.

Comparison of magnitude and intensity

It is very difficult to compare magnitude and intensity because intensity is linked with the particular ground- and structural conditions of a given area, as well as distance from the epicenter; whereas magnitude depends on the energy released at the focus of the earthquake. For example, in the San Francisco 8.3 earthquake of April 1906, the front porch of a house directly on the San Andreas fault was sheared off but the house otherwise was little damaged, whereas buildings many miles away in the city constructed on bay muds were totally destroyed. Similarly, during the Mexico City 8.1 earthquake of September 1985, structures in Acapulco (directly above the focus) that were built on solid granite fared well. Yet, more than 350 km away from the focus in Mexico City, extreme building damage and loss of life resulted from amplification of seismic waves in the soft, lakebed sediments underlying the city. Table 12.4 gives an approximate comparison of magnitude and intensity.

Table 12.4 - Richter Magnitude and Modified Mercalli Intensity Compared. (After Richter, 1958).

Richter Magnitude	Modified Mercalli Intensity	Description
2	I-II	Usually detected only by instruments
3	III	Felt indoors
4	IV-V	Felt by most people ; slight damage
5	VI-VII	Felt by all, many frightened and run outdoors; damage minor to moderate
6	VII-VIII	Everybody runs outdoors, damage moderate to major
7	IX-X	Major damage
8+	X-XII	Total and major damage

Table 12.3 - Modified Mercalli Intensity Scale

I	Not felt except by a very few under especially favorable circumstances.
II	Felt only by a few persons at rest, especially on upper floors of buildings. Delicately suspended objects may swing.
III	Felt quite noticeably indoors, especially on upper floors of buildings, but many people do not recognize it as an earthquake. Standing motor cars may rock slightly. Vibration like passing of truck. Duration estimated.
IV	During the day felt indoors by many, outdoors by few. At night some awakened. Dishes, windows, doors disturbed, walls make cracking sound. Sensation like heavy truck striking building. Standing motor cars rocked noticeably.
V	Felt by nearly everyone, many awakened. Some dishes, windows, etc. broken; a few instances of cracked plaster; unstable objects overturned. Disturbances of trees, poles and other tall objects sometimes noticed. Pendulum clocks may stop.
VI	Felt by all, many frightened and run outdoors. Some heavy furniture moved; a few instances of fallen plaster or damaged chimneys. Damage slight.
VII	Everybody runs outdoors. Damage negligible in buildings of good design and construction; slight to moderate in well-built ordinary structures; considerable in poorly built to badly designed structures; some chimneys broken. Noticed by persons driving motor cars.
VIII	Damage slight in specially designed structures; considerable in ordinary, substantial buildings, with partial collapse; great in poorly built structures. Panel walls thrown out of frame structures. Fall of chimneys, factory stacks, columns, monuments, walls. Heavy furniture overturned. Sand and mud ejected in small amounts. Changes in well water. Persons driving motor cars disturbed.
IX	Damage considerable in specially designed structures; well-designed frame structures thrown out of plumb; great in substantial buildings, with partial collapse. Building shifted off foundations. Ground cracked conspicuously. Underground pipes broken.
X	Some well-built wooden structures destroyed; most masonry and frame structures destroyed with foundations; ground badly cracked. Rails bent. Landslides considerable from river banks and steep slopes. Shifted sand and mud. Water splashed (slopped) over banks.
XI	Few, if any, (masonry) structures remain standing. Bridges destroyed. Broad fissures in ground. Underground pipelines completely out of service. Earth slumps and land slips in soft ground. Rails bent greatly.
XII	Damage total. Practically all works of construction are damaged greatly or destroyed. Earthquake waves seen on ground surface. Lines of sight and level are distorted. Objects are thrown upward into the air.

Laboratory Exercise 12.4
Earthquake magnitude and intensity

15. Using the Richter Scale nomogram in Figure 12.4, determine the Richter magnitude of the following earthquakes:

	S-P (sec)	**Amplitude (mm)**	**Magnitude**
(a)	02	20.00	____________
(b)	10	20.00	____________
(c)	50	20.00	____________
(d)	20	0.20	____________
(e)	20	2.00	____________
(f)	20	20.00	____________

16. For problems (a-c) of Question 15, the amplitude of the seismic wave remains constant but the distance to the epicenter (as measured by the difference in travel time between S waves and P waves [S-P]) changes. Can you identify and state a simple relationship between distance to epicenter (S-P), amplitude, and magnitude for these problems?

17. For problems (d-f) of Question 15, the value of S-P remains constant, but the amplitude changes by a factor of 10. Can you identify and state a simple relationship between S-P, amplitude, and magnitude for these problems?

LABORATORY REPORT (R3)

A FUTURE EARTHQUAKE IN NEW YORK CITY - It could happen here!

SCENARIO

On a snowy, windswept Friday night, late in the year 2027, a Richter magnitude 6.4 earthquake rips through the heart of New York City causing localized major damage. Chimneys are toppled as far away as Philadelphia, Pennsylvania and the quake is felt throughout southern New England. To those in the know, the temblor is not unexpected given the previous magnitude 5.0, 5.2, and 4.0 events in 1737, 1884, and 1985. Despite the warnings of scientists, the city is not prepared. Amazingly, very few people are injured excepting the usual weekend homocides and a painful hang-nail incident in Brooklyn. As described below, you've been assigned to collect eyewitness accounts on the intensity data for the event.

OBJECTIVE

The objective of the following set-in-the-future exercise is for you to become familiar with the **Modified Mercalli Intensity Scale** and to construct an **isoseismal map**. Isoseismals are lines that connect points of equivalent reported earthquake intensity and thus enclose areas of approximately equal earthquake damage. The resulting pattern discloses something about location of the **epicenter** and local ground conditions. This is not as simple as you might think as damage is a function of many variables. As stated in the late twentieth century by now retired Hofstra professor Charles Merguerian, "Given the frailty of the infrastructure of New York City, Mercalli intensity reports can be expected to exceed their Richter magnitude calibrations".

What to do:

In order to gather information on the pattern of damage after an earthquake, teams of people with intensity-data questionnaires are sent out to canvas for information on the earthquake's effects throughout a large area surrounding the epicenter. **On the following pages are damage reports such as would be collected in the aftermath of an earthquake.**

a. Using Table 12.4, analyze these reports to determine the local Mercalli intensities for each location and record them in the left margin adjacent to the damage reports.

b. Plot the intensities on the base map (Figure 12.5) for each of the given report locations and contour the isoseismal lines, dividing the map into regions according to intensity. Note that the isoseismal lines follow the basic rules of contour lines; they need not close and can just be drawn off the limits of the map. If necessary, refer back to Lab 11 for drawing contour maps. For ease of contouring, use the following combined intensity units:

I-III, IV, V, VI, VII, VIII-IX.
(See Table 12.4)

DISCLAIMER:

This laboratory exercise in no way reflects a desire for or endorsement of future seismicity in the New York City area. Rather, the tongue-in-cheek style is meant to amuse and educate by presenting a <u>plausible</u> scenario of what might occur if an earthquake were to strike close to home.

So, getting into your official Hofstra University Geology Department rental van, you follow this pre-planned survey route:

To start, you drive south from Hofstra University onto the Meadowbrook Parkway southbound to the Southern State/Belt Parkways westbound. Get onto the Interborough Parkway northbound then across the Triborough Bridge into Manhattan.

Hempstead - Hofstra University's Unispan swayed for 10 seconds and small cracks appeared in structure. Students, disturbed that root beer has spilled in Hofstra USA, start rumor that an oil truck has plowed into Unispan. The Old Midnight Mission in the center of town suffered minor damage.

Roosevelt Field - a few reports of broken dishes and windows in the mall. Surrounding neighborhoods report only old, poorly designed buildings suffered much damage.

JFK Airport - cracked plaster in older parts of terminal buildings. Quake was felt by most everyone who was waiting for flights. Flight crews and scared passengers drinking heavily at the airport bars did not notice anything unusual. Light aircraft waiting for takeoff rocked noticeably.

Coney Island - old pendulum clock in Veterans' Hospital stops; many residents report cracked plaster and minor beach erosion resulting from sand liquifaction and flow.

Prospect Park area - All residents feel the event and run outdoors. Sculptures in Prospect Park area and residential chimneys collapse. Hot dog stands overturned.

Long Island City - major cracks in older buildings with some failure of unreinforced building facades. People driving report experiencing severe shaking over and above monster pothole vibrations.

East Harlem at 125th Street near Triborough Bridge - considerable structural damage to many old buildings; a few buildings partially collapse with failure of most unreinforced building facades. Bridge swayed noticeably for 25 seconds with some suspension cables snapping and roadbed failure. As a result, bridge closed to all non-essential vehicular traffic. Not you, of course!

Central Park (110th Street and Fifth Avenue) - monuments thrown from their bases and fish leap from lakes. Sand and mud ejected from lakeshores. Buildings on 110th street in partial collapse with lots of heavy furniture overturned. Across town, turnstile jumpers at the Columbia University IRT station pay to **leave** the subway.

Next, take East Side Drive south to Lower East Side area then pick up West Side Highway northward to the Lincoln Tunnel into New Jersey.

Queensboro Bridge - large sections of bridge have collapsed and cables for tramway snapped sending Roosevelt Island cable cars bungee jumping. Bridge found to shift off of its foundation. All vehicular traffic diverted to other crossings. Underground pipes explode sending asbestos-rich steam geysers into air. Other East River bridges show similar damage.

Midtown Manhattan - a few serious pedestrian lacerations reported as thousands of glass windows fall from swaying skyscrapers. Unreinforced buildings are totally demolished. Broken gas mains explode and erupt into roaring fires at some manhole openings. Water mains rupture sending line pressure down to zero. Water towers fail sending millions of gallons of water to the streets. Luckily, this "meteoric" water puts out fires which started immediately after the first seismic jolt. Grand piano moves across stage of Carnegie Hall during performance of Khatchaturian's Gayane Ballet Suite. Conspicuous ground cracks noted in blacktop; some large enough to slow speeding taxi cabs.

