GL01308

Mesozoic - Cenozoic Stratigraphy

1/4/80

Triassic Stratigraphy of the Western Interior

Tectonics and Paleogeography

a) Persistence of the Paleozoic tectonic pattern through the Early Triassic. Modified miogeosyncline with maximum subsidence in southeastern Idaho. 5000 ft.

b) Transcontinental Arch or "continental backbone" in evidence - never covered by marine deposits but mostly buried by continental sediments as the period progressed.

  1) Remnants of Ancestral Rockies slowly buried
  2) Enlargement of Southern Arizona high = Mogollon High

c) Mild expression of Williston basin.

d) Elevation of Mesocordilleran Highland in mid-Triassic - exclusion of seas from continental interior.

e) Increased importance of islands and sites of sedimentation along Pacific border. Klamathonia.

f) Restriction or narrowing of seas in Montana and Alberta. Possible addition of western Canada-Alaska by plate tectonics

g) Isolation of area of Colorado Plateau by marginal uplifts.

Sedimentation

a) Prevalence of continental sediments

  1) red beds, origin and significance
  2) aeolian sandstones late in the Period to Jurassic

b) Low carbonate content - possible lowest of any period, general absence of reefs.

c) Complex interfingering, of most sedimentary units particularily marine and non-marine.

d) Contributions of volcanic material - tuff, ash, lava.

e) Widespread gravel units - Shinarump, etc.

f) Evaporites.

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Boundaries

a) Lower Triassic-Permian, world wide significance.

b) Middle Triassic-Upper Triassic, relation to conglomerates and erosional breaks. 10 my Hiatus

c) Upper Triassic-Jurassic, relation to world-wide evidence, difficulties in aeolian stratigraphy.

Economic Products

a) Oil and gas - marine tongues in red beds. Lower Ꝑ main | a SS

b) Uranium, copper, vanadium and other metals in continentel sediments, Silver Reef, etc. Chinle

c) Non-metallic products - building and ornamental stones. ripple stone - Moenkopi

d) Host rock for Mineralization in Park City.

Paleontology

a) Invertebrates

1) ammonites world-wide zones - abt. 18

2) other molluscs, fresh water and marine poorly represented

3) brachiopods declined mostly lingula

4) scarcity of corals, sponges, crinoids and bryozoans

b) Plants - transitional floras, petrified forests

1) cycads

2) conifers ← Aurucaria (transitional from coal type to angiosperms (Cret.))

3) ferns

1st coal mine in U.S. was in Triassic of N. Carolina

c) Vertebrates

1) fish - Ganoids

2) amphibians Abundant

3) reptiles beginning

4) mammals beginning (Trytilodonts in Kayenta)

Scenic Areas

a) Petrified Forest National Monument Chinle

b) Canyonlands, Nat. Park Lake Powell, Rec. Ar Zion, Nat. Pk. Capital Reef, Nat. Pk. etc. Wingate, Kayenta