Lower East Side - everyone who was awake felt quake, and those not awake were launched out of bed. Everyone runs outdoors to see what happened. Damage reports indicate slight- to moderate damage in most structures. Some chimneys broken and considerable damage to older structures.

Staten Island Ferry Terminal - some people waiting for ferry awakened by quake. Many people frightened and fear ferry has crashed into pilings. Slight damage reported.

Chelsea area - everyone felt quake. Many residents trapped in elevators for hours. Drivers are not able to steer properly resulting in pedestrian accidents.

Lincoln Tunnel area - severe shaking in tunnel reported as soft Hudson River silts amplify seismic waves. A sharp snapping noise was reported by toll takers and tunnel safety crews. Inspectors, fearing the worst, reported "minor cracking in tunnel walls but only minor amounts of river water leaking in as yet". Toll booths fall apart sending loose change all over roadway; major traffic jams ensue as motorists attempt to "get even". You decide to drive through tunnel, with a watchful eye on tunnel walls, to get lunch and complete your intensity survey.

Weehauken and West New York - everyone feels quake and runs outdoors in their pajamas. Hanging lamps swayed and oriental rugs slipped on floors. Minor cracks in plaster are reported.

North on U.S. Routes 1 & 9 then west on Route 80 across Newark Basin to Paterson, New Jersey then northeastward on the Garden State Parkway across the Tappan Zee Bridge into Westchester.

Edgewater - considerable damage to old buildings the result of blocks of the Palisades diabase collapsing onto their property.

Stop at Nathan's for lunch. After a few hot dogs and burgers, return northward on U.S. Routes 1 & 9.

Fort Lee - some people report fallen plaster; Ernest P. Worrell said his chimney broke.

Lodi - some people were frightened; some fallen plaster reported.

Paterson - one woman said she saw light posts swaying; most people felt shaking but thought it was a heavy truck not an earthquake.

Westwood - most people felt quake. Dave Franklin was up late. He felt the quake, saw his car move, and heard the house creaking. Went back to bed to "sleep it off".

Spring Valley (side trip) - a few objects suspended from ceilings swayed and felt by only a few residents. No damage reported.

West Nyack - a few Boy Scouts were awakened by quake; policeman at Dunkin' Donuts drops donut into coffee but not sure what caused it.

Tarrytown - some people awakened by creaking sounds from walls.

Briarcliff Manor - not felt except by one resident, a diehard Jimi Hendrix fan.

South on Route 87 (Thruway) to the Cross Bronx Expressway then eastward through the Bronx to Orchard Beach.

Ardsley - everyone felt quake. Older people, remembering the 1985 quake, report that this one was much worse. Some heavy furniture moved; fallen plaster reported; dogs slink away.

Yonkers - everyone felt quake; fallen plaster; cats bark.

Riverdale - many people awakened and run outdoors. Considerable damage reported from older buildings. Gas mains break and fires erupt throughout the area.

Bronx Zoo - animals behaving strangely for hours before the quake. Large desks in administration building move and cracked plaster reported. Damage is reported as slight. Most animals awakened except for ground sloth.

Orchard Beach - Light poles along concrete boardwalk disturbed and minor sand slumps into Long Island Sound. Trees sway and a Will Jefferson Clinton, who claims to a be a former U.S. President, reports his pendulum clock stopped swaying and he fears for his life as he is without health insurance.

Route 95 south to the Throgs Neck Bridge into Queens and then back to Hofstra's Geology Department.

Flushing - quake felt by all. Dishes in Chinatown district smashed from falling onto floor. Overall damage slight but a few localized instances of severe damage and cracked chimneys, the results of soft-sediment ground slumping. Whitestone bridge closed as 40-second-long harmonic swaying cracks the roadway.

LaGuardia Airport - All airport traffic halted as slumping of soils results in cracked runways. Building damage is negligible except for parking garage which partially pancakes onto 350 cars. Motorists still charged daily rate until vehicles moved.

Bayside - All residents feel quake and run outdoors. Damage is slight but pervasive.

Great Neck - Most residents feel quake but not sure what happened. Many reports of cracked plaster and interruption of shopping routines.

Glen Cove - Those indoors sense heavy truck or low-flying plane but those outdoors report nothing. Some slumping of sediments from north shore cliffs.

Hicksville - Many residents report shaking indoors but few outdoors notice event. Cars stopped in traffic on LIE report shaking event to their spouses on car phones. No damage to structures reported.

Hofstra Geology Department - Return to the department office and hand in your report.

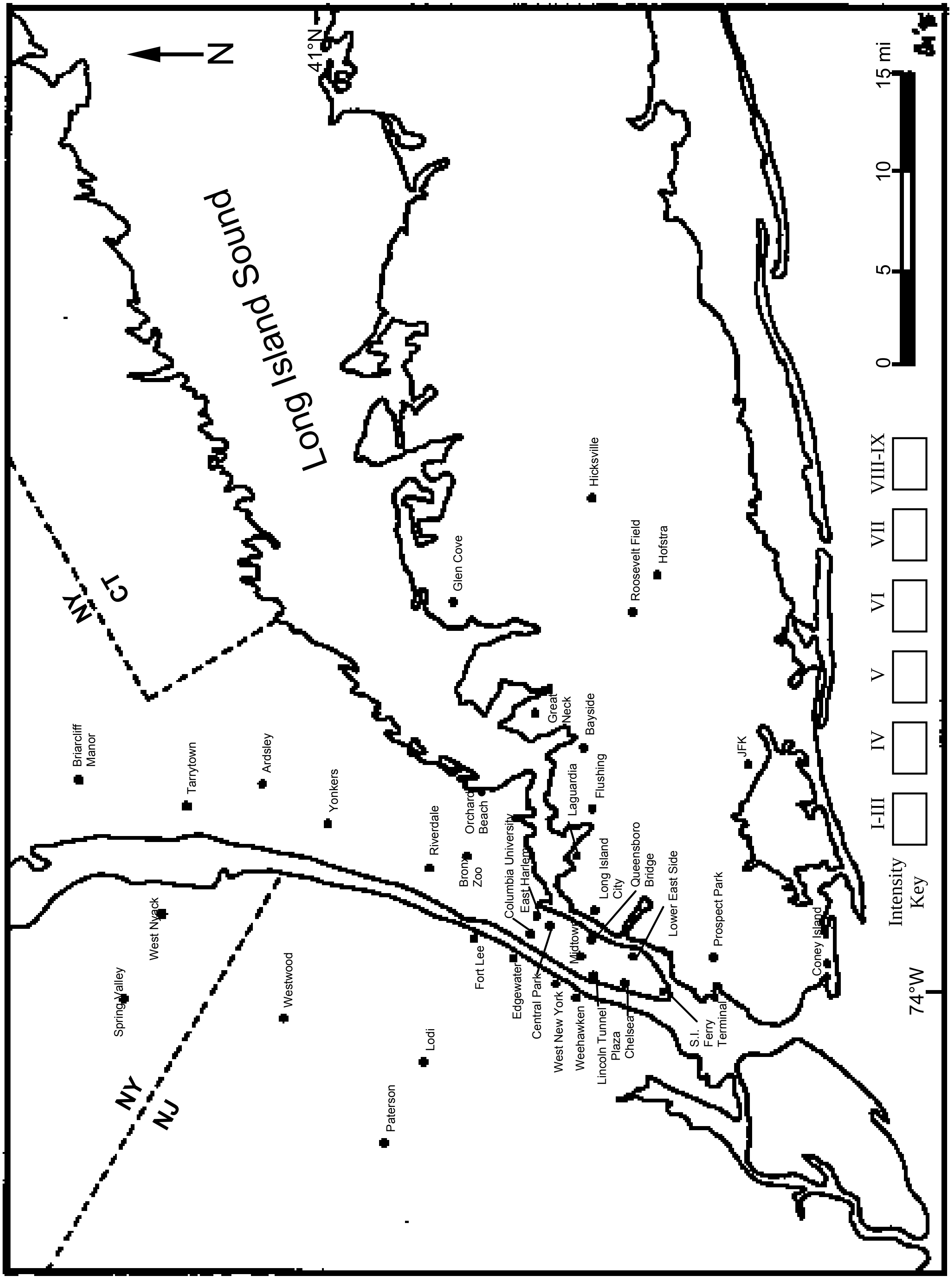


Figure 12.5 - Base map of New York City area for plotting isoseismals. Roads, bridges, and tunnels are ommitted for clarity.

LABORATORY REPORT (R3) **NAME: ____________________**

Supplemental Questions

1. Identify the probable epicenter of the earthquake by locating the region of highest damage intensity on your isoseismal map.

2. Are there any apparent trends to the pattern of damage caused by the earthquake? If so, identify the trends and speculate as to what might have caused them (think about what features cause earthquakes).

3. Explain the the differences between the three types of earthquake waves. Discuss differences in their style of propagation, speed and region of motion through the Earth, and in their destructive effects.

4. Explain the difference between earthquake intensity and magnitude.

5. Explain what factors control the damage (intensity) caused by an earthquake. Why might a smaller earthquake centered in New York City cause more damage than a larger earthquake centered elsewhere?

Optional. Using references such as newspapers, magazines, and encyclopedia yearbooks, research and report on an earthquake that has occurred in the last five years. Your report should be about a page in length and cover such points as location of the epicenter, magnitude, and effects on the surrounding landscape and residents. Don't forget to cite your reference(s) and include a bibliography.

Appendix A: Geologic History of the New York Region

PURPOSE

During Lab 2 or 3 your professor will describe the all-day field trip to be conducted later in the semester. From this description, you will need to decide whether or not you plan to participate in this field trip. The field trip is by election only and all students need to register, in person, with your lab professor **BEFORE** the end of Lab 4! Those of you who take the field-trip option can also opt to produce a photo essay to replace your lowest test grade in the lab. A field-trip guide will be provided for those of you who come and this guide will dovetail nicely with any pre-trip discussion.

The material on the following pages is intended to supplement your professor's presentation. The material is in two parts: (1) how James Hutton's "Great Geological Cycle" was used to turn the study of the Earth completely around and later became the basis on which the geologic time scale was established; and (2) an essay on the geologic history of the New York City area.