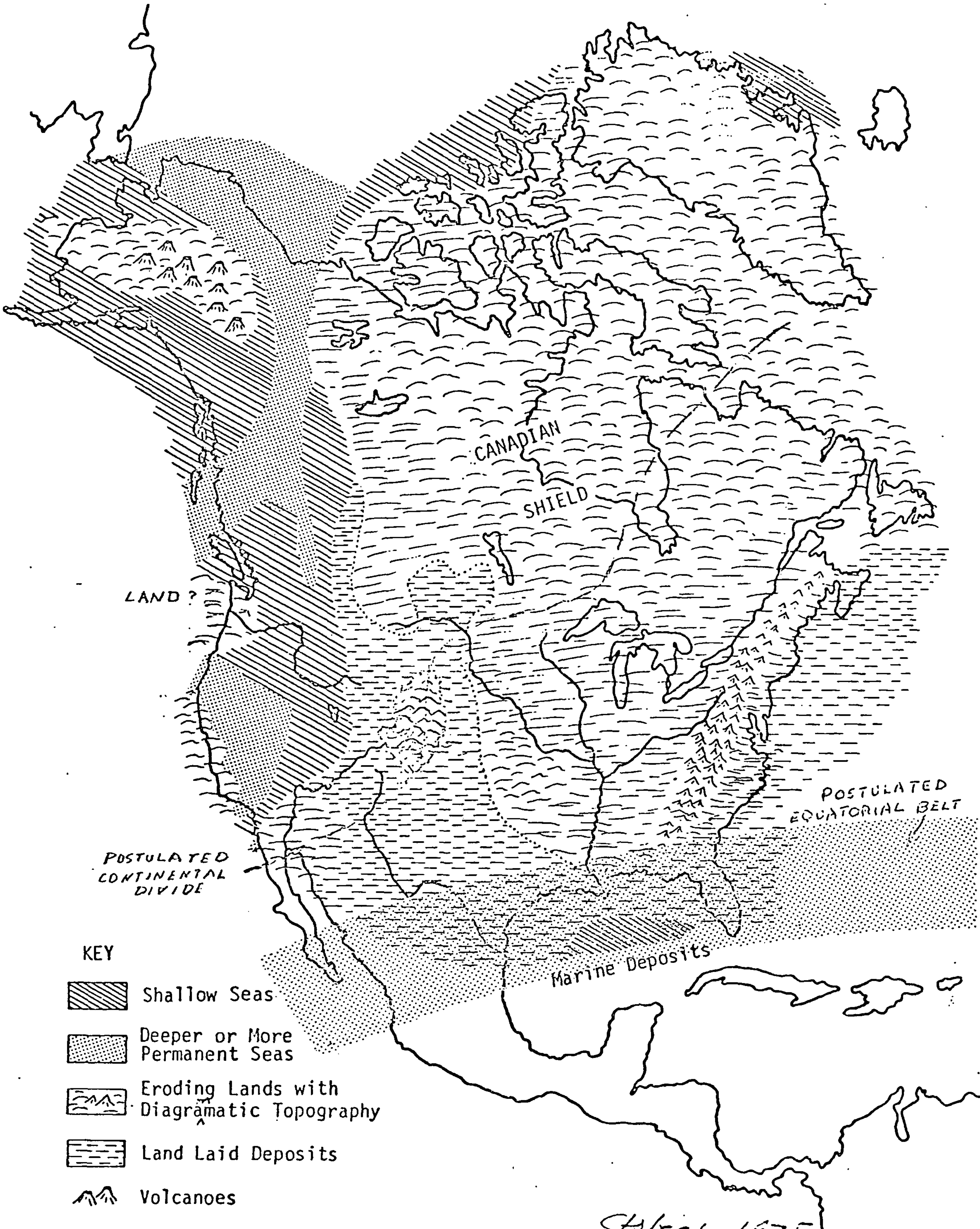
c) Painted Desert Chinle

d) Vermillion Cliffs, Echo Cliffs, etc. Vermillion Cliffs Moenkopi Group

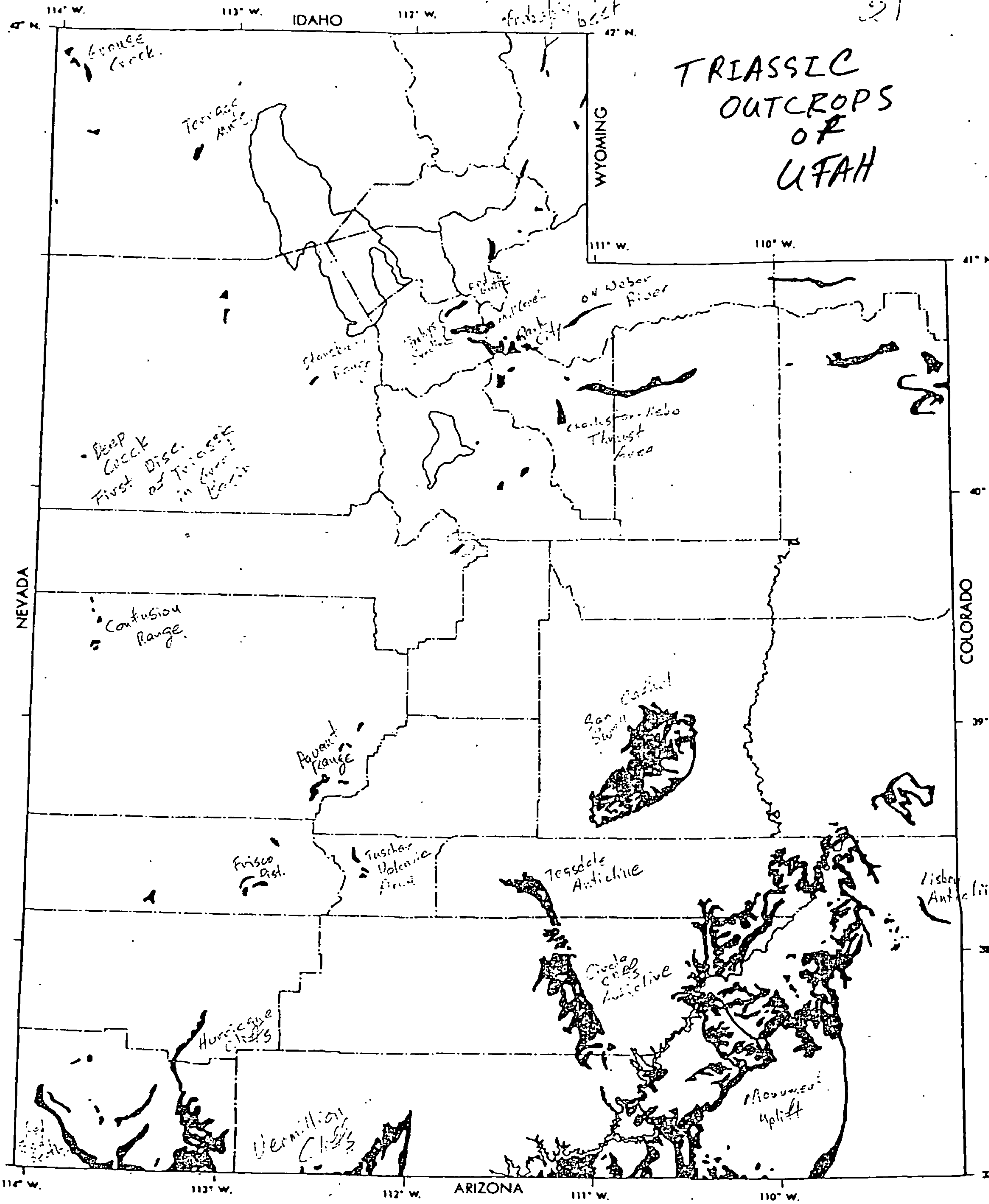
e. Flaming Gorge Moenkopi

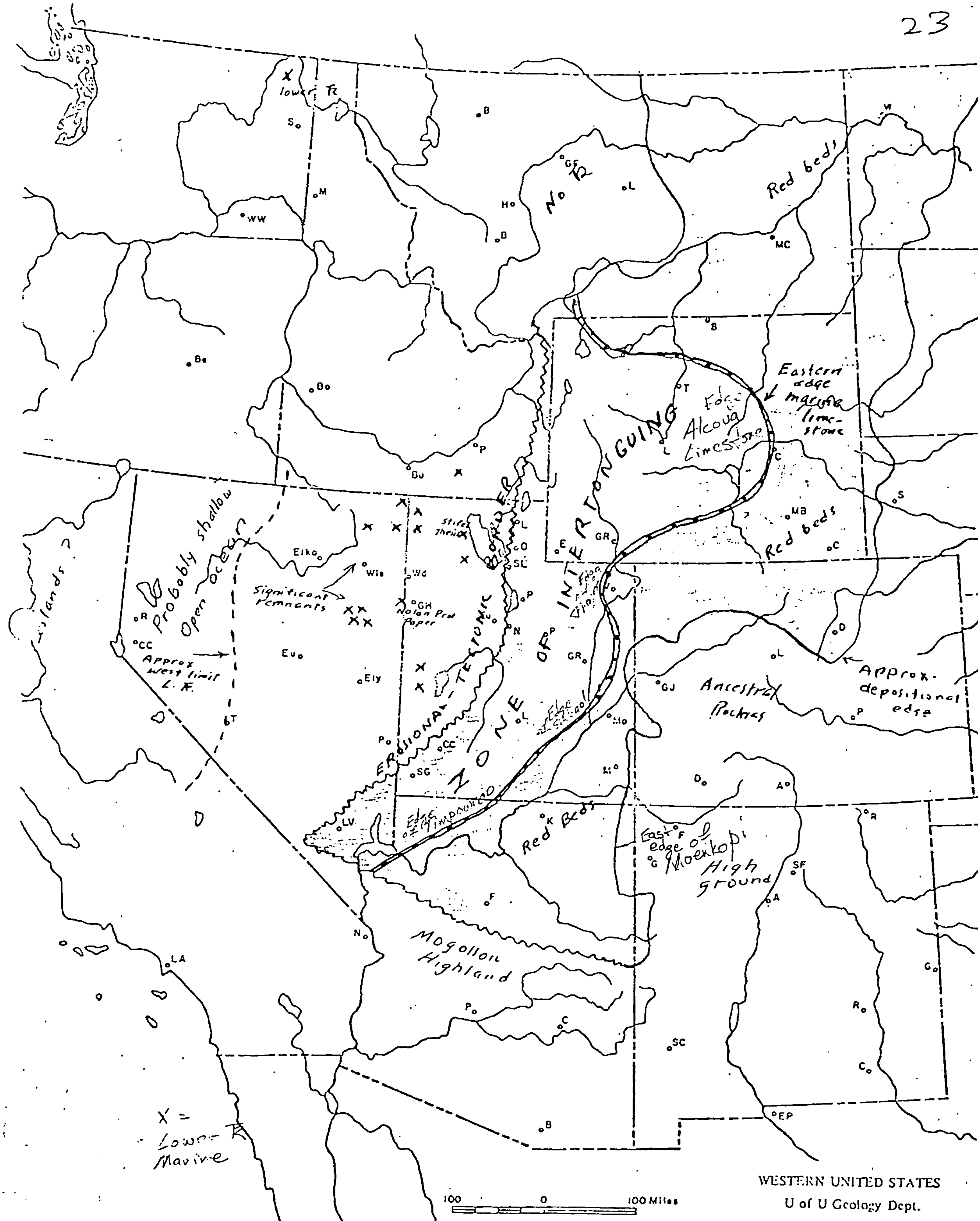
W. L. Stokes
1980

PALEOGEOGRAPHY OF THE TRIASSIC PERIOD

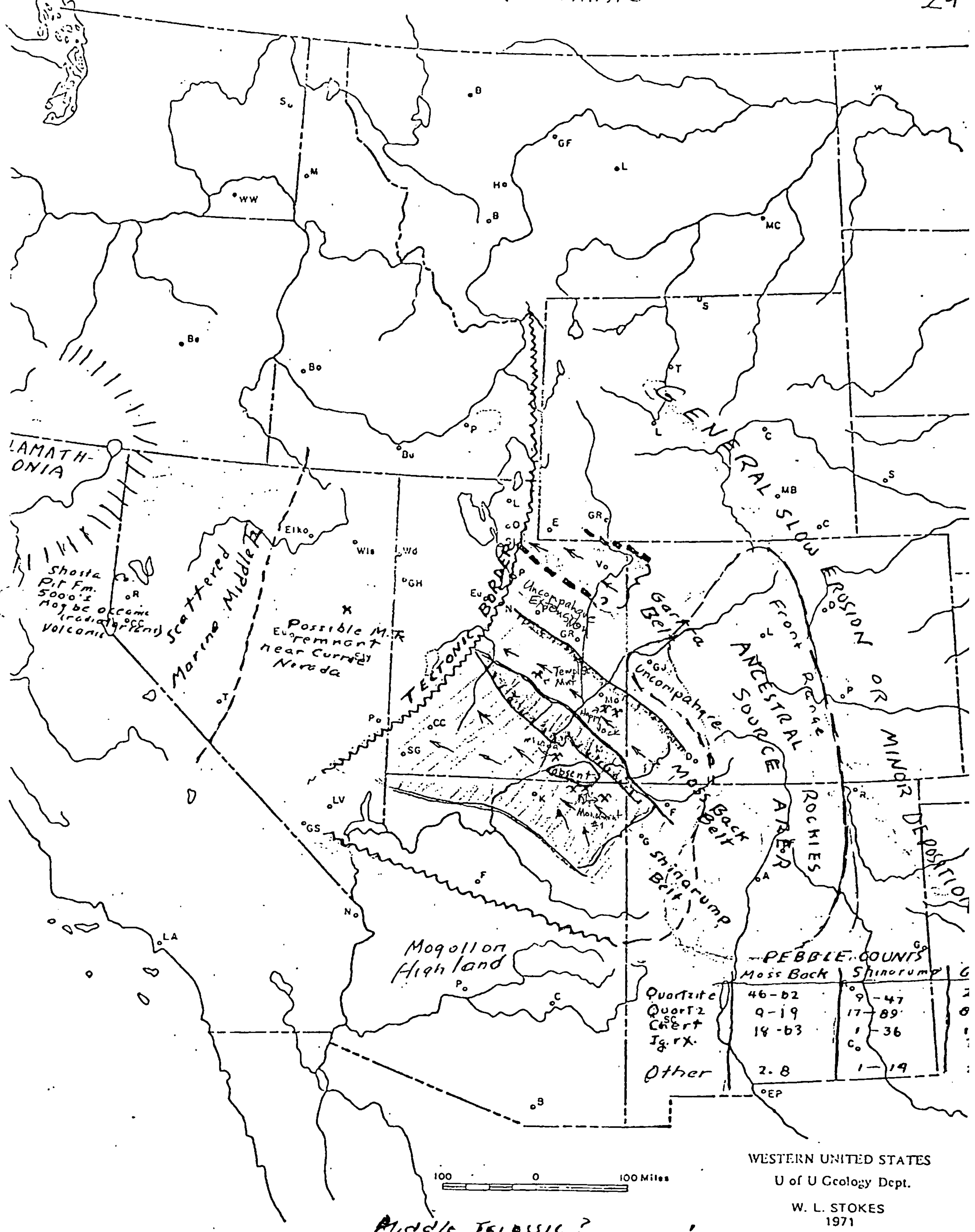


TRIASSIC OUTCROPS OF UTAH



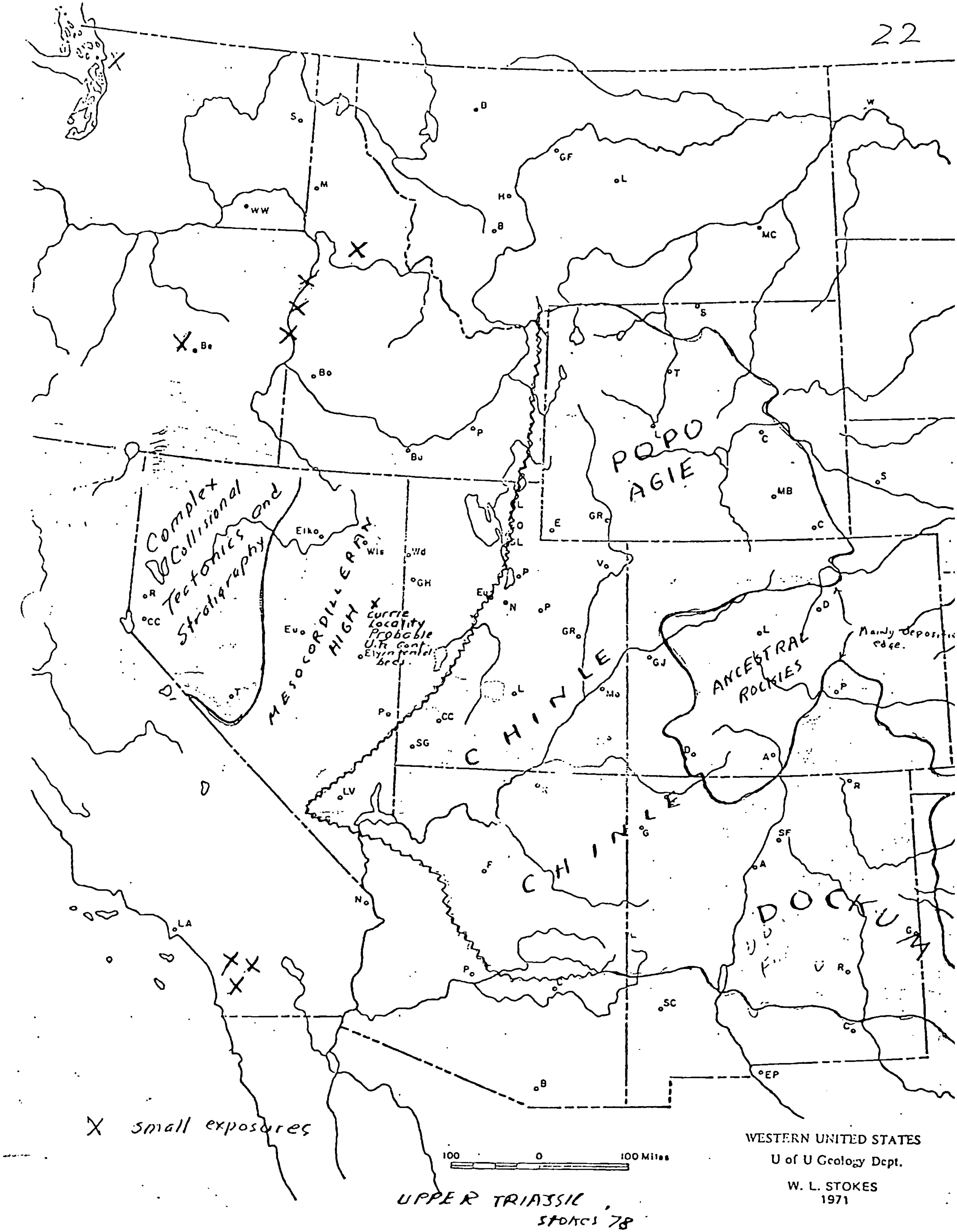


Lower TRIASSIC
Stokes' 78



Middle Triassic ?

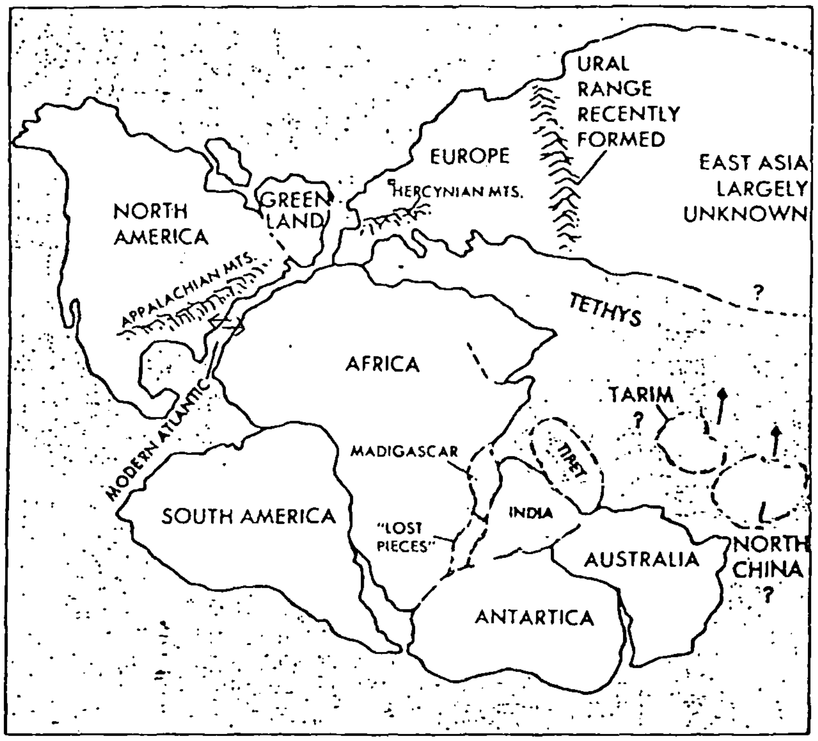
Stokes '78



Complex Collisional Tectonics and Stratigraphy
MESOCORDILLERAN HIGH
Currie Locality Probable U.Tr. Cont. Elyn feeder beds
POPO AGIE
CHINLE
CHINLE
ANCESTRAL ROCKIES
Mainly depositional edge
DOCKUM
Elko
X small exposures
100 0 100 Miles
WESTERN UNITED STATES
U of U Geology Dept.
W. L. STOKES
1971
UPPER TRIASSIC
Stokes '78

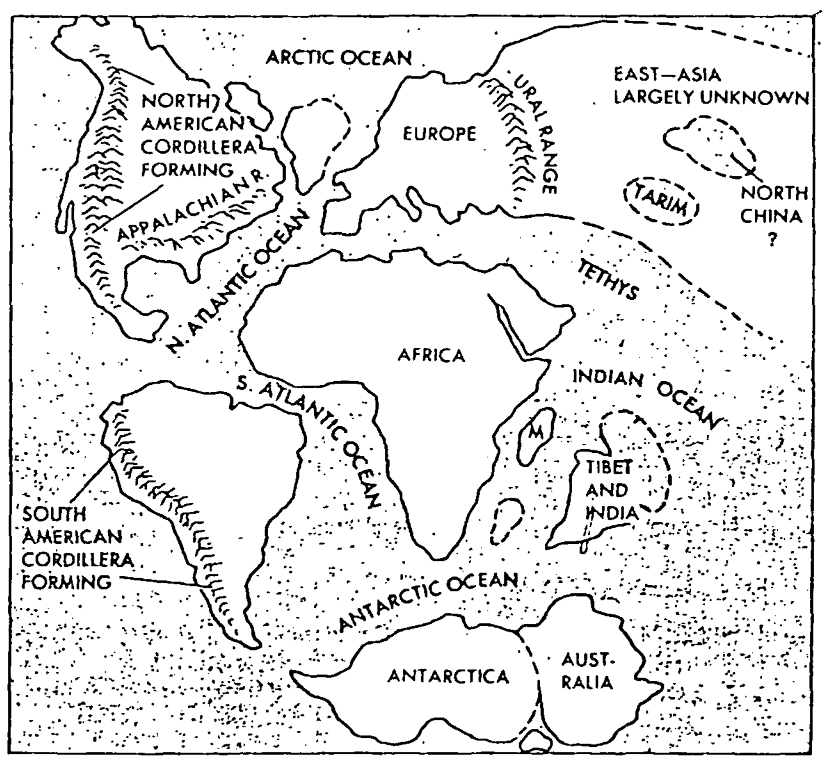
1/4/80

1



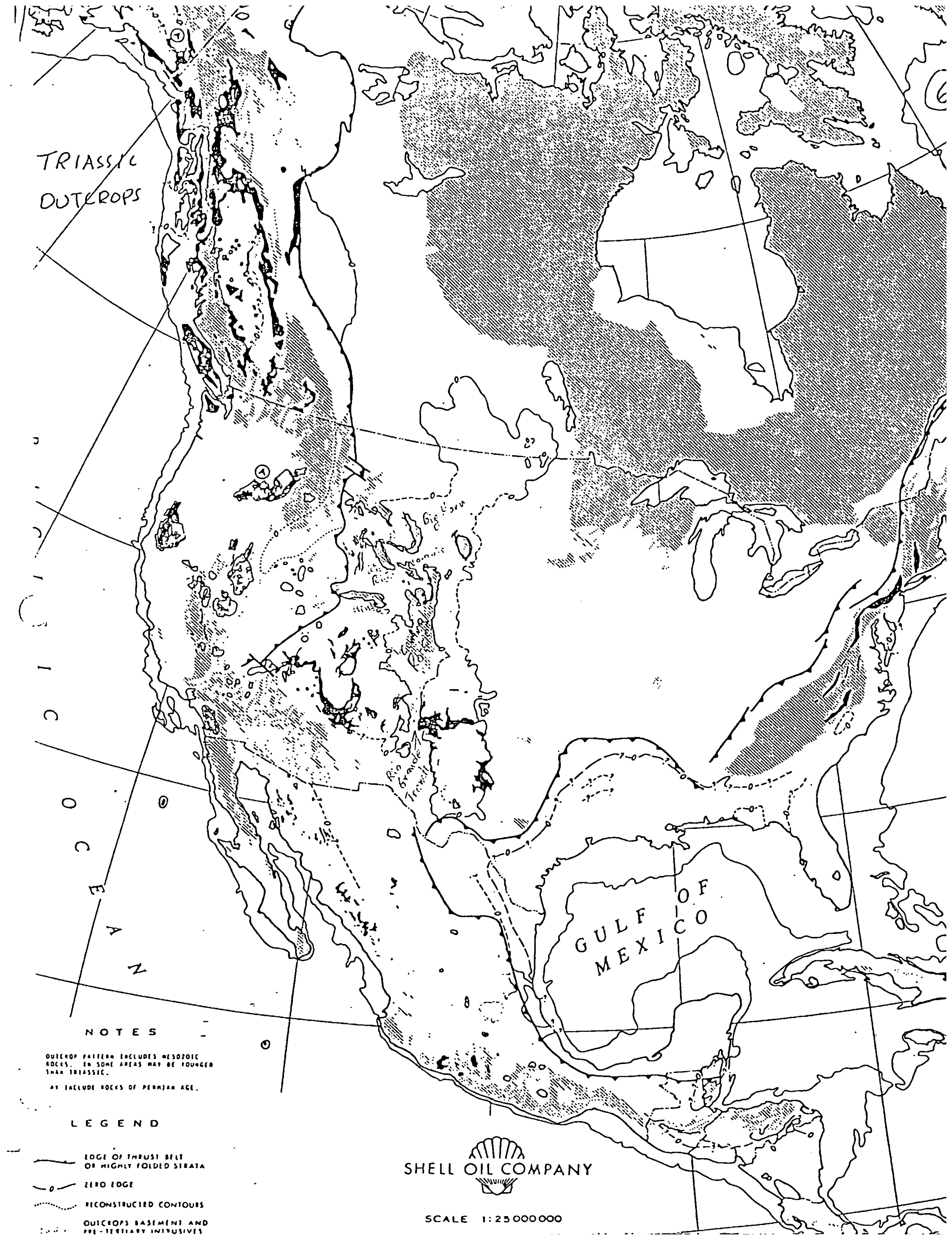
Diagrammatic
Arrangement
of Continents
Middle
Triassic

Distribution of major land-masses in the Middle Triassic.