How James Hutton used the "sands of time" to reverse the "hands of time"

Late in the eighteenth century, bookstalls were awash in volumes entitled "Theory of the Earth." The numerous authors who wrote such books did so because they thought they had it all figured out. These "Theory" books followed a simple pattern. They began with a statement about the origin of the Earth and then narrated a succession of events through which the author visualized that the Earth had passed in changing from its presumed nascent condition to the situation that exists today. Nearly all of them featured a "Universal Ocean," which by some means or other shrank down from covering the entire surface of the Earth to its present size (covering about two-thirds of the surface of the Earth). All authors wrote that in this marvelous body of water, many natural events happened. The list ranged from depositing sediments that would become rock to shaping its bottom into such a form that when the water level dropped far enough, there to behold would be the landscapes of the modern world. In all of these "Theories of the Earth," time flowed forward: the beginning of time was the assumed origin of the Earth.

A small group of brilliant Scotsmen in the circle of one **James Hutton** (1726-1797) of Edinburgh helped turn the whole study of the Earth upside down. Hutton's friends went with him on geological "walks," made sketches, and inspired him to write down what he had been discussing with them. One of Hutton's closest friends was the mathematician John Playfair (1748-1819). Playfair was so devoted to Hutton that after Hutton died before he was able to complete work on the final volume of his book, Playfair spent 5 years writing a book entitled "Illustrations of the Huttonian theory of the Earth." Indeed, most of our understanding of Hutton comes from those who have read Playfair's book, as Hutton's own writing was often dense and difficult to understand.

By Hutton's time, geologic science had progressed to the point where it was generally agreed that beneath a surface layer of loose soils or "dirt" (**regolith** of the modern usage) one encounters solid bedrock and that much of this bedrock consists of layers (named **strata**) that form layered successions. In many places, the strata are horizontal, the same position in which they were formed initially. But in mountain chains, the strata are inclined at various angles, including vertical.

Among Hutton's great contributions was his insistence that the strata consist of "materials furnished from the ruins of former continents." In today's language, we would say that Hutton was referring to **sedimentary rocks** and that the sediment of which they consist had been eroded from (thus being "the ruins" of) a former continent. To explain how this could happen, Hutton described what he called a "system of universal decay and degradation." In his view, sediment would be deposited in the sea, converted to strata, and later be elevated to form new land. These new lands would be eroded and the debris carried into another sea, and so on into the future. Hutton's friend, John Playfair gave the name "great geological cycle" to what Hutton called a "system of universal decay and degradation." Today, Hutton's ideas are known by Playfair's term. Applying the established concepts of Nicholas Steno (1667) who stated that younger strata overlie older strata and by tracing groups of strata, Hutton was able to work his way backward through the geologic record.

In trying to explain the geologic record, Hutton began by turning "time" around. Instead of following the approach used by all the other authors of "Theory-of-the-Earth" books, Hutton did just the opposite. As mentioned, in these "Theory..." books, time flowed forward. The other authors were, in effect, telling the world how it should have behaved (according to their imaginations about the past).

Hutton changed everything by turning the whole business completely around. He began by observing what was happening today and how modern processes such as erosion were creating new layers of sediments (a new geologic record). Then he started working backward, trying to answer the question, How far back into this record does one find rock products of the same general processes that prevail today? And to his utter astonishment, he found that the oldest rocks do not differ significantly from the youngest ones. Hutton announced that he could find "no vestige of a beginning nor prospect of an end."

The Geologic Record: Bedrock, Regolith, and Shape of the Earth's Surface

The geologic record consists of three components, two modern and one ancient. The modern components are (1) the shape of the Earth's surface and (2) the loose "dirt" (technically known as **regolith** or **sediment**) that underlies the surface to variable depths in most places. The ancient component is (3) the **bedrock**, which is defined as the continuous solid rock that everywhere underlies the regolith and locally is exposed at the Earth's surface itself. The geologist's tool of choice in working with the regolith is a shovel; that of the bedrock, a hammer.

According to the ways that the different bodies within the bedrock have formed, geologists classify them into three rock families: (A) **igneous rocks** (formed by the cooling of molten material, such as the lava that comes out of a volcano), (B) **sedimentary rocks** (formed by the lithification of sediments), and (C) **metamorphic rocks** (formed by the effects of heat and/or pressure on some preexisting rock).

By identifying all these rocks and noting the locations, dimensions, and relationships at the boundaries of one body of rock with another, geologists are able to organize the ancient bedrock into units. Furthurmore, geologists have established a relative chronology of time based on the sequential overlap of these units. Such an established series of units of known relative ages is the basis for the geologic time scale (Table A-1). In the next section, we explain the fundamental principles on which relative ages of rock units can be determined.

Telling time from the rocks

Using Hutton's approach, an armada of geologists was able to work out a relative **geologic time scale** (Table A-1). The most-basic principle (as proposed by Steno) is that sedimentary strata are horizontal in their original positions. This is known as the **principle of original horizontality**. Closely associated with this principle is that, unless they have been disturbed, the oldest layers are at the bottom and progressively younger layers are higher up (**Steno's principle of superposition**). A third Stenoan principle is known as the **principle of crosscutting relationships**, which states that if one body of rock cuts across another, the cut-across one is older. This is self evident; the older one had to be in place before something could happen to it. A somewhat comparable principle is known as the **principle of inclusions**, or that only an older rock can be found as an inclusion within a younger rock. The last basis for establishing a relative time scale (attributed to **William Smith**) is the **principle of faunal succession**. This is a statement of the observed fact that within the sedimentary strata are distinctive fossil remains. Given this situation, it is possible to use fossils to match the ages of layers from place to place with the general proviso being: **same fossils, same age**.

Keep in mind that these principles provide methods of relative-time analysis and thus form the basis of our relativistic geological time scale. Absolute time (the numbers printed in your textbook at geologic period- and era boundaries) are the products of twentieth-century geochronologic studies of radiogenic materials in minerals and rocks and have been (and will continue to be) updated as new dating methods are developed and utilized.

Essay on the Geologic History of the New York City region

Armed with Hutton's brilliant approach, geologists have been able to decipher the history contained in the Earth's geologic record. In attempting to unravel the geologic history of any area, a geologist first examines how the geologic cycle is operating there today. Of special importance in this examination are the climate (for its influence on the water circulation and distribution of plants and animals) and the position of the surface of the sea, or sea level (for its control on the flow of rivers, location of zones of breaking waves, and pattern of sites covered with seawater, another major influence on the distribution of organisms).

The purpose of this essay is to show you how modern geologists have used James Hutton's concept of the "Great Geological Cycle" and have applied his approach in working backward from the modern world into the geologic record. By doing this, we begin with the youngest components of the geologic record, the shape of the Earth's surface and the regolith, and work our way backward into the bedrock. After we have examined and unraveled all the major parts, we can synthesize a geologic history of our local area in which we start with the oldest event and let time flow forward to the present day.

In attempting to pry loose the geologic secrets of any area, geologists "take the region apart," so to speak, one single large layer at a time. For the purposes of our first look at New York City and vicinity, we need only six major layers (Table A-2). As we shall see, relationships found at the contact surfaces between successive layers contain evidence for inferring other events in the historical development of the area. The usual implication is that times of erosion intervened between the accumulation of the layers. Typically geologists number the layers "from the bottom up;" that is, the oldest layer is numbered 1; the next oldest, 2; and so forth. We shall follow this method of numbering, but because we are going to work "from the top down," the first layer we discuss bears the number 6, the next 5, and so forth.

Layer 6: The Holocene sea and its sediments. Time span: from now back to roughly 4,000 years ago; an era of generally rising sea level.

The sea and its associated sediments form the topmost layer of the New York City region's geologic "layer cake." Although the sea itself will not be preserved as part of the future record now being created by the operation of the geologic cycle, in the sea is where much of the material which will compose that record is accumulating. The New York City region possesses a varied shoreline. Along the S coast of Long Island and E coast of New Jersey (the shores of that part of the Atlantic Ocean known as the **New York Bight**) are marvelous sandy ocean beaches. The waves have smoothed these stretches of the shoreline facing the ocean and are actively shifting sand along them. Along the northern sector of the New Jersey coast, the sandy shore sediments travel relentlessly northward toward Sandy Hook. Along the south shore of Long Island, the sediment moves predominantly westward toward Rockaway (Breezy) Point.

Away from these beaches exposed to open-ocean waves, the shoreline plan is irregular. Typical features are numerous bays, estuaries, and intertidal marshes. New York is being submerged at a rate of 1 mm/yr. This means that the edges of bays and marshes tend to overspread the adjacent higher land and that the tops of the marshes are thickening upward at the same rate of 1 mm/yr. (This thickening is possible only on marshes that have not been diked or reclaimed by humans. The thickening takes place because the marsh grasses can grow upward with the rising water and are able to trap silt brought in by the incoming tides.)

What is more, the rising seawater tends to check the speed of inflowing river waters and to trap stream sediment in the estuaries at the river mouths. Despite the rising seawater, the Hudson River pours 23,000 cubic feet of fresh water per second into the New York Bight, and with it, 3,000 tons of suspended sediment per day. Despite this amount of suspended sediment, no delta is present at the mouth of today's Hudson River, as is the case in the Gulf of Mexico, for example, at the mouth of today's Mississippi River.

The rising seawater and its veneer of sediments are now in contact with each of the five older "layers" of the New York City region's geologic underpinnings. About 12,000 yr ago, the shoreline stood 75 miles or so seaward of its present location (Figure A.1) as the growth of continental glaciers (Layer 5, below) produced a marked global (**eustatic**) drop in sea level. The Holocene beaches then lay at what is now the edge of the continental shelf, many miles out to sea.

Layer 5: Pleistocene deposits made by a continental glacier. Time span: from about 12,000 yr ago back to 20,000 yr ago; the **last** of a number of glaciers that spread into the New York City region.

Layer 5 brings us face to face with the last great Ice Age. Geologists do not know as much as they should about the climatic situation during the so-called Ice Ages, the last 10 million years or so during which cyclical changes in the Earth's climate brought about the alternate spreading- and melting of continental glaciers. The geologic record available in the New York City region sheds no light on the long interval between 10 million years ago and the time when the first of the several irregular layers of stony debris (**till**) was left behind by the oldest of the continental glaciers to visit this region. According to Sanders and Merguerian (1994 a,b; 1995) and Sanders et al. (1995, 1997), as many as five time separated glaciations may have affected the New York City area over the last 400,000 years (Table A-3).

The outermost margin of the largest of these ice sheets (age not well known) deposited what is known as a **terminal moraine**--an irregular ridge containing numerous gigantic boulders. One such terminal moraine extends along the center line of Long Island. Nothing but a continental glacier could have transported the boulders and have built Long Island's "backbone" ridge (now followed by the LIE). The terminal moraine not only extends E-W across Long Island into Long Island City, but it crosses the lower part of New York Harbor at The Narrows (site of the Verrazano-Narrows bridge). From The Narrows, the terminal-moraine ridge can be followed across Staten Island and then picks up again in northern New Jersey.

When the ice melted, the water created various "temporary" lakes (temporary in the geologic sense; actually some of these lakes lasted hundreds- or even thousands of years). From the margin of the glacier southward, rivers spread an apron of sand and gravel (**glacial outwash**).

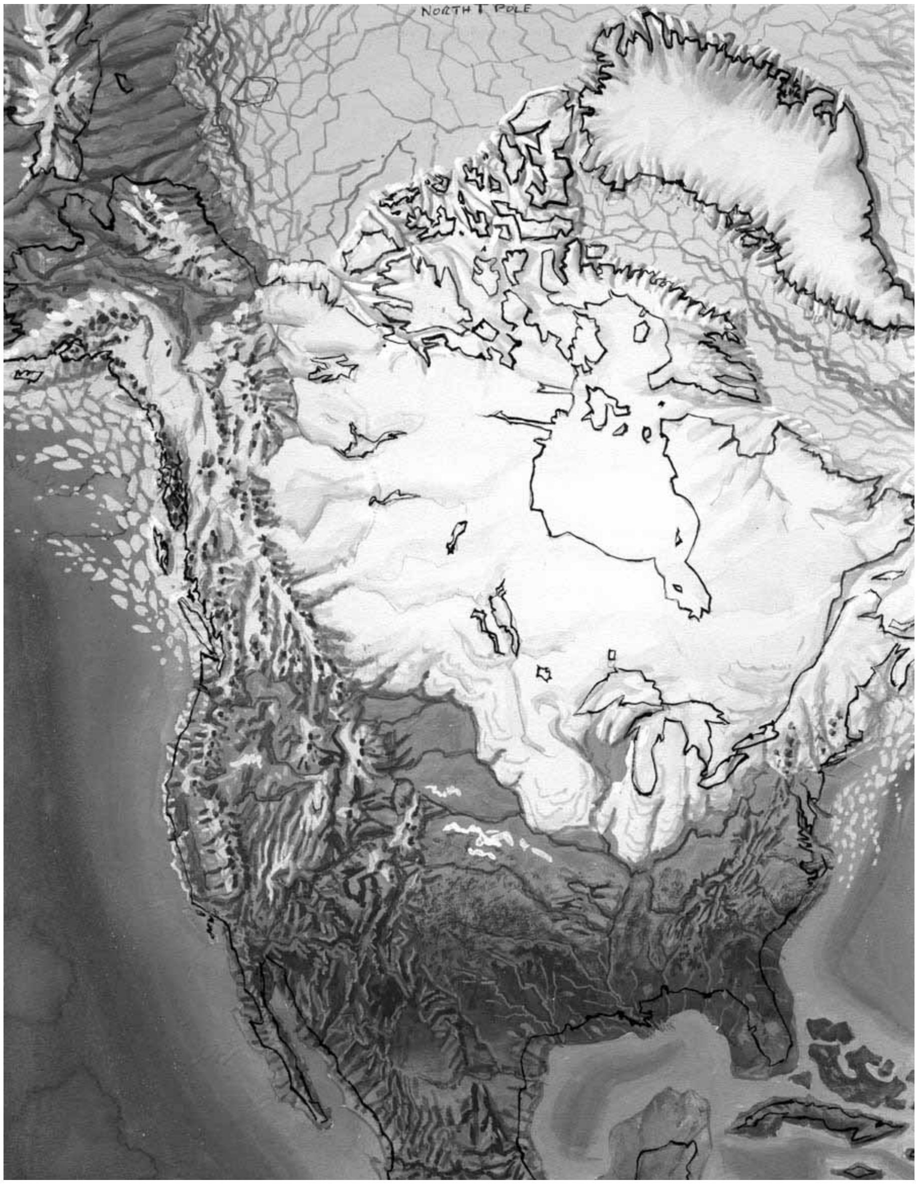


Figure A.1 - Physiographic map of North America during the Pleistocene showing the extent of continental glacial ice and the oceanward migration of the shoreline, the result of a global (eustatic) drop in sea level. (Courtesy Dr. Rhodes W. Fairbridge).

From its outer fringe northward, the glacier extended as far as Labrador. (See Figure A.1.) It covered the highest peaks in New England and everything else for hundreds of miles. The only way to begin imagining what it must have been like in New York City and on Long Island during an Ice Age is to make comparisons with the edges of one of the two ice sheets present in the world today. Make your choice: Greenland? or Antartica? Imagine a 5,000 foot thick glacier descending on New York City today (Figure A.2).

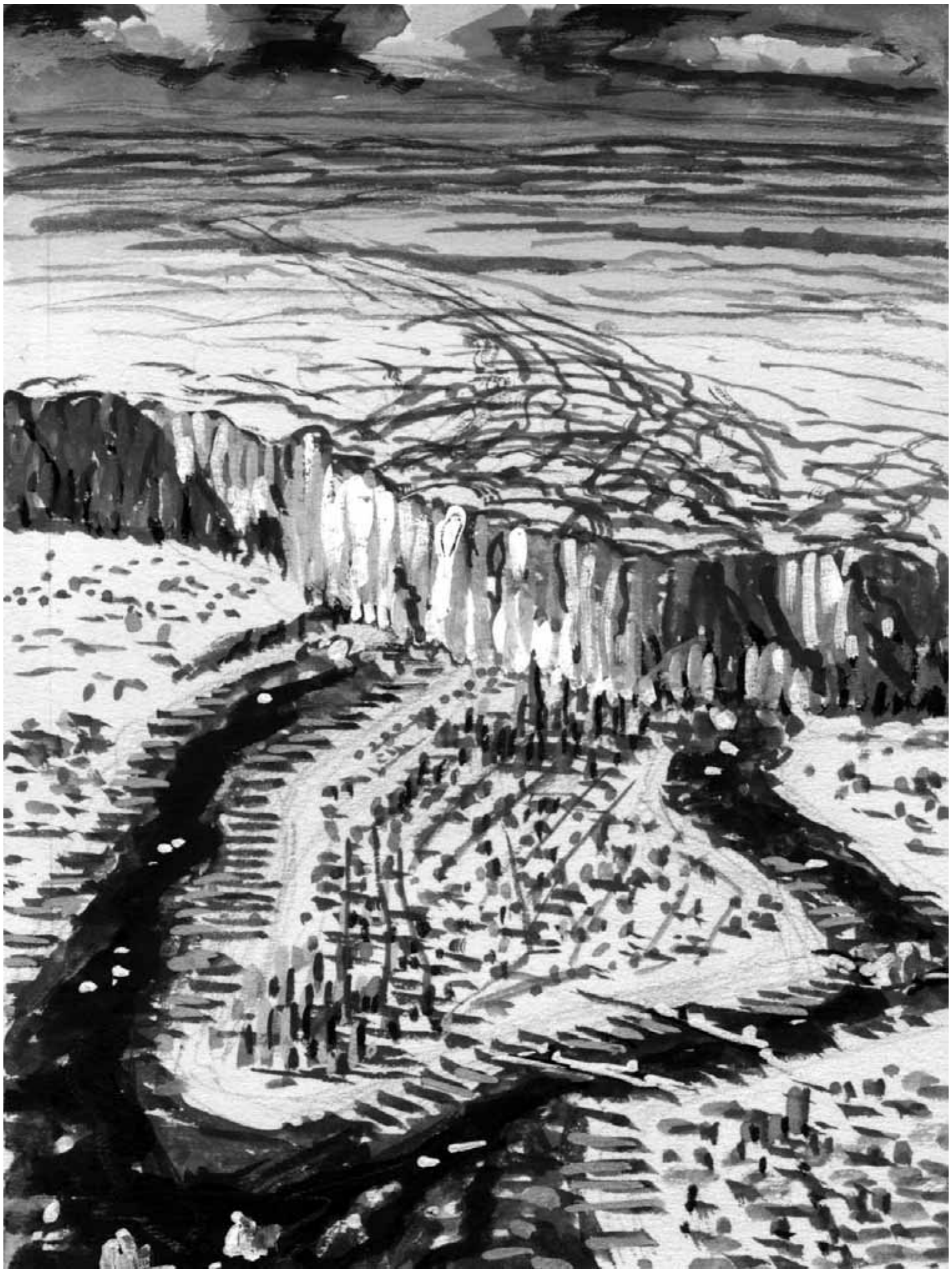


Figure A.2 - Imaginary sketch of the New York metropolitan area after the descent of a continental ice sheet, similar to the one that descended upon the region during the Pleistocene. Sea-level is shown as it is at present to keep the geography recognizable (Courtesy Dr. Rhodes W. Fairbridge.)

Glaciers totally rearrange the countryside over which they flow. The now-vanished glaciers that visited the New York City region filled in some valleys, opened up other valleys, dammed up glacial lakes, transported big blocks of rock far and wide, and churned up the soil, mixing it liberally with stones plucked from near and far. Research by Sanders and Merguerian cited above strongly suggest that four and possibly five glaciers affected the New York City area. The youngest glacier flowed SSW down the Hudson Valley (Figure A.3a) and deepened and widened the valley's bedrock bottom. Two older glaciers flowed across the New York City region from NW to SE transporting and distributing indicator stones and a distinctive reddish till from New Jersey over the New York-Long Island region (Figure A.3b). The next oldest glacier followed the same course as the youngest one - from NNE to SSW. (See Figure A.3a.) These glaciers nearly obliterated evidence for the oldest glacier which flowed from the NW to SE. (See Table A-3.)