Distribution of major land-masses near close of the Cretaceous period.

W. L. Stokes
1978



TRIASSIC OUTCROPS
Big Horn
Rio Grande Trench
O C E A N
GULF OF MEXICO
NOTES
OUTCROP PATTERN INCLUDES MESOZOIC ROCKS. IN SOME AREAS MAY BE YOUNGER THAN TRIASSIC.
AT INCLUDE ROCKS OF PERMIAN AGE.
LEGEND
EDGE OF THRUST BELT OR HIGHLY FOLDED STRATA
ZERO EDGE
RECONSTRUCTED CONTOURS
OUTCROPS BASEMENT AND PRE-TERTIARY INTRUSIVES
SHELL OIL COMPANY
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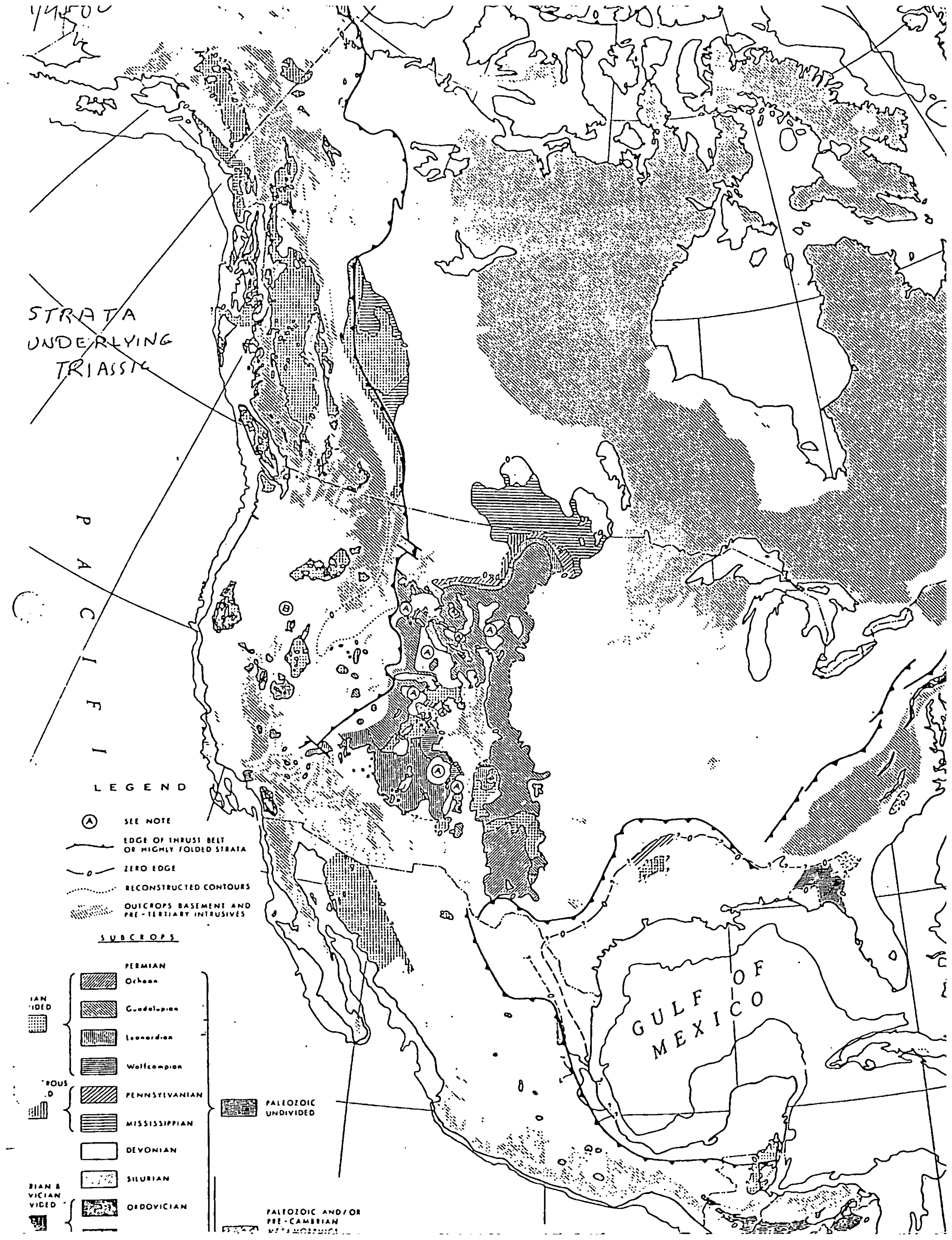
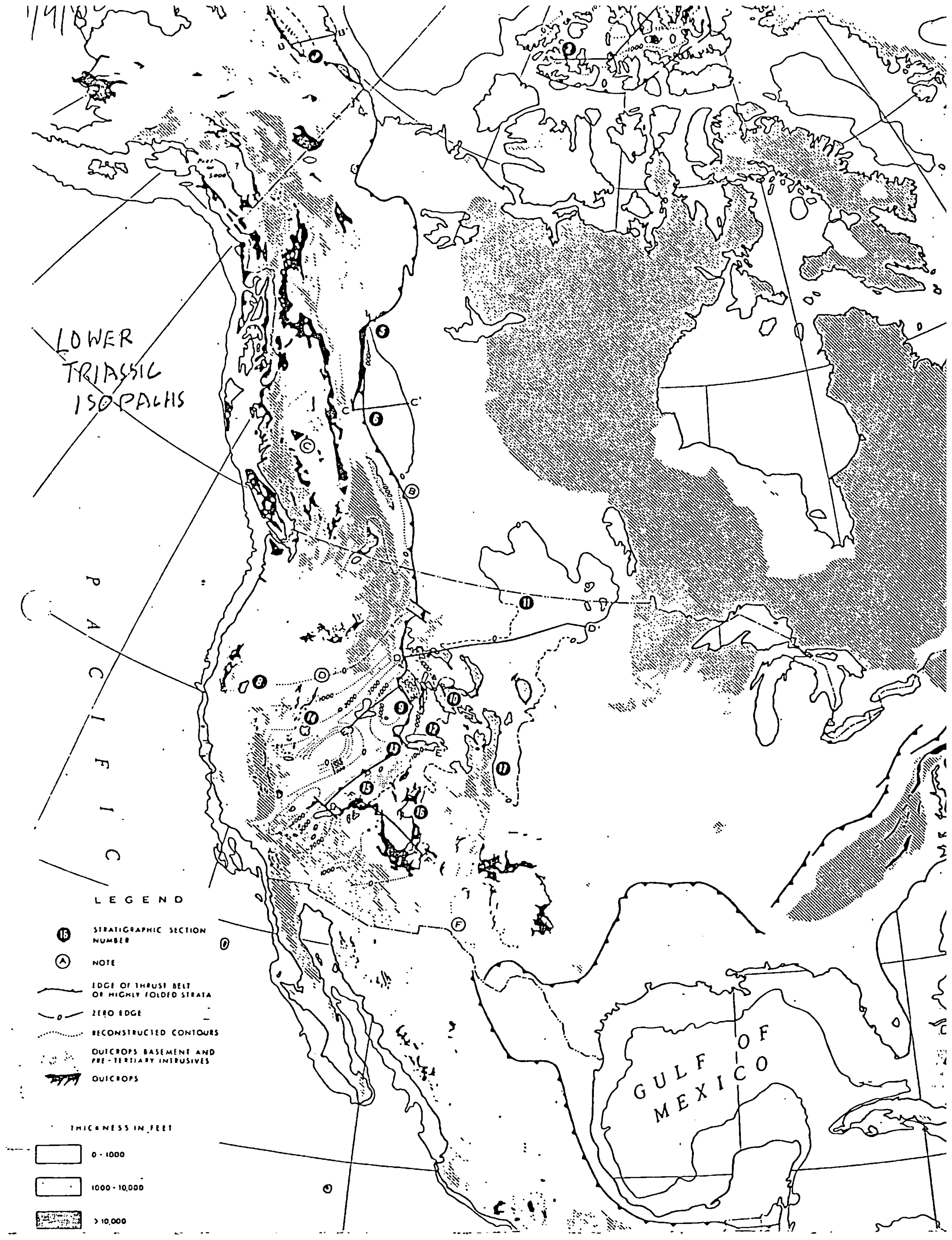
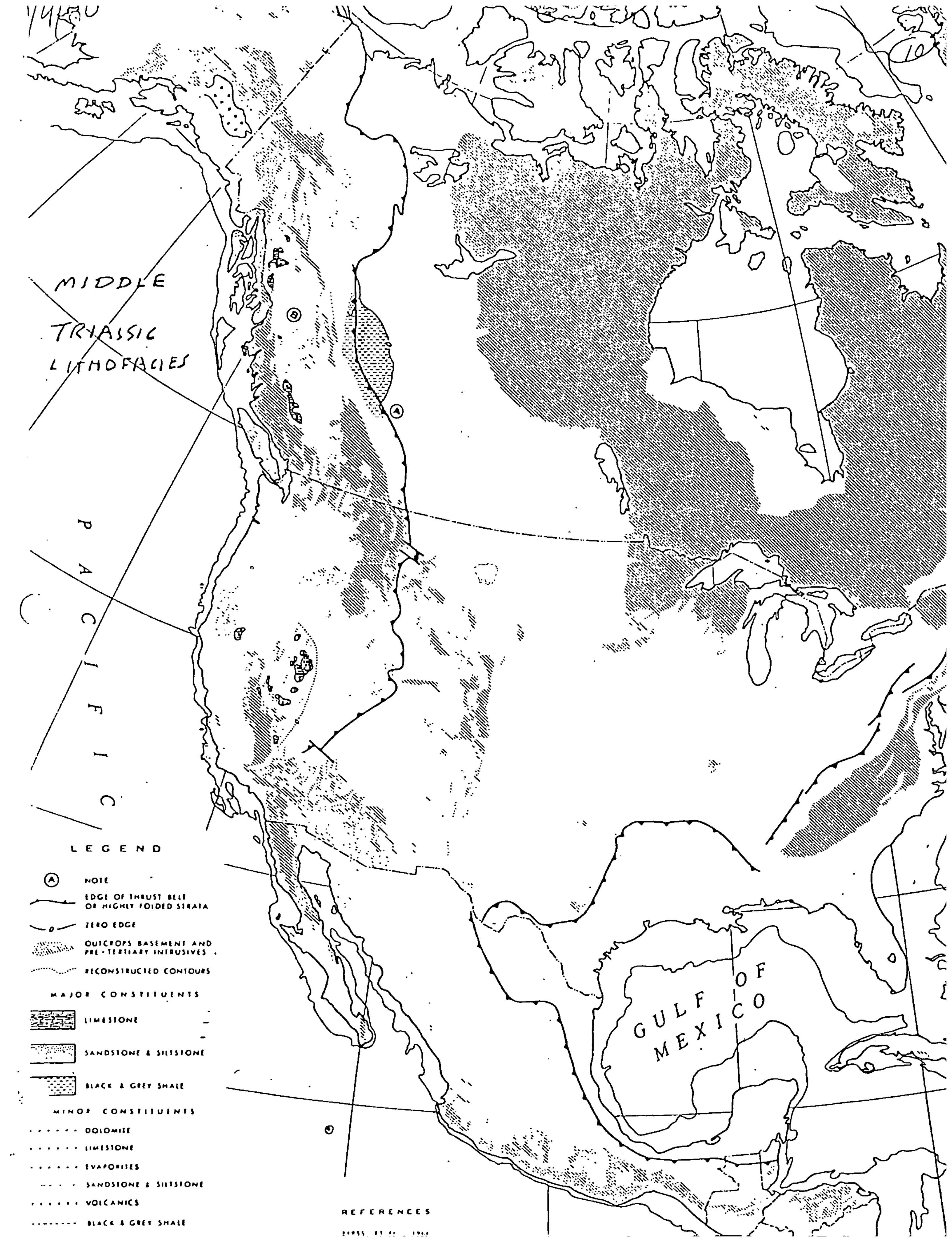
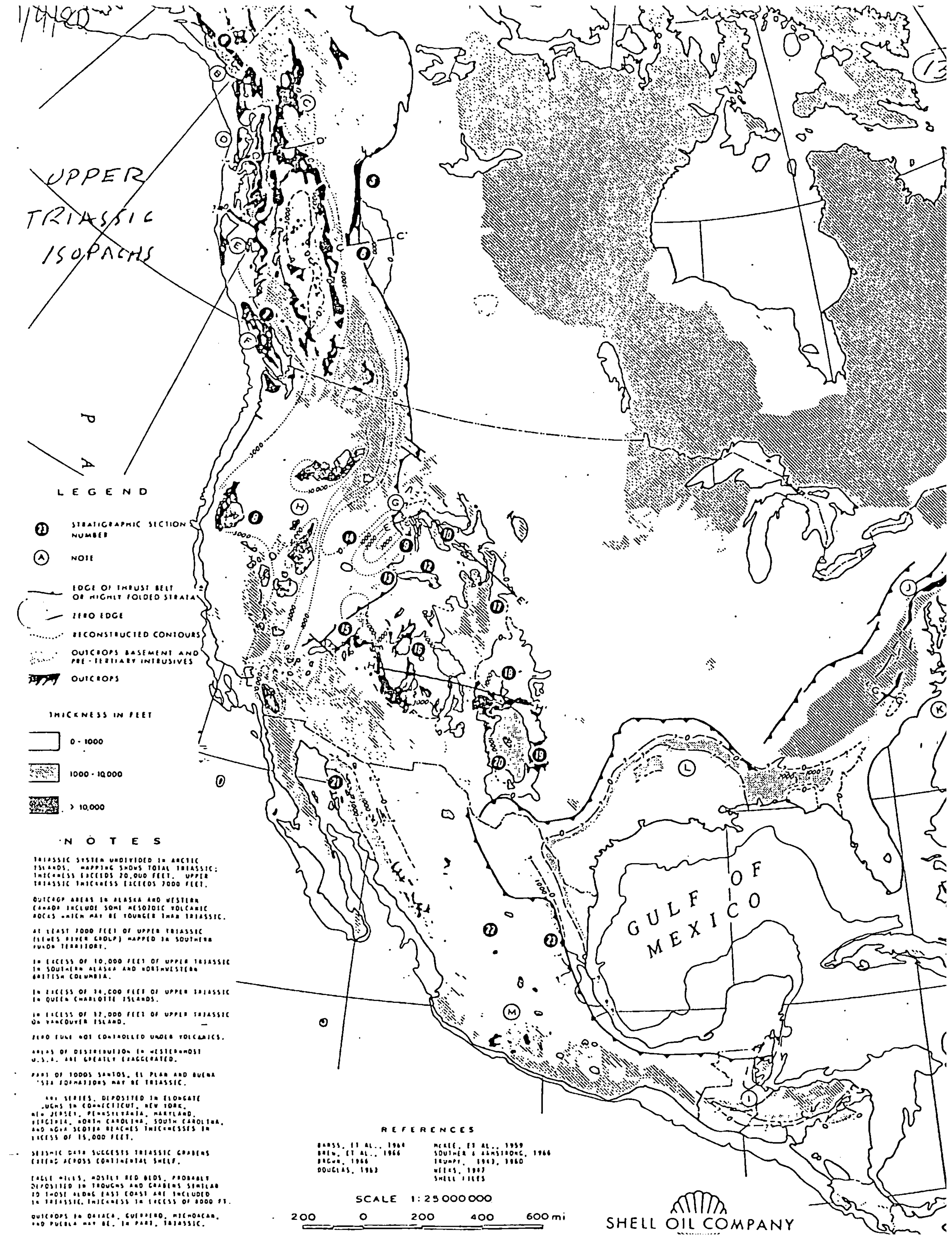