Of special interest in the interpretation of any glacial deposit is the direction in which the glacier flowed. This can be determined by studying the marks made by the flowing ice on the bedrock and by tracing distinctive boulders (known as **erratics**) transported by a glacier back to the parent bedrock areas from which they were eroded. Numerous distinctive kinds of rocks are found in the New York metropolitan region. One particularly distinctive variety is a coarse-textured igneous rock found only near Peekskill. Blocks of such Peekskill-derived rocks have been found at Orchard Beach; they confirm that the latest glacier flowed from NNE to SSW.

While all that ice was spreading as far south as New York City, it was steadily robbing the ocean of water. As a result, sea level dropped and the exposed margins of the continents "grew" outward. (See Figure A.1.) When the ice melted, the water returned to the oceans and sea level rose. The repeated waxing and waning of glaciers and rise and fall of sea level took place many times. Unfortunately, in the New York City region, not much evidence remains to enable us to decipher how all the many past glaciations affected the local area. As with Layer 6, Layer 5 lies in contact with all older layers. This indicates that the great period of erosion that exposed all these older layers is older than Layer 5.

Layers 4 and 3: The Miocene(?) to Cretaceous Coastal-Plain sediments. Time span: from about 10 Ma to about 100 Ma; a long period of marine sedimentation on the submerged continental margin.

Perhaps you have just noticed that the amounts of time involved in each of our layers is growing by leaps and bounds as the layer numbers decrease by 1. Part of this time growth is real and part of it is an artifact of our ignorance and the progressive decrease in the amounts of geologic record preserved as one pushes backward into it.

The coastal-plain strata are the first of the layers in which the plate-tectonic setting affects the accumulation of sediments. The coastal-plain strata are typical products of the continued seaward tilting of the passive margin of a continent.

Layer 4 (about 10 Ma; Miocene?) is distinctive because it consists of layers of gravel that were spread out as fans eastward from the ancestral Appalachians. These gravels drove the shoreline eastward and in many areas since that happened, the sea has not returned.

Layer 3 (15 Ma to 110 Ma) refers to the generally marine coastal-plain sediments. Although we list them here as consisting of only two layers, they really form a complicated assemblage of materials. Nevertheless, the general conditions that prevailed while the coastal-plain strata were accumulating remained generally constant.

To show some of the diversity, we can consider two sublayers of Layer 3, designated A (for the lower and older) and B (for the upper and younger). In a slight departure from the "top-down" approach followed to this point, we discuss the lower, older sublayer first.

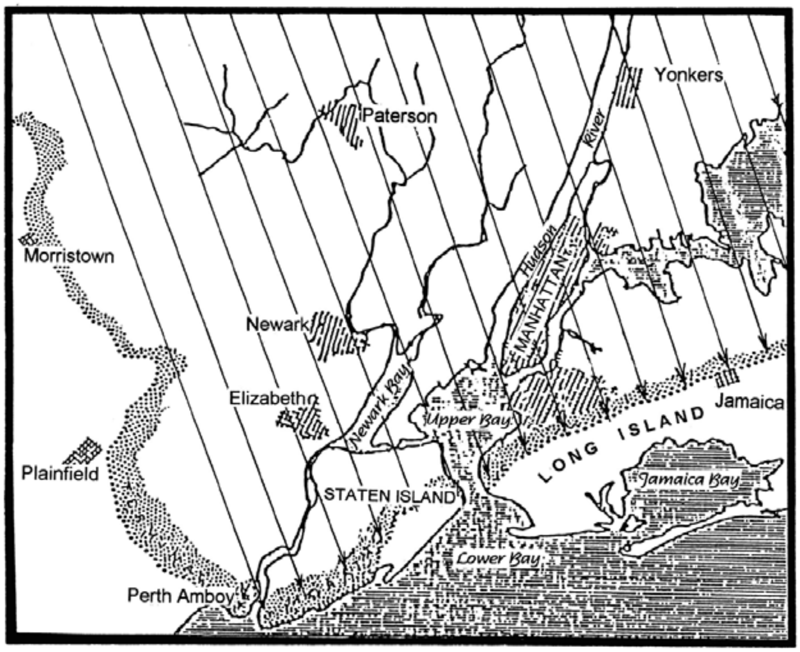


Figure A.3a - Rectilinear flow from NW to SE of glacier older than the latest Wisconsinan. This glacier flowed across the Hudson Valley and deposited red-brown till and -outwash on the east side of the Hudson River. (J. E. Sanders.)

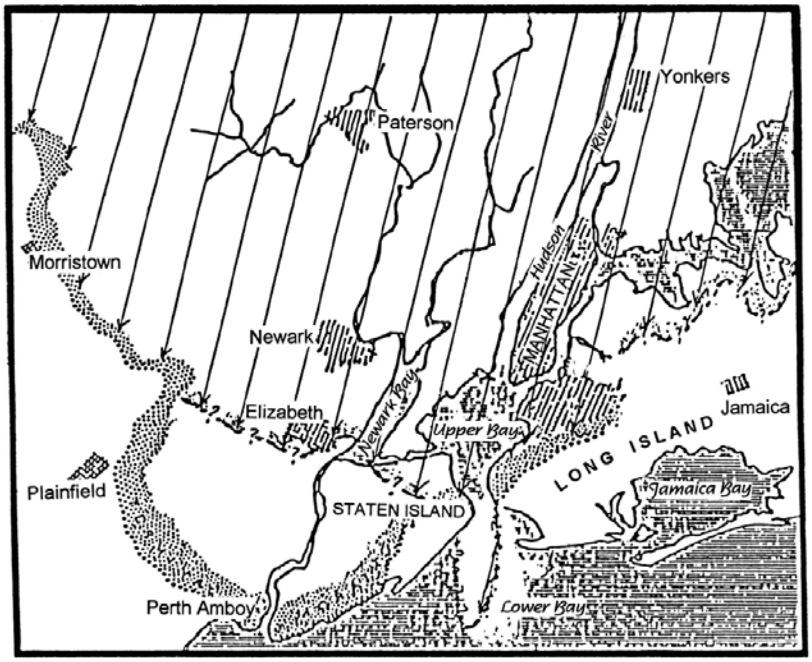


Figure A.3b - Inferred flow pattern of latest Wisconsinan glacier, the Woodfordian, down the Hudson and Hackensack lowlands from NNE to SSW. As shown here, the Woodfordian glacier affected only westernmost parts of Long Island; elsewhere, its terminal moraine was along the south coast of Connecticut. (J. E. Sanders.)

Coastal-plain Sublayer A: (age Late Cretaceous; time span about 100 Ma to 65 Ma) consists of sands and related sediments that accumulated on a great delta complex. We do not yet know all the details, but some combination of ancient rivers draining the NE part of the USA fed into the sea in the general vicinity of New York City. South of the delta complex, the shoreline of the time trended nearly E-W. We can reconstruct this from the fact that both to the SW and to the E of the delta, the sediments that are of the same age as the sediments of the delta are of marine origin and represent the kinds typically deposited well away from the shoreline.

The thickness of coastal-plain sublayer A is about 1,000 feet. This layer is exposed in parts of New Jersey, on Staten Island, and in a few places on Long Island. It underlies all of Long Island and there serves as a very important source of underground water. Hundreds of deep wells draw fresh water from the sand layers (named the Magothy and Lloyd Sand) of this sublayer. If these sandy layers were not present, the number of Long Islanders would be much smaller than it is; the water supply would not have been large enough to support a large population.

Coastal-plain sublayer B: (age Miocene; time span 20 Ma to about 10 Ma) lies on the eroded upper surface of coastal-plain sublayer A. Sublayer B, only a few hundred feet thick, consists of sediments deposited well offshore on a shelf much like the present continental shelf lying east of the shoreline along the US east coast today. The characteristics of sublayer B and its relationships to sublayer A permit us to infer that crustal movements took place after sublayer A had been deposited and before sublayer B accumulated. The New York region was first relatively uplifted (shown by the erosion of sublayer A) and then resubmerged to a greater extent than previously (to enable offshore-type marine strata of sublayer B to overlie the eroded delta deposits of sublayer A).

Nearly all of the coastal-plain layers consist of still-"soft" sediments. They have been cemented only locally if at all, and have not been deeply buried nor tilted by more than a few degrees from their initial horizontal positions. The general shape of the shoreline of the New York Bight originated as a result of the gentle tilting of the coastal-plain layers.

The elongate E-W depression forming what is now Long Island Sound was excavated by streams during this post-sublayer B period of erosion. The Long Island Sound depression lies adjacent to and just north of the preserved edge of sublayer A. We think also that much of the pattern of present-day rivers and valleys was established during this same erosion interval.

Layer 2: Triassic- to Jurassic sedimentary- and igneous rocks of the Newark basin. Time span: from 180 Ma back to about 220 Ma; rocks related to the rifting and opening of the Atlantic ocean.

With Layer 2, we get down not to brass tacks but to something everyone will recognize as bedrock. What is more, we come to the rocks that form some of our region's most-prominent landscape features. Rocks of Layer 2 extend westward from the Hudson River (south of Stony Point) to the Ramapo Mountains in NW New Jersey (Figure A.4). The thickness of the strata of Layer 2 is many thousands of feet. All parts of Layer 2 have been bent into folds. In addition, the strata are now inclined about 10° to 15° to the NW.

The Hudson River has eroded a valley along the base of Layer 2 that extends from Stony Point, NY to Hoboken, NJ (opposite the Lincoln Tunnel). This stretch of the Hudson Valley is geologically analogous to Long Island Sound. In both, one side is formed by ancient, eroded metamorphic rocks, and the other side, by strata that dip away from the older rocks and whose basal part contains a resistant layer that underlies a linear ridge. In the stretch of the Hudson Valley referred to, the strata of Layer 2, dipping 10° to 15° to the NW, form the W side and the Palisades is the ridge formed by the basal resistant layer.