STRATA
UNDERLYING
TRIASSIC
P A C I F I
GULF OF MEXICO
LEGEND
SEE NOTE
EDGE OF THRUST BELT OR HIGHLY FOLDED STRATA
ZERO EDGE
RECONSTRUCTED CONTOURS
OUTCROPS BASEMENT AND PRE-TERTIARY INTRUSIVES
SUBCROPS
PERMIAN
Ochoan
Guadalupian
Leonardian
Wolfcampian
PENNSYLVANIAN
MISSISSIPPIAN
DEVONIAN
SILURIAN
ORDOVICIAN
PALEOZOIC UNDIVIDED
PALEOZOIC AND/OR PRE-CAMBRIAN


LOWER TRIASSIC ISOPACHS
PACIFIC
GULF OF MEXICO
LEGEND
STRATIGRAPHIC SECTION NUMBER
NOTE
EDGE OF THRUST BELT OR HIGHLY FOLDED STRATA
ZERO EDGE
RECONSTRUCTED CONTOURS
OUTCROPS BASEMENT AND PRE-TERTIARY INTRUSIVES
OUTCROPS
THICKNESS IN FEET
0 - 1000
1000 - 10,000
> 10,000


MIDDLE
TRIASSIC
LITHOFACIES
PACIFIC
GULF OF MEXICO
LEGEND
NOTE
EDGE OF THRUST BELT OF HIGHLY FOLDED STRATA
ZERO EDGE
OUTCROPS BASEMENT AND PRE-TERTIARY INTRUSIVES
RECONSTRUCTED CONTOURS
MAJOR CONSTITUENTS
LIMESTONE
SANDSTONE & SILTSTONE
BLACK & GREY SHALE
MINOR CONSTITUENTS
DOLOMITE
LIMESTONE
EVAPORITES
SANDSTONE & SILTSTONE
VOLCANICS
BLACK & GREY SHALE
REFERENCES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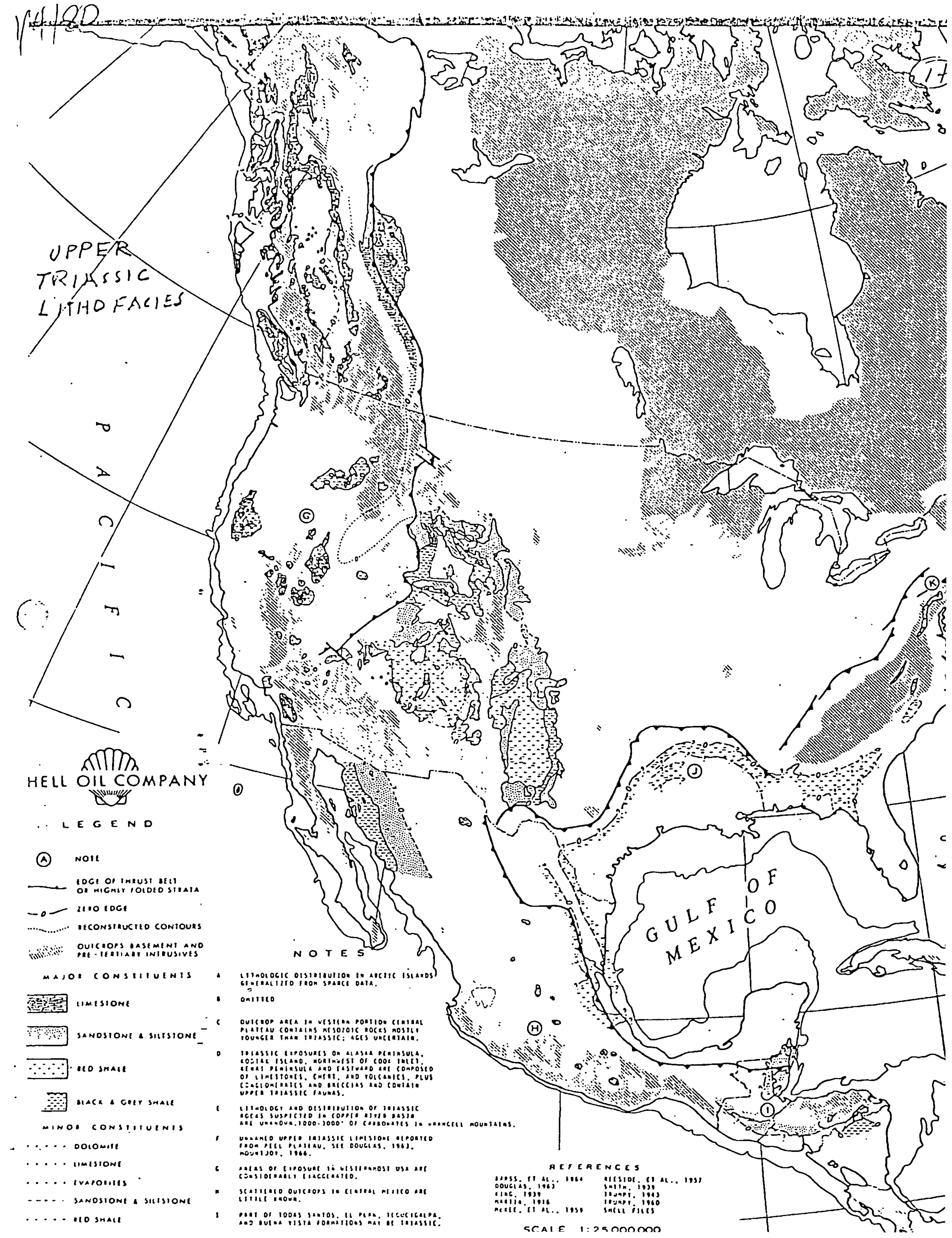
UPPER TRIASSIC ISOPACHS
GULF OF MEXICO
LEGEND
STRATIGRAPHIC SECTION NUMBER
NOTE
EDGE OF THRUST BELT OR HIGHLY FOLDED STRATA
ZERO EDGE
RECONSTRUCTED CONTOURS
OUTCROPS BASEMENT AND PRE-TERTIARY INTRUSIVES
OUTCROPS
THICKNESS IN FEET
0 - 1000
1000 - 10,000
> 10,000
NOTES
TRIASSIC SYSTEM UNDIVIDED IN ARCTIC ISLANDS. MAPPING SHOWS TOTAL TRIASSIC; THICKNESS EXCEEDS 20,000 FEET. UPPER TRIASSIC THICKNESS EXCEEDS 7000 FEET.
OUTCROP AREAS IN ALASKA AND WESTERN CANADA INCLUDE SOME MESOZOIC VOLCANIC ROCKS WHICH MAY BE YOUNGER THAN TRIASSIC.
AT LEAST 7000 FEET OF UPPER TRIASSIC (LEWES RIVER GROUP) MAPPED IN SOUTHERN YUKON TERRITORY.
IN EXCESS OF 10,000 FEET OF UPPER TRIASSIC IN SOUTHERN ALASKA AND NORTHWESTERN BRITISH COLUMBIA.
IN EXCESS OF 14,000 FEET OF UPPER TRIASSIC IN QUEEN CHARLOTTE ISLANDS.
IN EXCESS OF 12,000 FEET OF UPPER TRIASSIC ON VANCOUVER ISLAND.
ZERO EDGE NOT CONTROLLED UNDER VOLCANICS.
AREAS OF DISTRIBUTION IN WESTERNMOST U.S.A. ARE GREATLY EXAGGERATED.
PART OF TODOS SANTOS, EL PLAN AND HUENA VISTA FORMATIONS MAY BE TRIASSIC.
SERIES, DEPOSITED IN ELONGATE TROUGHS IN CONNECTICUT, NEW YORK, NEW JERSEY, PENNSYLVANIA, MARYLAND, VIRGINIA, NORTH CAROLINA, SOUTH CAROLINA, AND NOVA SCOTIA REACHES THICKNESSES IN EXCESS OF 15,000 FEET.
SEISMIC DATA SUGGESTS TRIASSIC GRABENS EXTEND ACROSS CONTINENTAL SHELF.
EAGLE MILLS, MOSTLY RED BEDS, PROBABLY DEPOSITED IN TROUGHS AND GRABENS SIMILAR TO THOSE ALONG EAST COAST ARE INCLUDED IN TRIASSIC. THICKNESS IN EXCESS OF 8000 FT.
OUTCROPS IN OAXACA, GUERRERO, MICHOACAN, AND PUEBLA MAY BE, IN PART, TRIASSIC.
REFERENCES
BARSS, ET AL., 1964
BREW, ET AL., 1966
BROWN, 1966
DOUGLAS, 1962
MCKEE, ET AL., 1959
SOUTHER & ARMSTRONG, 1966
TRUMPY, 1961, 1960
WEEKS, 1947
SHELL FILES
SCALE 1:25 000 000
200 0 200 400 600 mi
SHELL OIL COMPANY



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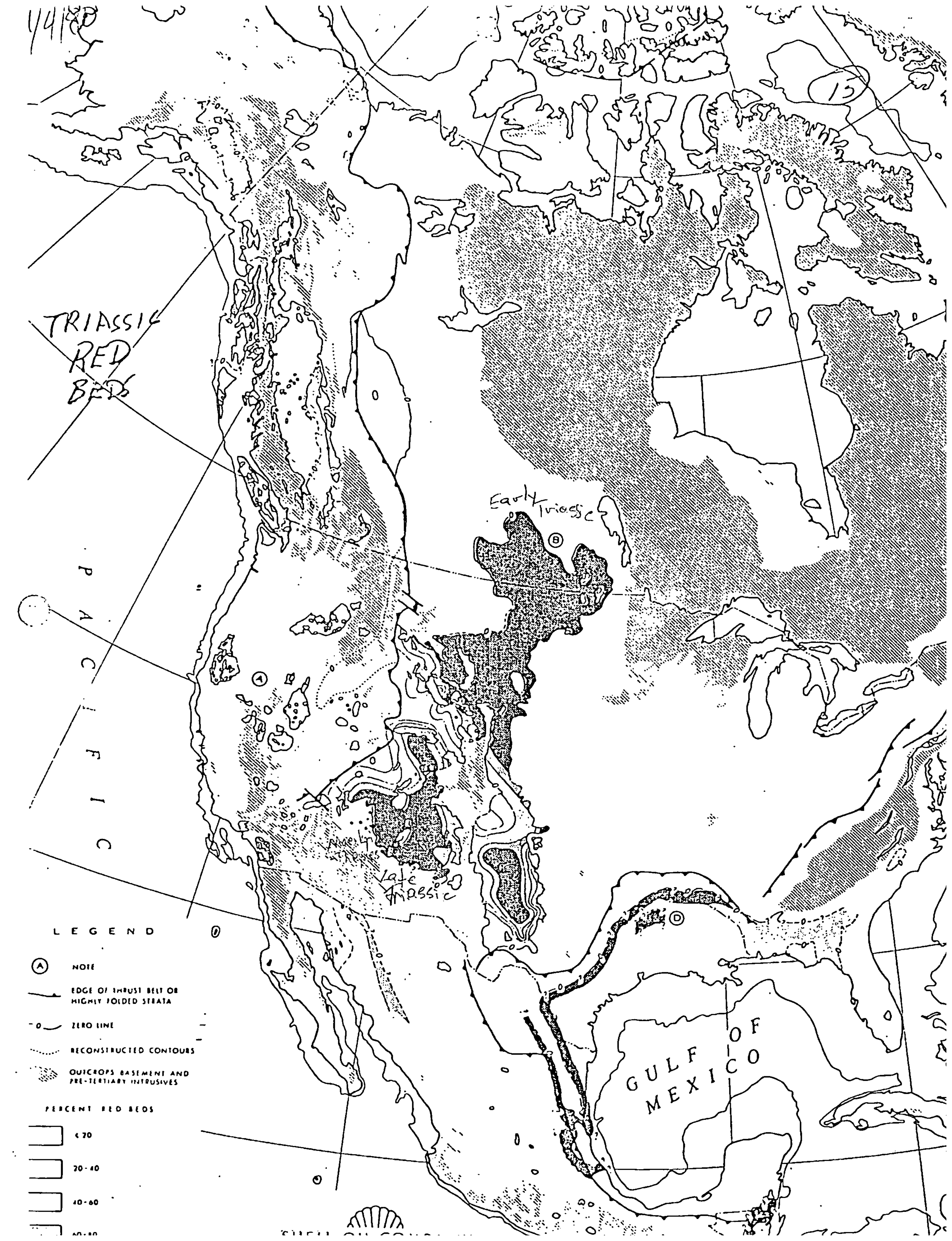
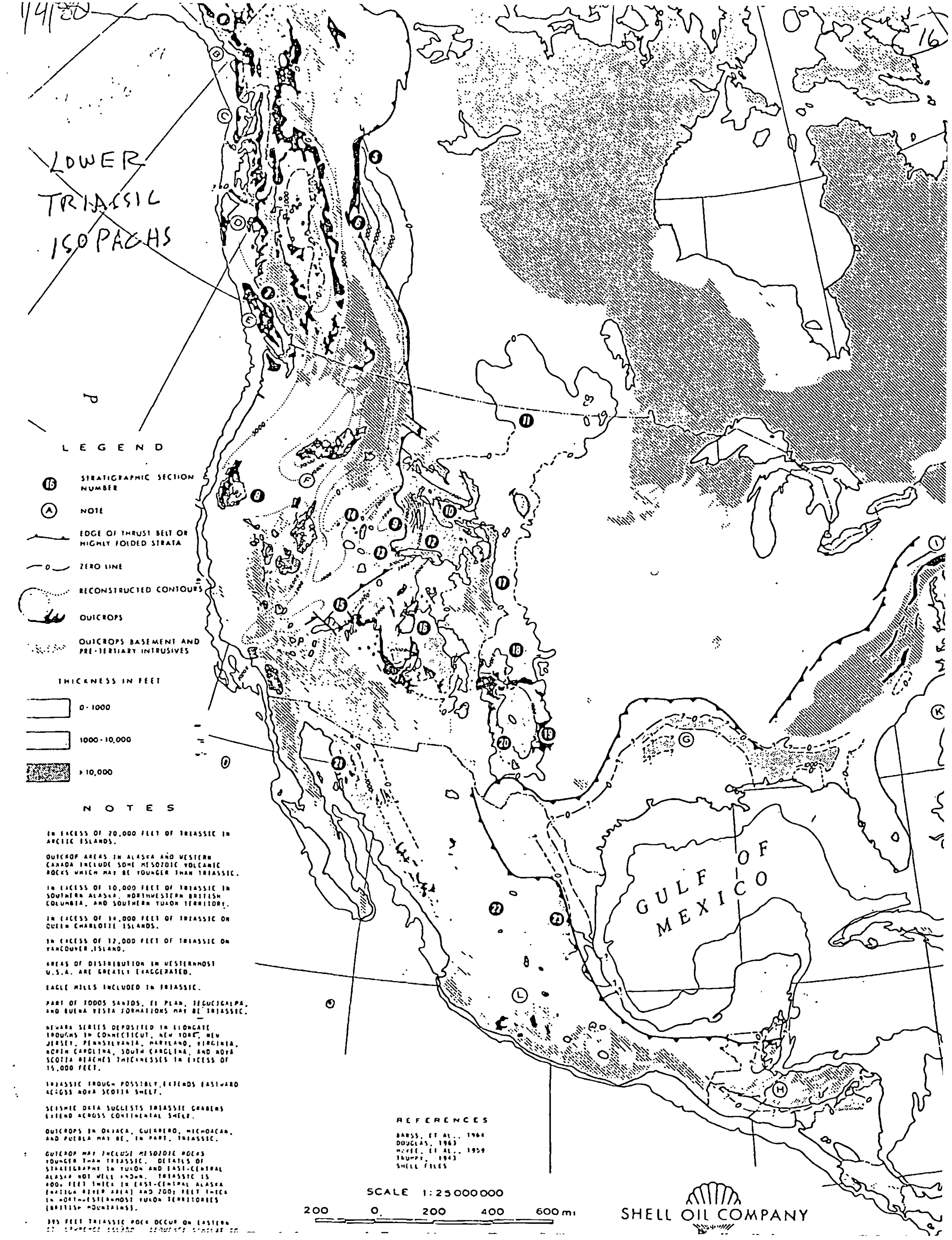
UPPER TRIASSIC LITHOFACIES
P A C I F I C
GULF OF MEXICO
HELL OIL COMPANY
LEGEND
(A) NOTE
EDGE OF THRUST BELT OR HIGHLY FOLDED STRATA
ZERO EDGE
RECONSTRUCTED CONTOURS
OUTCROPS BASEMENT AND PRE-TERTIARY INTRUSIVES
MAJOR CONSTITUENTS
LIMESTONE
SANDSTONE & SILTSTONE
RED SHALE
BLACK & GREY SHALE
MINOR CONSTITUENTS
DOLOMITE
LIMESTONE
EVAPORITES
SANDSTONE & SILTSTONE
RED SHALE
NOTES
A LITHOLOGIC DISTRIBUTION IN ARCTIC ISLANDS GENERALIZED FROM SPARCE DATA.
B OMITTED
C OUTCROP AREA IN WESTERN PORTION CENTRAL PLATEAU CONTAINS MESOZOIC ROCKS MOSTLY YOUNGER THAN TRIASSIC; AGES UNCERTAIN.
D TRIASSIC EXPOSURES ON ALASKA PENINSULA, KODIAK ISLAND, NORTHWEST OF COOK INLET, KENAI PENINSULA AND EASTWARD ARE COMPOSED OF LIMESTONES, CHERT, AND VOLCANICS, PLUS CONGLOMERATES AND BRECCIAS AND CONTAIN UPPER TRIASSIC FAUNAS.
E LITHOLOGY AND DISTRIBUTION OF TRIASSIC ROCKS SUSPECTED IN COPPER RIVER BASIN ARE UNKNOWN. 1000-3000' OF CARBONATES IN WRANGELL MOUNTAINS.
F UNNAMED UPPER TRIASSIC LIMESTONE REPORTED FROM PEEL PLATEAU, SEE DOUGLAS, 1963, MOUNTJOY, 1966.
G AREAS OF EXPOSURE IN WESTERNMOST USA ARE CONSIDERABLY EXAGGERATED.
H SCATTERED OUTCROPS IN CENTRAL MEXICO ARE LITTLE KNOWN.
I PART OF TODAS SANTOS, EL PLAN, TEGUCIGALPA, AND BUENA VISTA FORMATIONS MAY BE TRIASSIC.
REFERENCES
BARSS, ET AL., 1964
DOUGLAS, 1963
KING, 1939
MARTIN, 1916
MCKEE, ET AL., 1959
REESIDE, ET AL., 1957
SMITH, 1939
TRUMPY, 1943
TRUMPY, 1960
SHELL FILES
SCALE 1:25,000,000


TRIASSIC RED BEDS
15
Early Triassic
Late Triassic
P A C I F I C
GULF OF MEXICO
LEGEND
NOTE
EDGE OF THRUST BELT OR HIGHLY FOLDED STRATA
ZERO LINE
RECONSTRUCTED CONTOURS
OUTCROPS BASEMENT AND PRE-TERTIARY INTRUSIVES
PERCENT RED BEDS
< 20
20 - 40
40 - 60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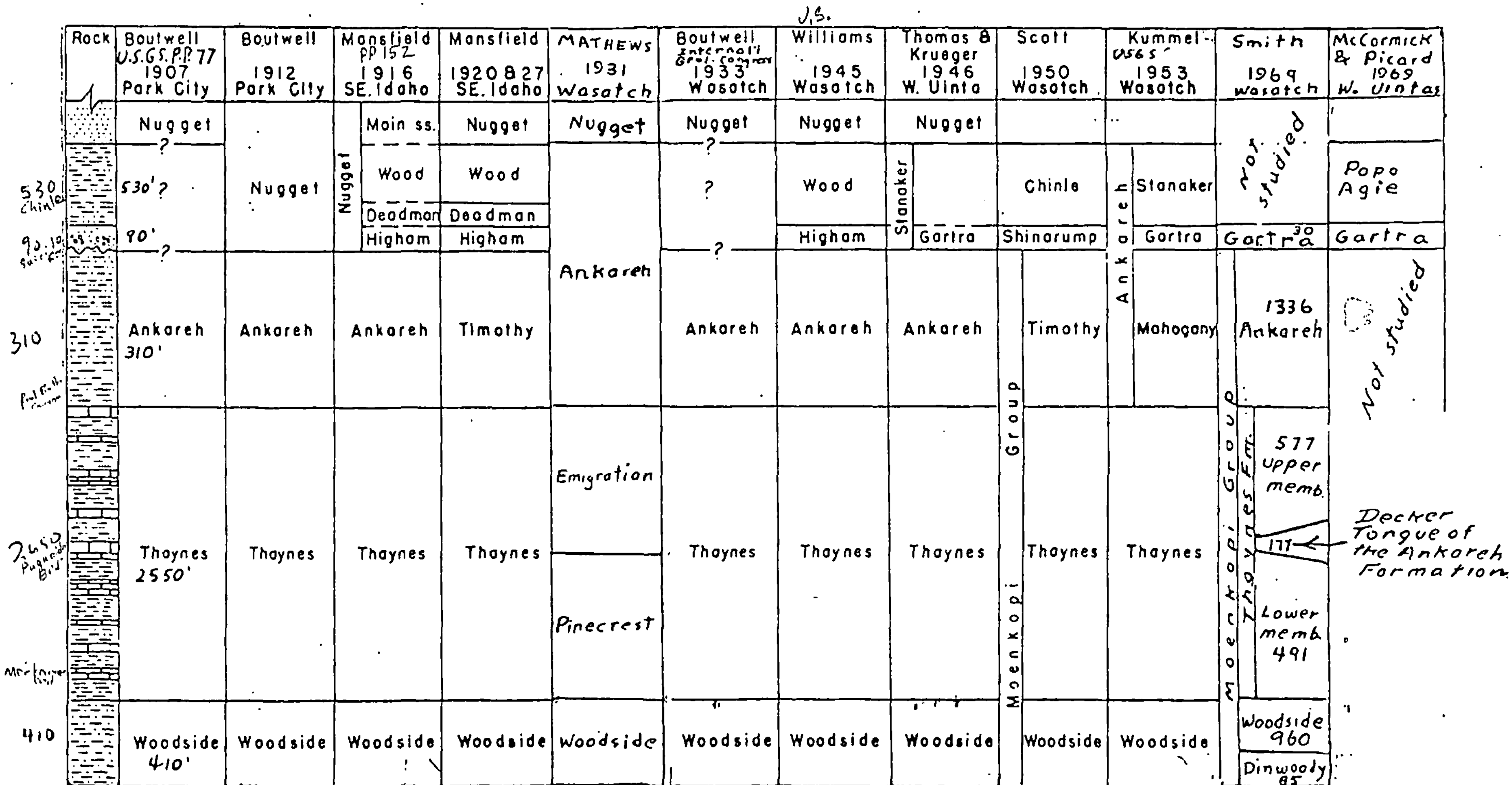
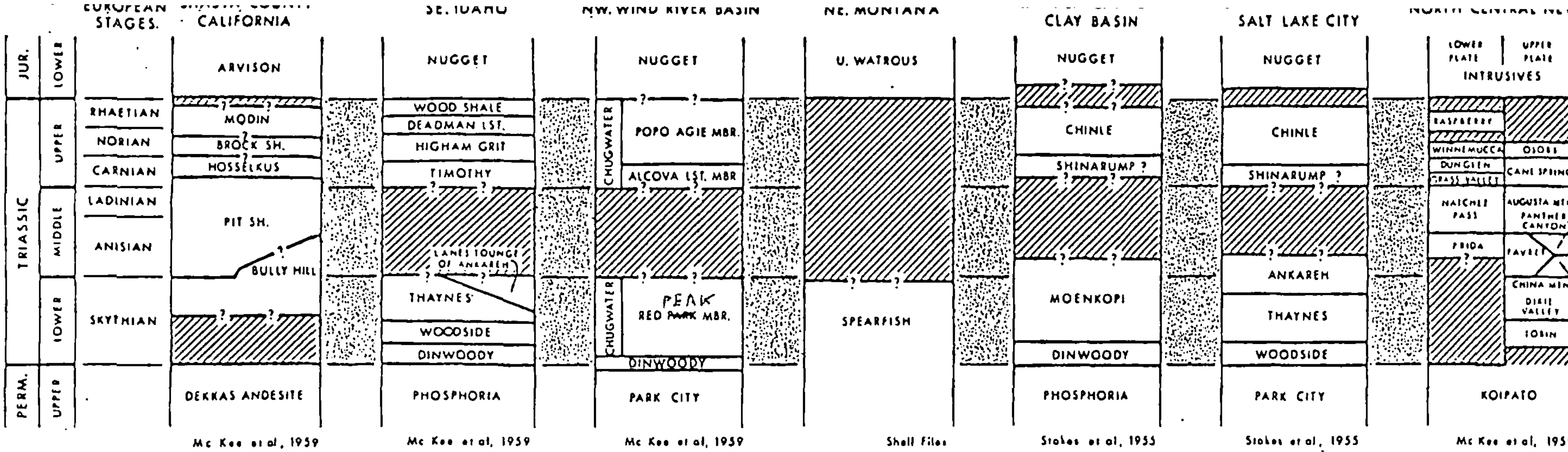
16
LOWER TRIASSIC ISOPACHS
LEGEND
STRATIGRAPHIC SECTION NUMBER
NOTE
EDGE OF THRUST BELT OR HIGHLY FOLDED STRATA
ZERO LINE
RECONSTRUCTED CONTOURS
OUTCROPS
OUTCROPS BASEMENT AND PRE-TERTIARY INTRUSIVES
THICKNESS IN FEET
0 - 1000
1000 - 10,000
> 10,000
NOTES
IN EXCESS OF 20,000 FEET OF TRIASSIC IN ARCTIC ISLANDS.
OUTCROP AREAS IN ALASKA AND WESTERN CANADA INCLUDE SOME MESOZOIC VOLCANIC ROCKS WHICH MAY BE YOUNGER THAN TRIASSIC.
IN EXCESS OF 10,000 FEET OF TRIASSIC IN SOUTHERN ALASKA, NORTHWESTERN BRITISH COLUMBIA, AND SOUTHERN YUKON TERRITORY.
IN EXCESS OF 14,000 FEET OF TRIASSIC ON QUEEN CHARLOTTE ISLANDS.
IN EXCESS OF 12,000 FEET OF TRIASSIC ON VANCOUVER ISLAND.
AREAS OF DISTRIBUTION IN WESTERNMOST U.S.A. ARE GREATLY EXAGGERATED.
EAGLE MILLS INCLUDED IN TRIASSIC.
PART OF TODOS SANTOS, EL PLAN, TEGUCIGALPA, AND BUENA VISTA FORMATIONS MAY BE TRIASSIC.
NEWARK SERIES DEPOSITED IN ELONGATE TROUGHS IN CONNECTICUT, NEW YORK, NEW JERSEY, PENNSYLVANIA, MARYLAND, VIRGINIA, NORTH CAROLINA, SOUTH CAROLINA, AND NOVA SCOTIA REACHES THICKNESSES IN EXCESS OF 15,000 FEET.
TRIASSIC TROUGH POSSIBLY EXTENDS EASTWARD ACROSS NOVA SCOTIA SHELF.
SEISMIC DATA SUGGESTS TRIASSIC GRABENS EXTEND ACROSS CONTINENTAL SHELF.
OUTCROPS IN OAXACA, GUERRERO, MICHOACAN, AND PUEBLA MAY BE, IN PART, TRIASSIC.
OUTCROP MAY INCLUDE MESOZOIC ROCKS YOUNGER THAN TRIASSIC. DETAILS OF STRATIGRAPHY IN YUKON AND EAST-CENTRAL ALASKA NOT WELL KNOWN. TRIASSIC IS 400± FEET THICK IN EAST-CENTRAL ALASKA (NATION RIVER AREA) AND 200± FEET THICK IN NORTHWESTERNMOST YUKON TERRITORIES (BRITISH MOUNTAINS).
395 FEET TRIASSIC ROCK OCCUR ON EASTERN
REFERENCES
BARSS, ET AL., 1964
DOUGLAS, 1963
McKEE, ET AL., 1959
TRUMPY, 1943
SHELL FILES
GULF OF MEXICO
SCALE 1:25 000 000
200 0 200 400 600 mi
SHELL OIL COMPANY