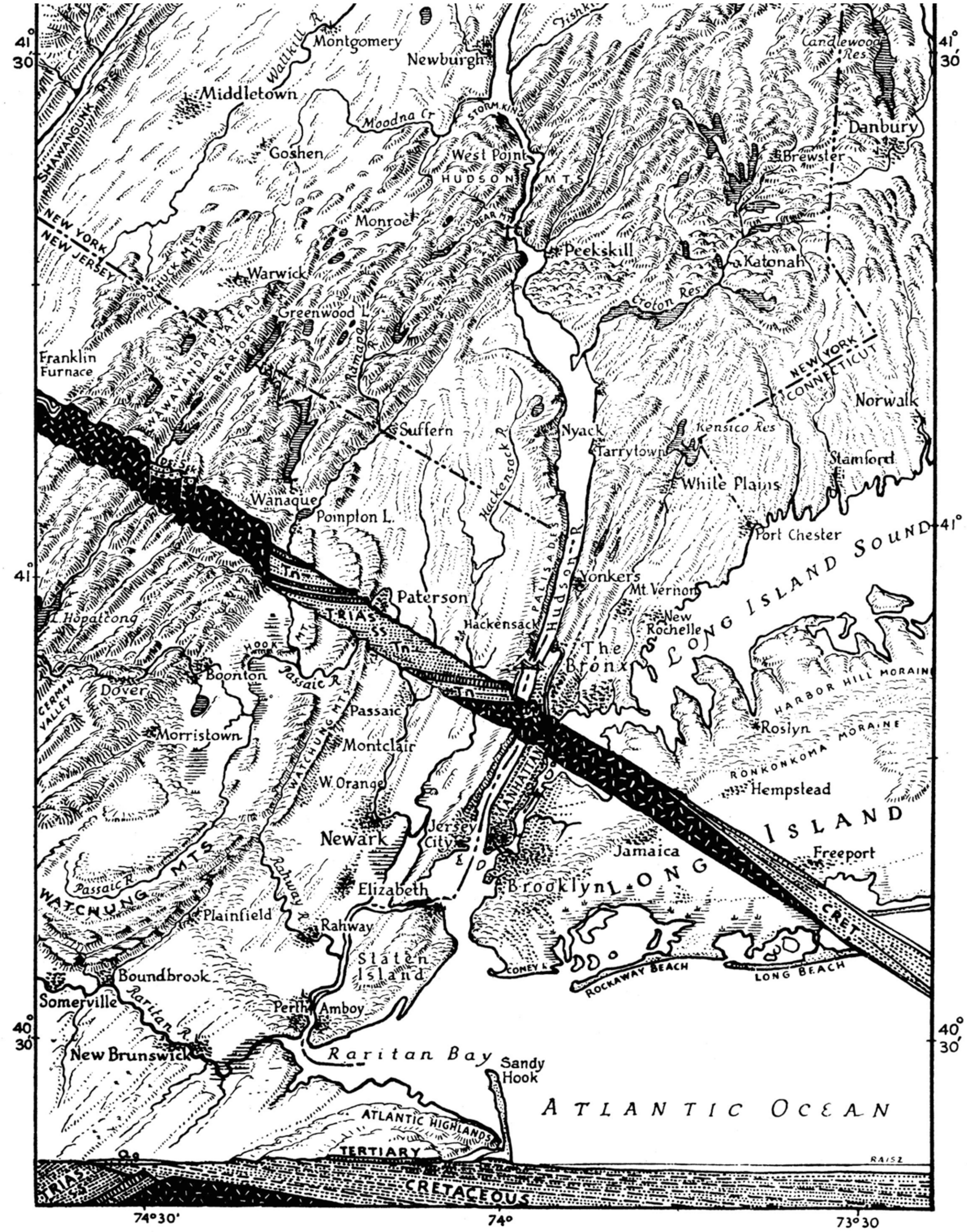


Figure A.4 - Physiographic diagram of New York City and vicinity showing two cut-away vertical slices to illustrate the geological structure. (Erwin Raisz.)

The strata of Layer 2 were deposited in lakes and on vast river plains that occupied part of an ancient rift valley analagous to the East African rift system of today. Typically, the lake deposits are black; some have yielded fine specimens of carbonized fossil fish. From the overlying red-colored ancient river deposits have come footprints and bones of dinosaurs.

Debate continues over the interpretation of the ancient East American rift valley in which the rocks of Layer 2 accumulated. Numerous similarities can be cited between the characteristics of Layer 2 and the materials that are now filling the rift valleys of East Africa. Whether or not these two valley settings were identical, no one disputes the point that Layer 2 originated in a nonmarine valley during a time of active vertical movements of great blocks (or huge plates) of the Earth's lithosphere. Some of these local blocks sank actively. On them, strata that became thousands of feet thick accumulated. Nearby blocks were persistently elevated; from them, great quantities of coarse debris were dumped onto the sinking blocks. This debris now exists as coarse conglomerates (lithified gravel beds).

According to modern ideas about the behavior of lithosphere plates, when two plates diverge, the lithosphere is stretched, thinned, and eventually may split (Figure A.5). If such total rupture takes place, then new material wells upward from below along the new fracture, and the conditions are established for starting a new ocean basin (in this case the modern Atlantic). The earliest feature formed in such plate divergence, one that appears before the lithosphere is totally ruptured, is a rift valley. If the forces causing the plates to move apart cease to operate before the point of total rupture, then the only feature formed may be a rift valley. It's a long way from a rift valley to an ocean and the ocean-forming chain of events does not always go to completion. (In today's world, narrow basins such as the Gulf of California or the Red Sea-Gulf of Aqaba-Dead Sea depression in the Middle East are taken to be examples of ocean basins in early stages of development.) Granted all the above, then the East American rift valley of Early Mesozoic age might logically be entitled: "The Little Ocean That Couldn't" (continue to open, that is). As far as the fossils indicate, no ocean ever entered the low parts of the basins in which the strata of Layer 2 were deposited.

The rocks of Layer 2 display a second striking characteristic not present in any of the layers examined so far: igneous rocks. At least three times, molten material derived from within the Earth's mantle moved upward and was spread out on the ancient Earth's surface as vast lava flows. The layers of igneous rocks formed by the cooling of these great lava flows are so uniform through such wide areas that we must conclude that in the ancient valley areas, vast lava lakes formed. Their depths ranged from 350 feet to close to 800 feet. After the lava in each lake had cooled and solidified, streams of water spread new layers of sediment across the then-cooled igneous rocks. Between each of the separate lava outbursts, sediments accumulated and became as thick as 1,000 feet (300 meters), so that over time the entire basin filled with strata that are as much as 8 km thick in some places.

After all the layers had been tilted and much eroded, these layers of igneous rock project higher than their surroundings. In New Jersey, the Watchung Mountains are examples of such tilted ancient lava flows whose tilted and much-eroded resistant edges form ridges. The lowlands between the Watchung ridges are underlain by the relatively weaker sedimentary layers composed of the debris that buried each of the ancient lava flows.

Beginning at the time of the first of these ancient lava flows and possibly extending for as long as the time required for the second flow to appear, an enormous sheet of igneous material (called the Palisades Intrusive Sheet) inserted itself between the sedimentary layers near the base of Layer 2. Because this sheet of molten material never reached the Earth's surface, the igneous material forming this sheet is not considered to have been a lava flow. Instead, the molten material cooled roughly 3 to 4 km underground, steaming and baking the overlying layers of enclosing sedimentary material that formed the roof of the tabular chamber in which the hot magma cooled (Merguerian and Sanders, 1995 a,b). After tilting and much erosion, this sheet, too, formed a prominent ridge--the Palisades which forms the west bank of the Hudson River from Haverstraw, NY to Hoboken, NJ. (See Figure A.4.) Great columns of blue-gray rock having a distinctive brownish weathering color have broken along the cracks that opened when the molten material shrank during cooling. The thickness of the Palisades sheet of igneous rock is 800 to 1,000 feet (approximately 300 meters), making it one of the thickest such bodies in the world.

The rocks of Layer 2 rest upon a sequence of much older, highly deformed and modified igneous- and metamorphic rocks of Layer 1. When we come to rocks of Layer 1, we have really reached "rock bottom." Nothing older is known in this vicinity.

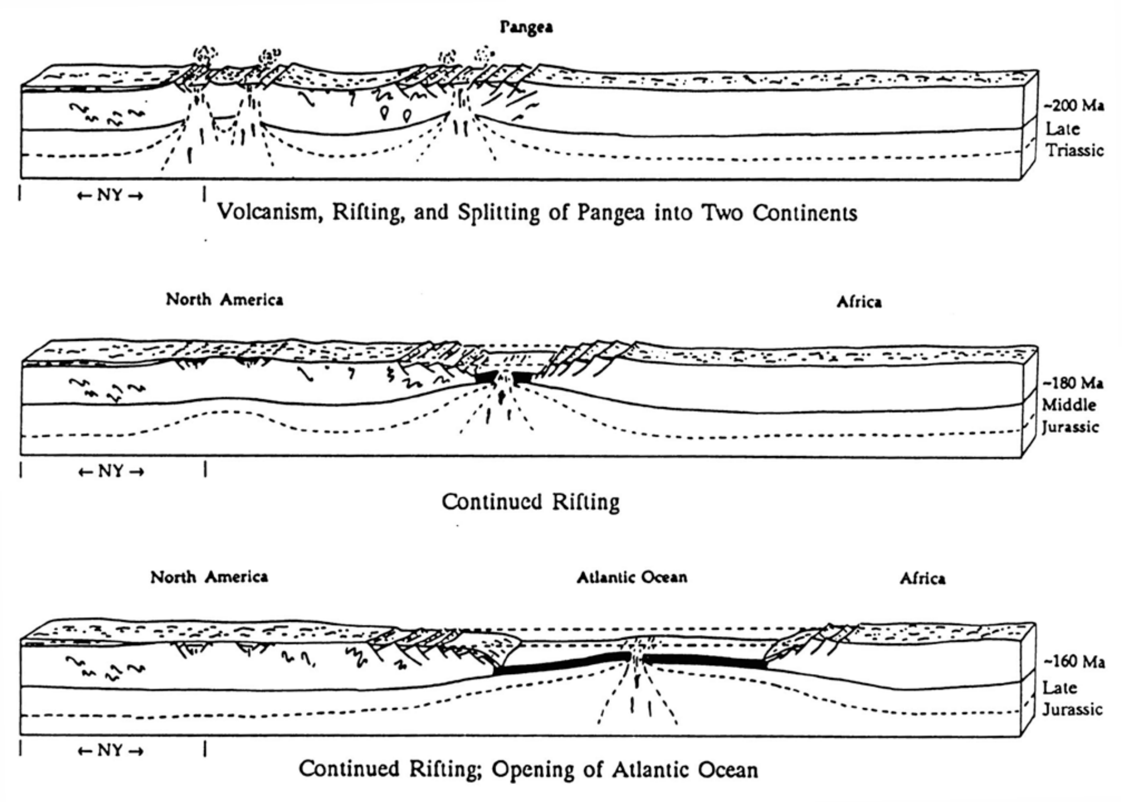


Figure A.5 - Block diagrams from Late Triassic to Late Jurassic time showing early rift stages of Pangea and subsequent opening of the modern Atlantic Ocean. (New York State Museum, Educational Leaflet 28, 1991, p. 259.)