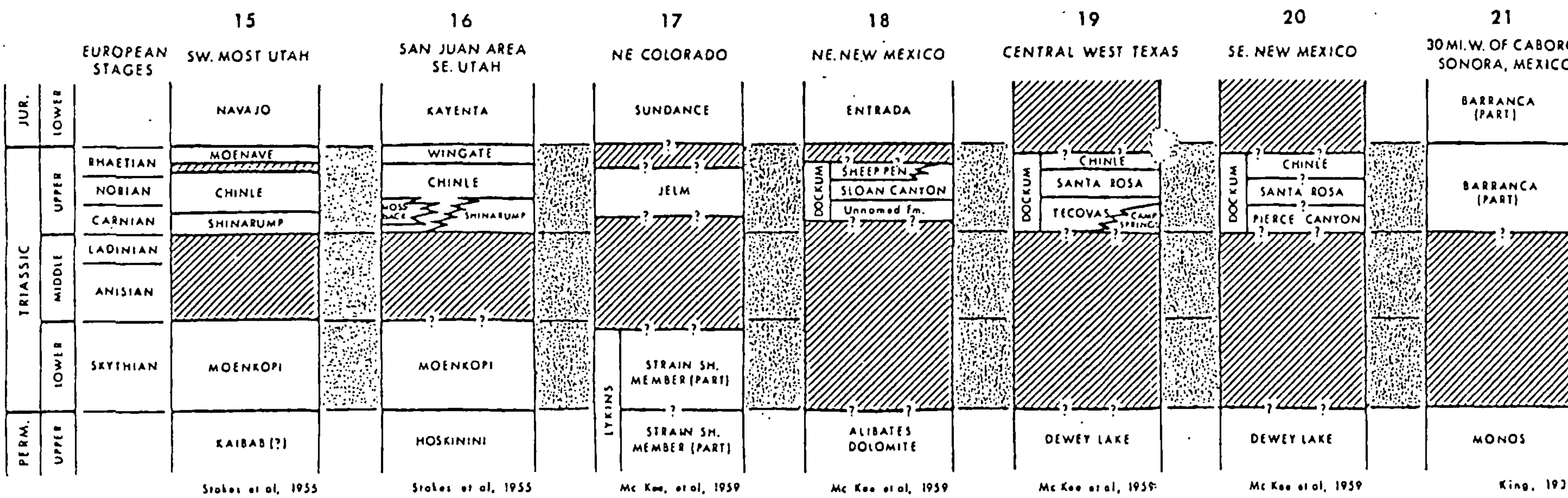
Evolution of Nomenclature of Triassic Rocks
near Salt Lake City

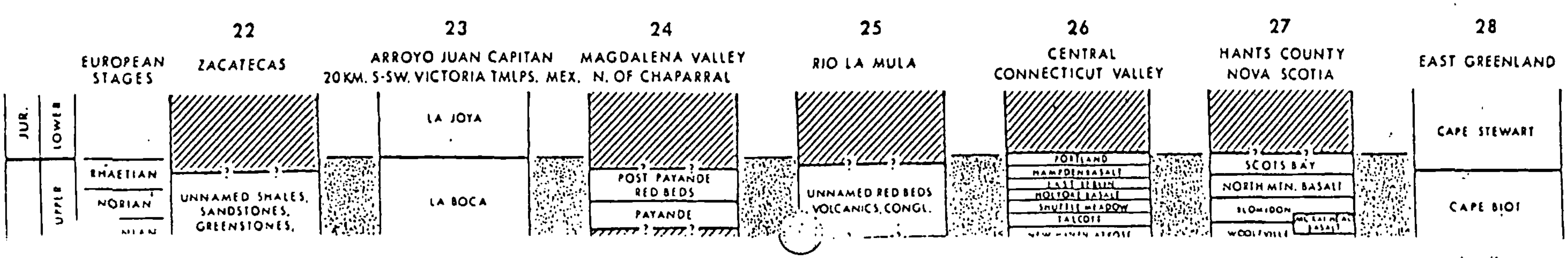


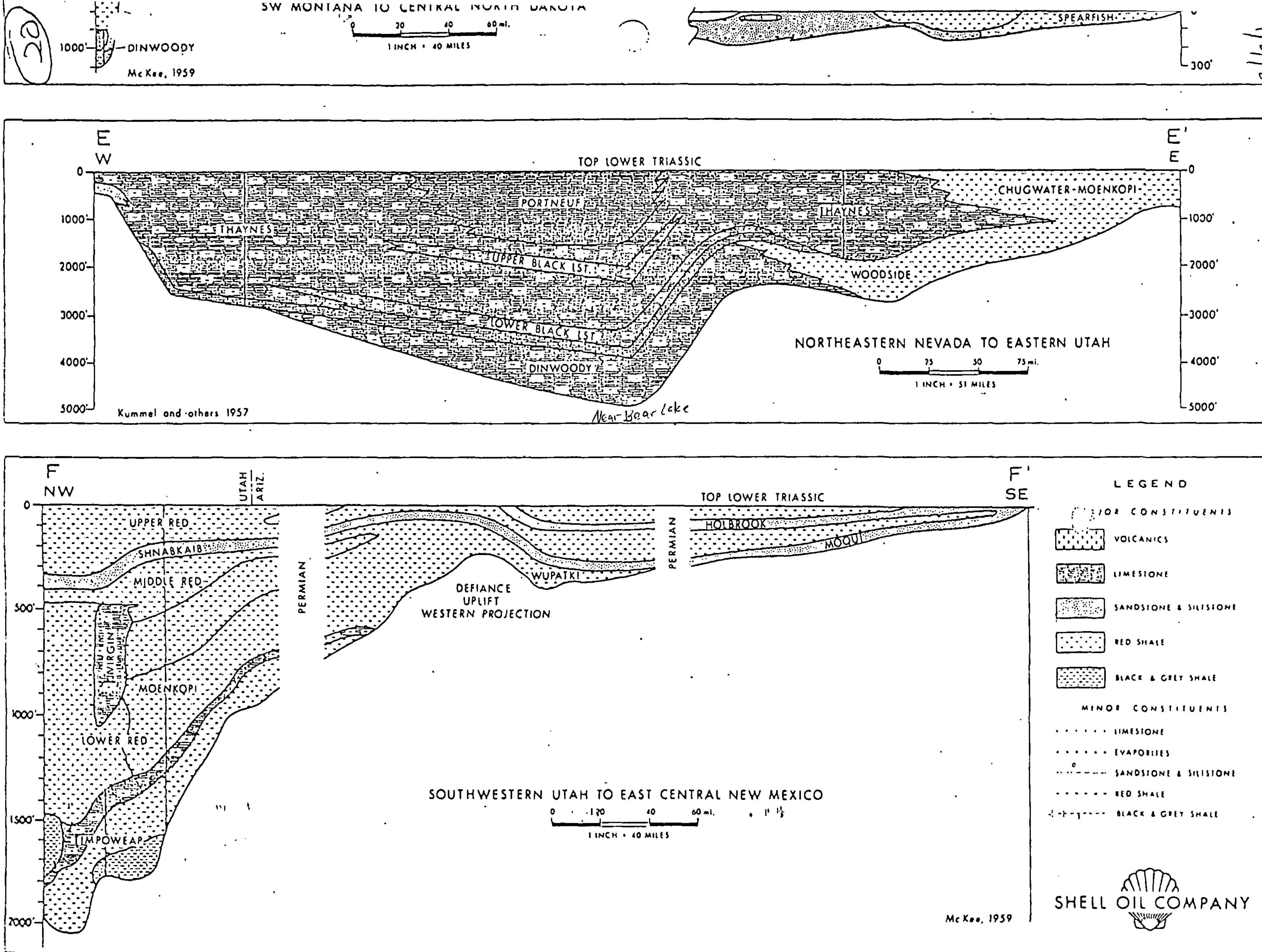
Adapted chiefly from "Stratigraphy of
the Triassic Sequence in the Wasatch
and Uinta Mountains". by W. F. Scott,
Guidebook to the 10th Ann. Field Conference



TRIASSIC STRATIGRAPHIC SECTIONS - WESTERN U.S.

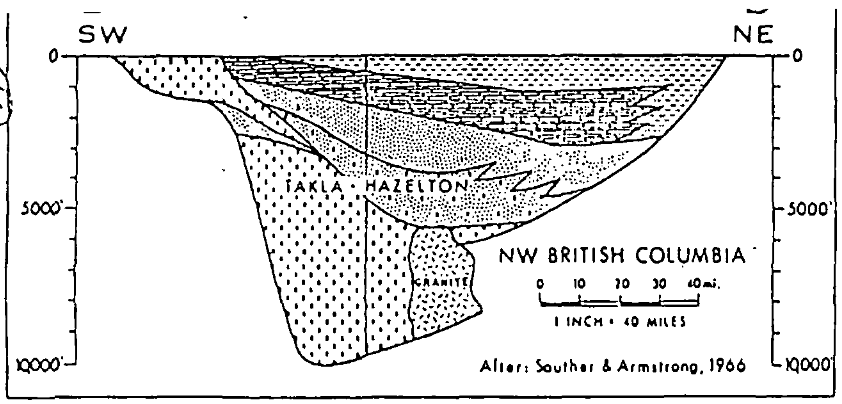


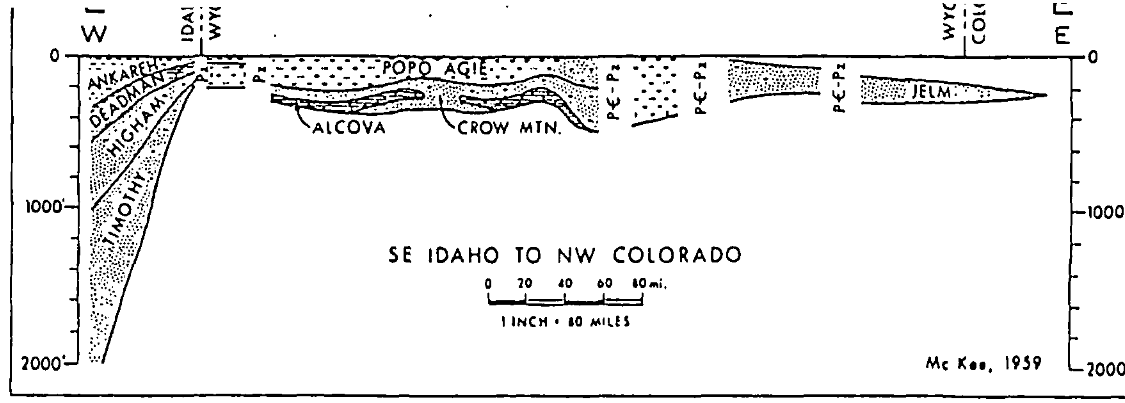


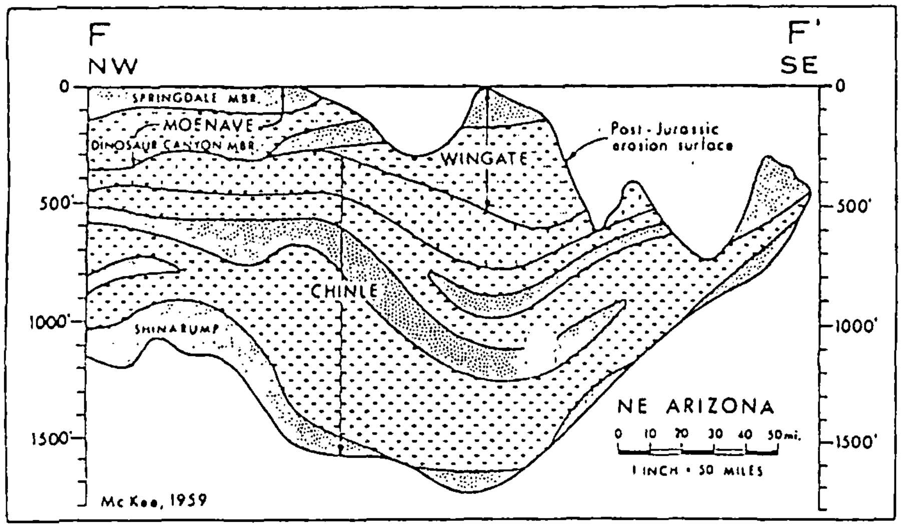


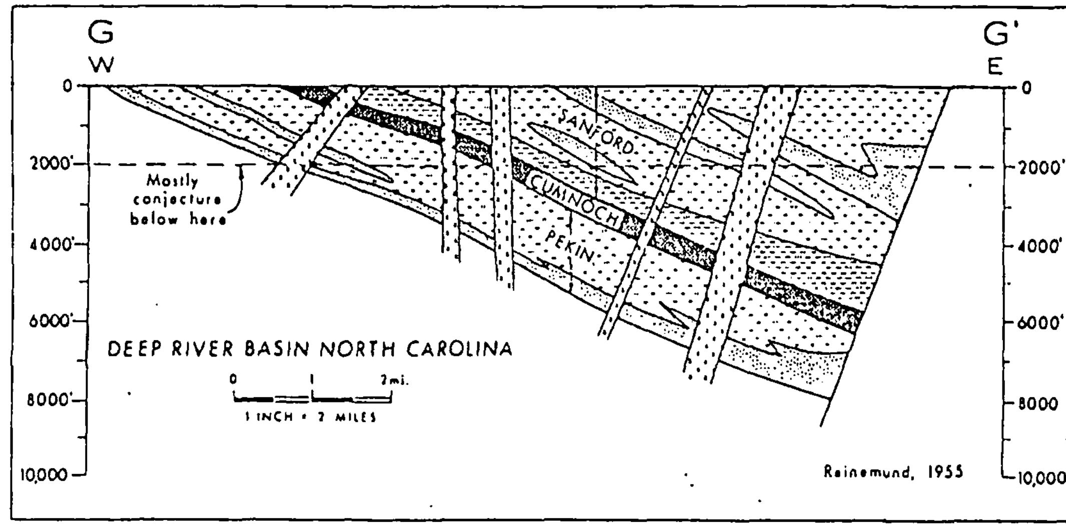
LOWER TRIASSIC CROSS SECTIONS

(14)