Layer 1: Proterozoic and Paleozoic metamorphic- and igneous rocks "The Basement". Time span: 430 Ma back to 1,100 Ma; rocks formed at great depth during ancient continental collisions.

The rocks of Layer 1 have been much deformed and heated. As a result, their minerals have recrystallized. In their reconstituted form, they appear as shimmering mica schists, glistening marbles, striped gneisses, and dense amphibolites. The layers have been compressed into great folds; the sides of these folds are very steep so that the resulting layers are close to vertical. The rocks of Layer 1 are typical of those found in mountain chains. Perhaps the best way to characterize Layer 1 is to refer to it as the "roots" of a former mountain chain. If we could sort them out by recovering the full evidence of their ancestry, we would probably find that the sediments now converted into the rocks of Layer 1 were deposited on a continental margin that was first passive but later became active and convergent.

Layer 1 rocks underlie all of Manhattan Island, The Bronx, and are exposed in a few places on Staten Island and along the west bank of the Hudson River, in New Jersey, from Hoboken to Bayonne. (See Figure A.4.) East of Manhattan and The Bronx, Layer 1 lies directly beneath Layers 6, 5, or 3. Beneath the Hudson River, rocks of Layer 1 are overlain by those of Layer 2. In fact, as noted above, the Hudson follows the boundary between these two great layers from Haverstraw, in Rockland County, New York, to Hoboken, in Hudson County, New Jersey.

The rocks of Layer 1 had been deformed and recrystallized long before those of Layer 2 were deposited. In fact, measurements of radioactive decay products, such as those that began to accumulate in the crystal lattices of Layer 1 rocks after their most-recent episode of heating, enable us to assign an age of 350 Ma to the last major episode of metamorphism of the rocks in the New York region. Notice that the age of origin of the precursor rocks of the youngest rocks of Layer 1 is thought to be about 450 Ma. Thus, in some places, the rocks of Layer 1 were heated to the point of recrystallization about 100 million years after they had accumulated as sediments.

In eastern Connecticut, rocks belonging to the same "basement complex" as New York's Layer 1 went back into nature's pressure cooker for another treatment about 230 Ma. During any time when the original sedimentary material was recrystallized, various igneous rocks were also formed. Local bodies of both mafic rocks and granites have been discovered. In addition, ancient volcanic rocks were formed, but the heat and pressure so changed the volcanic rocks that amateur observers would scarcely recognize them as being of volcanic origin.

Our studies of modern- and ancient mountain belts suggest that the rocks of Layer 1 were produced in a deep-sea trench formerly bordering the east coast of ancestral North America (Robinson and Hall, 1980; Merguerian 1983, 1996). The driving force behind the mountain building, which brought sediments and volcanic materials down to depths exceeding 12 miles, were a number of dominantly convergent Paleozoic plate-tectonic interactions that closed the ancestral proto-Atlantic ocean called **Iapetus.** The mountain building involved collisions between intervening volcanic island chains (Taconic arc), microcontinents (Avalonia), and continents (proto Africa) and ultimately resulted in creation of a supercontinent known as **Pangea** (Figure A.6). The oldest rocks in the region (1,100 Ma Proterozoic rocks found exposed in the Bronx, Westchester, and central New Jersey) were initially formed during an even older series of convergent plate interactions (known as the **Grenville Orogeny**) for which we have very little evidence to reconstruct these events.

Yet, for all their ancient appearance and venerable ages, rocks of Layer 1 are no older than about 1,100 Ma. Thus, they reach back in time to a point marking only the last 20 percent or so of the Earth's history. All ideas about the other 80 percent of the earlier history of the Earth have been derived from localities outside the New York metropolitan area, such as in the northwestern part of the Canadian Shield, where it is possible to examine "Grand-daddy" rocks as old as about 4,000 Ma.

Regional Geological Synthesis

Now that we have reached all the way to the bottom of the stack of layers in New York City and vicinity's geological "cake," let us recapitulate by working our way upward. At the base, we find ourselves just about at the point where we started at the top--not in the soup, but in the sea. The materials of Layer 1 as we have defined it are composite and are products of several episodes of deposition of sediments in an ancient sea, of deformation and heating of these marine sediments, and of uplift and erosion. As mentioned, the most-ancient materials of Layer 1 began to accumulate before 1,100 Ma and the most-recent ones, about 400 Ma. Mountain building began roughly 450 Ma and by about 350 Ma, the New York region may have resembled the Alps or other great modern mountain chains.

As soon as they had formed, New York City's ancient mountains began to be eroded and were eroded continuously for a period lasting about 150 Ma. After this long episode of erosion, the local lithosphere, which had been so plastic when the rocks of Layer 1 were being deformed and recrystallized, became brittle and began to split apart along great cracks. Roughly 200 Ma, a great rift-valley system formed, and a new cycle of sediment accumulation began. This time, the sea did not enter the Newark basin where the sediments were being deposited. The environments of deposition of the sediments were nonmarine; they included fans, river-bottom plains, and large lakes whose levels fluctuated with long-term changes in the climate (Olson, 1980). At least three times while the Newark basin was being filled by sediments washed in from the adjacent highlands, volcanic activity took place. Lava oozed out of cracks and spread over the basin floor to create gigantic lava lakes several hundreds of feet deep. The basin-filling materials of the Newark basin are represented by Layer 2. About 175 Ma, the Newark basin disappeared. The formerly horizontal layers of its filling strata were folded and tilted and another great episode of erosion commenced.

About 50 Ma later, during the Cretaceous Period, a passive continental margin was fully established and began to sink beneath the sea. The margin being submerged was the western margin of the newly formed Atlantic Ocean. For about 40 Ma, our area bobbed up and down, but probably never extended very far below nor very high above sea level. Deposition took place on an ancient shelf at variable distances from the ancient shoreline. The result was the coastal plain strata of Layer 3.

Starting about 10 Ma, two important changes took place. (1) The Appalachian region was greatly elevated. A sheet of coarse fan sediments was spread eastward away from this newly elevated tract. The climate may have been semi-arid or even arid. The complex mass of Miocene fan sediments, which pushed back the sea, formed Layer 4. (2) Sea level dropped greatly so that rivers began to erode deep valleys; the depression that is now Long Island Sound was excavated. This great drop in sea level probably was related to climatic oscillations that are still in progress. During much of this interval, the New York City region continued to be eroded. Our present pattern of valleys originated. Eventually, however, continental glaciers spread over much of our area from two general directions. The oldest- and intermediate ice sheets advanced from central Canada and crossed the Hudson Valley on a SE course. An old glacier and the youngest glacier originated from northeastern Canada and flowed toward the

SW, down the Hudson Valley. The glaciers left behind superposed irregular blankets of sediment as Layer 5 and associated features of glacial erosion.

After the last glacier finally had melted away and all the water that had been locked up on land as ice had finally returned to the oceans, sea level rose to its present position. Along the margins of the present Atlantic Ocean, we can see a new layer of marine sediment being deposited. The sea and its deposits form Layer 6. Tables A-1, A-2, and A-3 summarize the layers and their formative geological relationships.

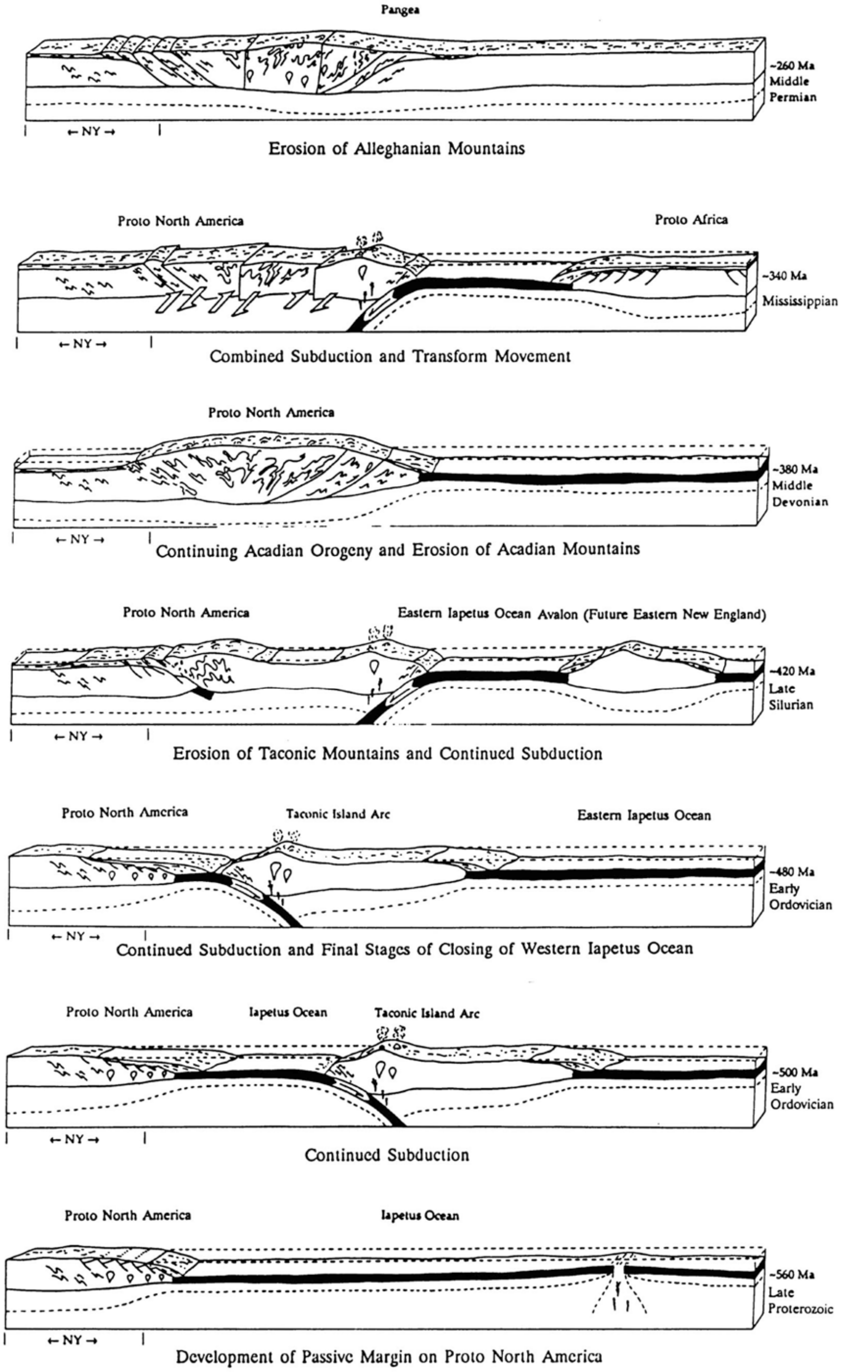


Figure A.6 - Sequential block diagrams from Late Proterozoic to Middle Permian time showing the plate tectonic development of the Appalachian mountain belt. (New York State Museum, Educational Leaflet 28, 1991, p. 256-258.)

TABLE A-1 - GEOLOGIC TIME CHART

with selected major geologic events from southeastern New York and vicinity

Eras	**Periods (Epochs)**	**Years (Ma)**	**Selected events**
CENOZOIC			
	(Holocene)		Rising sea forms Hudson Estuary, Long Island Sound, and other bays.
		0.1	Barrier islands form and migrate.
	(Pleistocene)		Melting of last glacier forms large lakes. Drainage from Great Lakes overflows into Hudson Valley. Dam at The Narrows suddenly breached and flood waters erode Hudson shelf valley.
			Repeated continental glaciation with glaciers flowing from NW and NE form moraine ridges on Long Island.
		1.6	
	(Pliocene)		Regional uplift, tilting, and erosion of coastal-plain strata; sea level drops.
		6.2	Depression eroded that later becomes Long Island Sound.
	(Miocene)		Fans spread E & SE from Appalachians and push back sea. Last widespread marine unit in coastal-plain strata.
		26.2	
	(Oligocene) (Eocene) (Paleocene)		
		66.5	
MESOZOIC	Cretaceous		Passive eastern margin of North American plate subsides and sediments (the coastal plain strata) accumulate.
		96	
		131	***(Passive-margin sequence II).***

	Jurassic		Baltimore Canyon Trough forms and fills with 8,000 feet of pre-Cretaceous sediments. Atlantic Ocean starts to open. Newark basins deformed, arched, eroded.
		190	Continued filling of subsiding Newark basins and mafic igneous activity both extrusive and intrusive.
	Triassic		Newark basins form and fill with nonmarine sediments.
PALEOZOIC			Pre-Newark erosion surface formed.
	Permian		***Appalachian orogeny.***
		260	(Terminal stage.) Folding, thrusting, and metamorphism of Rhode Island coal basins; granites intruded.
	Carboniferous		
			Southeastern New York undergoes continued uplift and erosion.
	Devonian	365	***Acadian orogeny.*** Deep burial of sedimentary strata. Metamorphism in New York City area. Peekskill Granite intruded.
	Silurian		
		440	***Taconic orogeny.*** Intense deformation and metamorphism.
		450	Cortlandt Complex and related mafic igneous rocks intrude Taconian suture zone. (Cameron's Line).
	Ordovician		Great overthrusting from ocean toward continent. Taconic deep-water strata thrust above shallow- water strata. Ultramafic rocks (oceanic lithosphere) sliced off and juxtaposed with deposits of continental shelf. Shallow-water clastics and carbonates accumulate in west of basin (= protoliths of Lowerre Quartzite, Inwood

Cambrian		Marble, lower part of Manhattan Schist Formation). Deep-water terrigenous silts form to east. (= Taconic sequence; protoliths of Hartland Formation, parts of Manhattan Schist). ***(Passive-margin sequence I).***
PROTEROZOIC	570	Period of erosion.
Z	600	Rifting with rift sediments (Ned Mountain Formation); Pound Ridge and Yonkers gneiss protoliths.
Y	1100	***Grenville orogeny.*** Sediments and volcanics deposited (protolith of Fordham Gneiss).
ARCHEOZOIC 2500	No record in New York.	
	4600	Solar system (including Earth) forms.

TABLE A-2 - GENERALIZED SUMMARY OF GEOLOGIC LAYERS, NEW YORK AREA

LAYER NO.	AGE (years) (Ma = millions of yr.)	GEOLOGIC AGE	CHARACTERISTICS
6	0 to 10,000+	Holocene	Sea water, junk, beach sand, marsh deposits, bay-bottom silts, etc.
5	10,000 to 100,000+ (could be older)	Pleistocene	Deposits from continental glaciers. Top layer (yellowish brn.) poss. 13,000 yr old, deposited by ice flowing SSW, down Hudson Valley. Older layers (reddish brn. in SE NY) deposited by glaciers that flowed to SSE, across the Hudson Valley. Older glacial sediments contain decayed granite stones.
	~~~Period of great erosion; rivers eroded deep valleys~~~		
4	ca. 10 Ma	Pliocene (?)	Yellow gravels, sands, and clays, spread SE from Appalachians on great fan complex; the climate probably was arid.
3	10 Ma to 110 Ma	Miocene to Late Cretaceous	Marine sands and clays of coastal plain; sediments of subsiding passive continental margin.
B	10 Ma to 20 Ma	Miocene	Continental-shelf and nearshore sands & silts that overlap farther inland than U. Cret.
A	l00 Ma to 65 Ma	Late Cretaceous	Deltaic sands, silts, & clays in NY City area; open-shelf sediments in other areas; gentle dip seaward at ca. 50 ft/mi.

~~Period of deformation of Newark rocks & great erosion~~

2	180 Ma to 200 Ma	Late Triassic to Early Jurassic	Newark Supergroup; nonmarine filling of Newark basin (deposited on fans, river plains, and lakes whose levels fluctuated cyclically with climate change); prevailing color is reddish brown (but some deep-lake deposits are black and associated mafic igneous rocks are dark bluish gray). Three extrusive sheets underlie the Watchungs; an intrusive sheet forms the Palisades. In NJ, the strata dip 15˚ NW

Period of great erosion; formation of pre-Newark age peneplain

1	350 Ma to 1150 Ma	Paleozoic & Proterozoic	Schists, gneisses, marbles, serpentinite, and related metamorphic rocks; ancient sedimentary strata and underlying igneous complex that were metamorphosed several times, most recently at 350 Ma.

TABLE A-3 - NEW CLASSIFICATION OF THE PLEISTOCENE DEPOSITS OF NEW YORK CITY AND VICINITY. (Sanders et al., 1997.)

Age	Till No.	Ice-flow Direction	Description; remarks
Late Wisconsinan ("Woodfordian"?)	I	NNE to SSW	Gray-brown till in Westchester Co., Staten Is., Brooklyn, & Queens (but not present on rest of Long Island); Hamden Till in CT with terminal moraine lying along the S coast of CT; gray lake sediments at Croton Point Park, Westchester Co.
Mid-Wisconsinan (?)			Paleosol on Till II, SW Staten Island.
Early Wisconsinan(?)	II	NW to SE	Harbor Hill Terminal Moraine and associated outwash (Bellmore Fm. in Jones Beach subsurface); Lake Chamberlain Till in southern CT.
Sangamonian(?)			Wantagh Fm. (in Jones Beach subsurface).
	IIIA	NW to SE	Ronkonkoma Terminal Moraine and associated outwash (Merrick Fm. in Jones Beach subsurface).
Illinoian(?)	IIIB		Manhasset Fm. of Fuller (with middle Montauk Till Member; in lower member, coarse delta foresets (including debris flows) deposited in Proglacial Lake Long Island dammed in on S by pre-Ronkonkoma terminal moraine.
	IIIC		
Yarmouthian			Jacob Sand, Gardiners Clay.
Kansan(?)	IV	NNE to SSW	Gray till with decayed stones at Teller's Point (Croton Point Park, Westchester Co.); gray till with green metavolcanic stones, Target Rock, LI.
Aftonian(?)			No deposits; deep chemical decay of Till V.
Nebraskan (?)	V	NW to SE	Reddish-brown decayed-stone till and -outwash at AKR Co., Staten Island, and at Garvies Point, Long Island; Jameco Gravel fills subsurface valley in SW Queens.
			Pre-glacial (?) Mannetto Gravel fills subsurface valleys.

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Geol. 1C **Mineral Practicum** Your Name:

#	Luster	Hardness	Cleavage yes / no, number?	Other useful character streak, color, etc.	Mineral Name
1					
2					
3					
4					
5					
6					
7					
8					
9					
10					
11					
12					

Geol. 1C **Rock Practicum** Your Name:

#	Igneous Sedimentary Metamorphic	**Textures** and / or other features	**Minerals Present** two (or one)	**Rock Name**
1				
2				
3				
4				
5				
6				
7				
8				
9				
10				
11				
12				