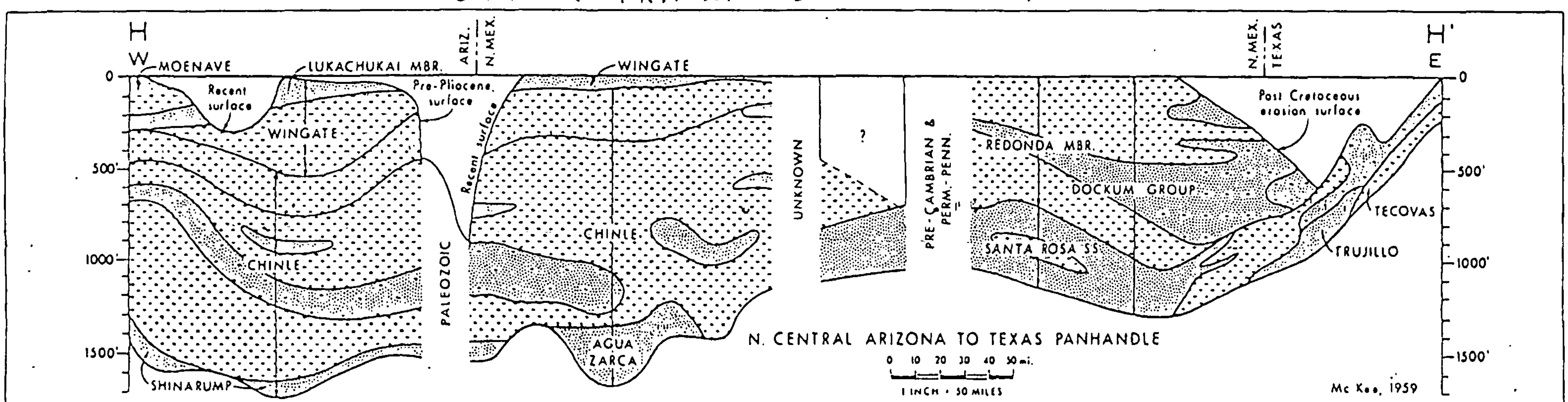








UPPER TRIASSIC CROSS-SECTIONS



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TRIASSIC PROBLEMS

Do no. 1; one from no's 2, 3 and 4 and one from 5, 6, 7, 8, 9, and 10.

1. Ancestral Rockies: Sketch on a page-size map of the Western Interior an outline of the various elements of the Ancestral Rockies as they existed at the beginning of the Triassic Period, do the same for the end of the Period. Show on this same map the edge of Permo-Triassic red beds. Discuss evidence for and against the theory that western red-beds are genetically related to uplifts of the Ancestral Rockies

2. Triassic Depositional Centers: Locate and briefly describe the two thickest Triassic sections of the Western Interior. Think of a miogeosynclinal section and a eugeosynclinal section. Relate these sedimentary accumulations to tectonic influences - in other words, under what conditions was space created to receive unusually large volumes of sediment.

3. Sediment dispersal problem: Note that practically all Triassic formations east of the Wasatch Line represent sediment in transit westward. Where did it ultimately come to rest? Read Lupe 1979 GSA abstract, p. 470, San Diego Meeting. Assume he is right, make a paleogeographical map showing the essential facts, include source area, transportation zone and depositional site. Think about how you are going to get sediment across the Great Basin.

4. Sheet- and Channel-Sandstones: a) Construct a tabular comparison of sedimentary features of the Shinarump, Moss Back, and Gartra. Include items such as clast lithology (be specific), fine components (matrix and cement), color, primary structures and any other features that you think important. b) Think about and comment on various ideas as to how thin sheets or shallow channel accumulations of continental conglomerate might have formed.

5. Formations v Facies: Study U.S.G.S. Professional Paper 691 and Utah Geological and Mineral Survey Bull. 104. The former describes the Moenkopi in terms of formal stratigraphy the latter in terms of facies. Contrast and compare t... two approaches and comment on their reliability, usefulness, and information content. Be subjective, which method is most helpful to you? Is a combination of methods (both verbal and graphic) possible?

6. Central Wasatch Section: Considering the historical development of formal stratigraphic nomenclature for the central Wasatch near Salt Lake City express an opinion as to which combination of names you prefer. Explain and defend your opinions. Keep in mind priority in naming.

7. Triassic of southern Arizona: Describe how Triassic rocks came to be discovered in southern Arizona. Describe these rocks and tell how they are significant paleogeographically.

8. What happened to the Middle Triassic: Middle Triassic is a recognized rock-stratigraphic unit that accompanies the Lower and Upper divisions almost everywhere but in the Western Interior. Describe and evaluate evidence that rocks of this age do exist in the Interior. Describe from both positive and negative evidence what you think was going on during this 10-million year interval.

9. The Chinle Complex: Construct a chart showing the evolution of nomenclature for the Chinle Formation of southeastern Utah. Comment on the utility of the latest schemes - are they working? (See Stewart; '57, p. 446 and Doelling, Bull 107.) Do you think that facies designations would serve better than formations. Mention by name some facies that might be designated, what environments might thus be designated?

10. Permian-Triassic Boundary: Discuss the Permian-Triassic boundary problem as it is perceived in the Western Interior. Where is the boundary located in terms of formal lithostratigraphy; is it apparent through facies changes, and are there systematic geographic patterns in the manner in which it is expressed? Think of both marine and non-marine stratigraphy. Hint: investigate the Hoskinnini Formation.

Generalizations

JURASSIC STRATIGRAPHY OF WESTERN INTERIOR

I Historical:

a) One of the first-named periods - von Humboldt, 1795

b) First system to be mongraphed on country wide scale - Arkell, 1933.

c) First system to be monographed on world wide scale - Arkell, 1956.

d) First to be covered by U. S. Geological Survey Paleotectonic Map Series (Miscellaneous Geologic Investigations Map I-175), 1956.

II Tectonics and Paleogeography:

a) Radical changes in general tectonic patterns. Previous positives are now negative, and previous negative areas are now positive. Mesocordilleran High seems to be gradually rising at the site of the Rocky Mountain Paleozoic Geosyncline, increased sedimentation on Utah-Wyoming Shelf.

b) Transcontinental Arch still weakly evident with last remnants of Ancestral Rockies finally buried in Late Jurassic.

c) Accentuated downfolding along Wasatch Line

d) Distant effects of disturbances on Pacific Border, Klamathonia

III Sedimentation

a) Dominance of aeolian sand deposits early in period.

b) Red beds in relation to Ancestral Rockies.

c) Complex interfingering of continental and marine beds.

d) Gypsum, salt and glauconite.

e) Alluvial plains and pedimentary deposits - late in period.

f) Volcanic contributions - ash beds, bentonites, tuffs

IV Boundaries:

a) See Triassic notes for lower boundary.

b) Upper Jurassic-Cretaceous boundary is in continental beds.

V Economic Geology

a) Leading source of uranium in North America, at least until recently.

b) Associated with the uranium are vanadium and copper

c) Oil and gas in both marine and non-marine beds; Montana, Wyoming, Utah (various Book Cliffs fields)

d) Salt and gypsum

e) Cement rock, Utah-Idaho, Twin Creek Limestone

f) Building stone locally, Navajo-Nugget sandstone

Paleontology:

a) Invertebrates

1. Ammonites -- world wide guides, rather poor in United States, better in Mexico and Alaska.

2. Other molluscs -- gastropods and pelecypods are varied and important in Western Interior.

3. Microfossils are gaining in importance but not plentiful, non-marine types are more important than are marine.

b) Plants

1. Cycads

2. Ferns

3. Conifers -- uranium "tree formations".

4. Where are the angiosperms?

5. Charophytes -- marine and non-marine algae.

c) Vertebrates

1. All classes now present, fish, amphibians, reptiles, birds and mammals

2. Dinosaurs

Scenic Areas:

a) Zion Park and Monument

b) Dixie Park

c) Capital Reef National Park

d) Glen Canyon - Lake Powell Recreation area

e) Rainbow Natural Bridge

f) Canyonlands National Park

g) Arches National Park

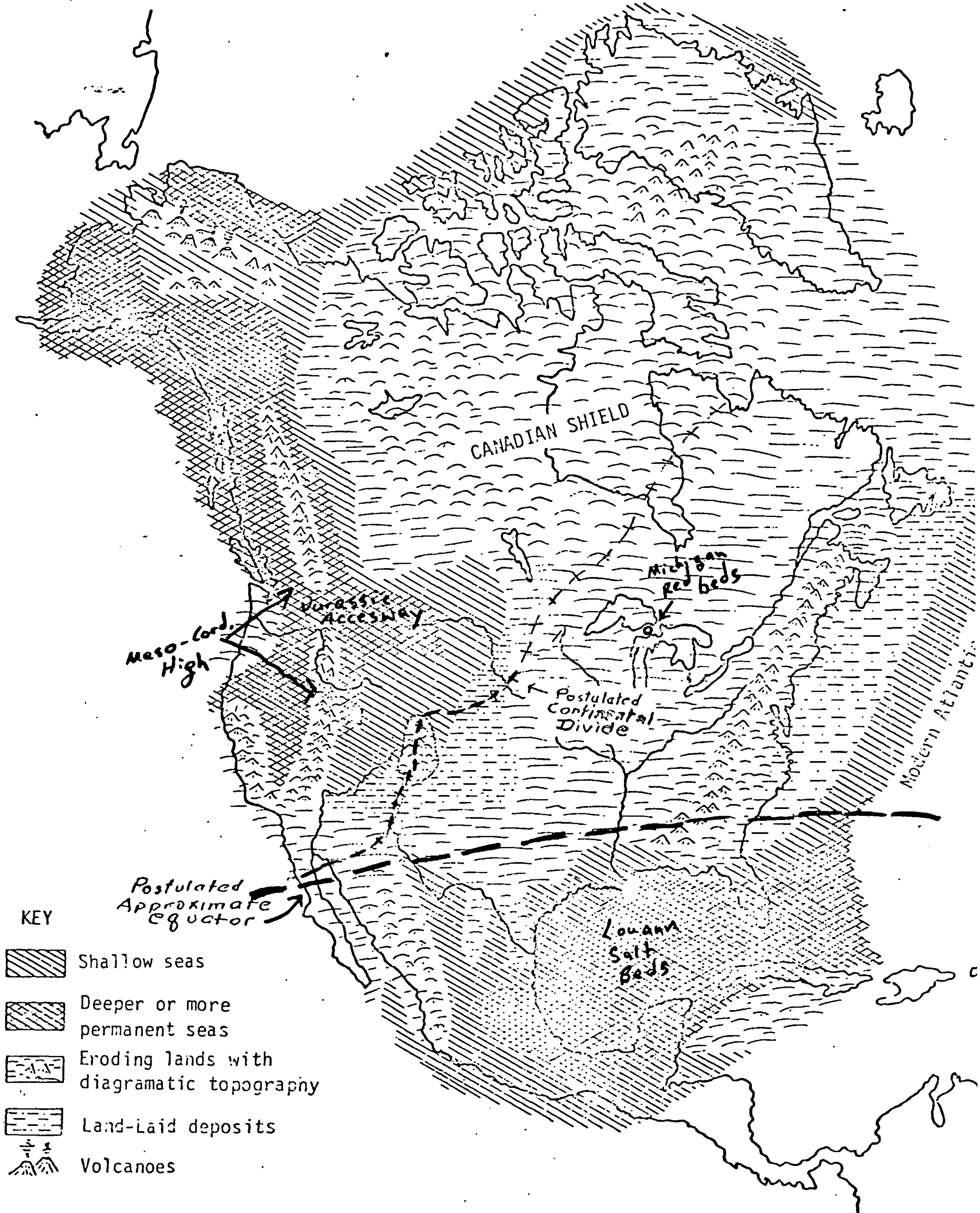
h) Dinosaur National Monument

i) Cleveland-Lloyd Quarry

j) Goblin and Cathedrial Valleys

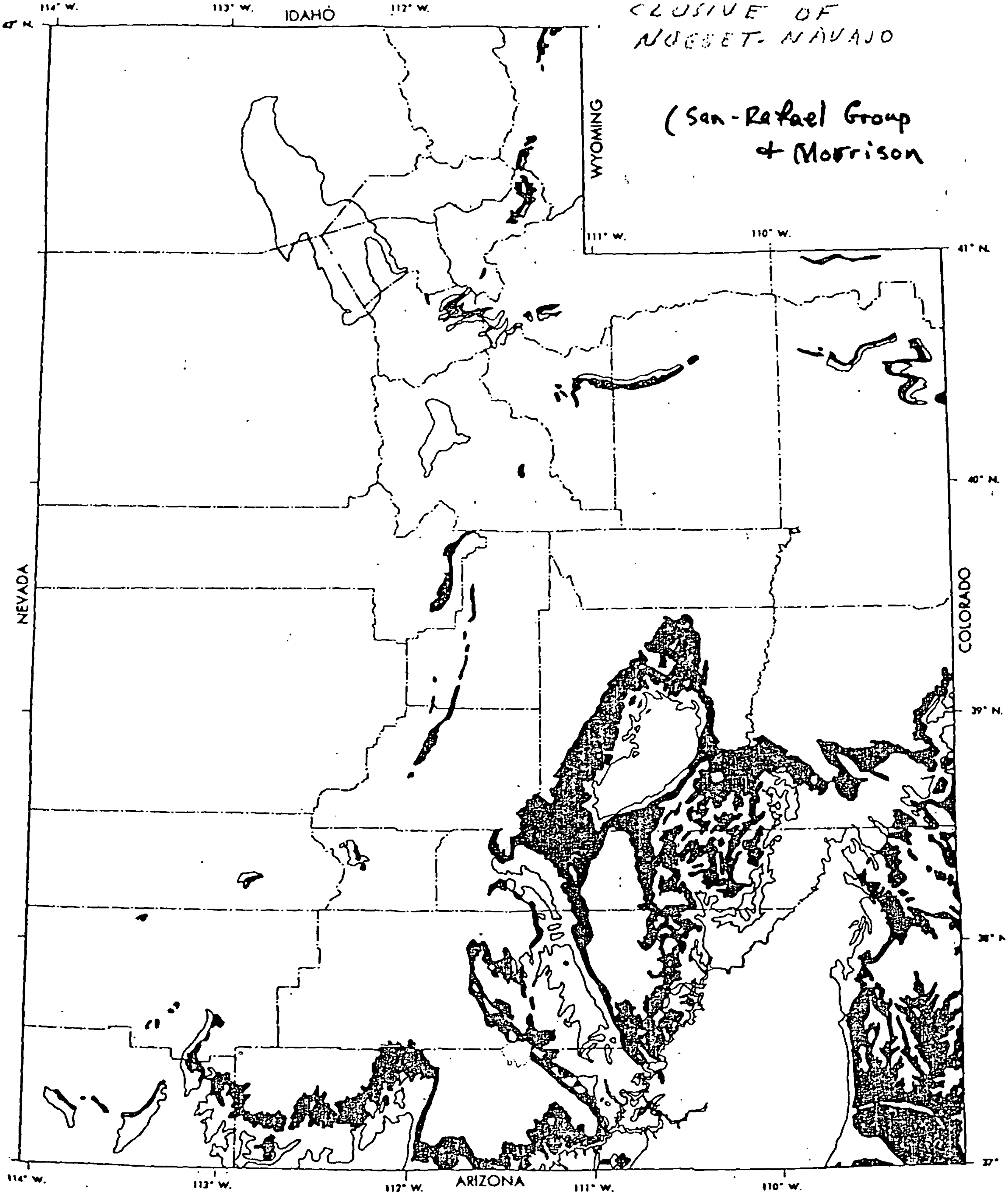
k) Hole in the Rock, Crossing of the Fathers.

W. L. Stokes


CANADIAN SHIELD
Michigan Red beds
Jurassic Accessway
Meso-Cord. High
Postulated Continental Divide
Postulated Approximate Equator
Louann Salt Beds
Modern Atlantic
KEY
Shallow seas
Deeper or more permanent seas
Eroding lands with diagramatic topography
Land-Laid deposits
Volcanoes

JURASSIC - EX-
CLUSIVE OF
NUGGET- NAVAJO

(San-Rafael Group
+ Morrison



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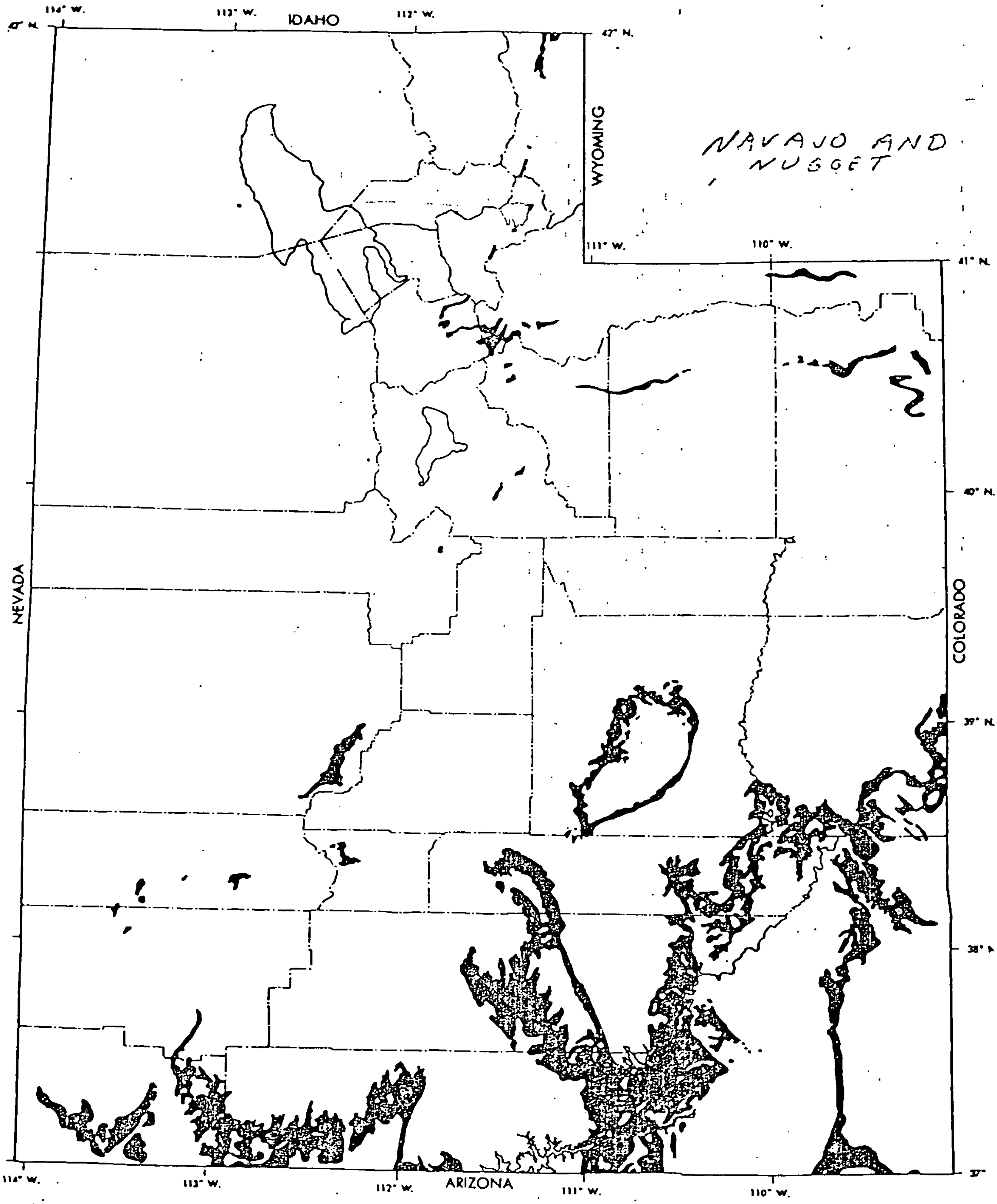
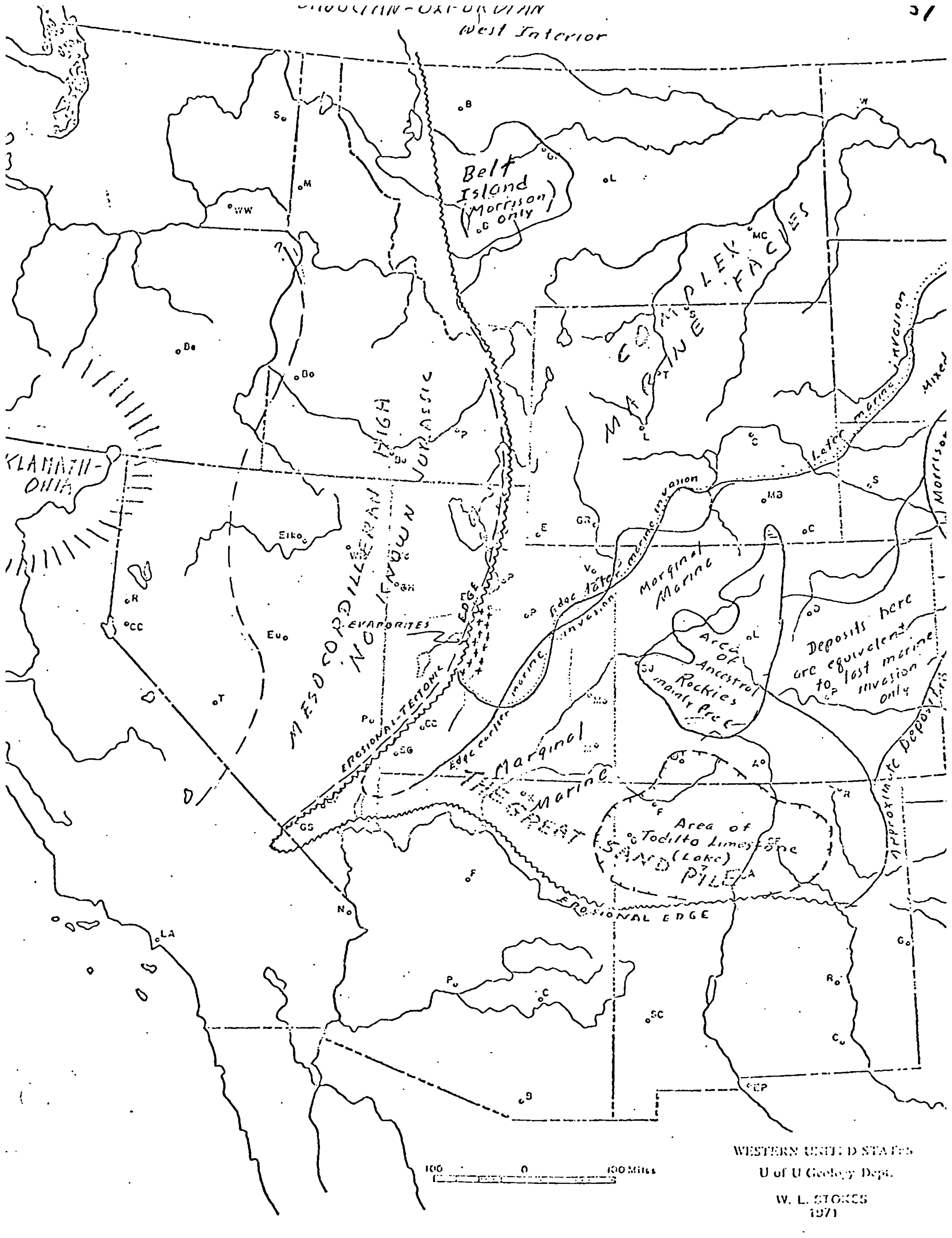
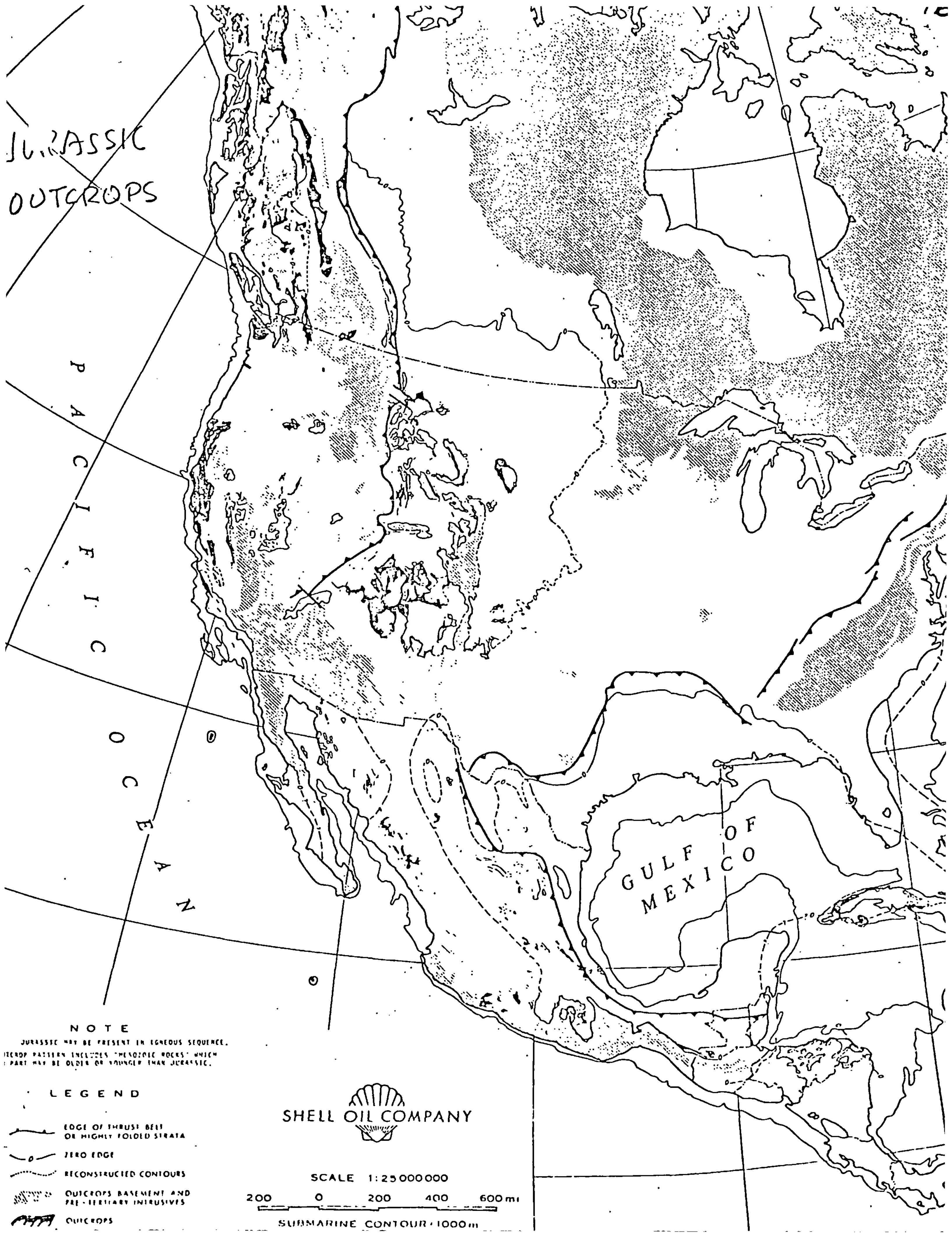


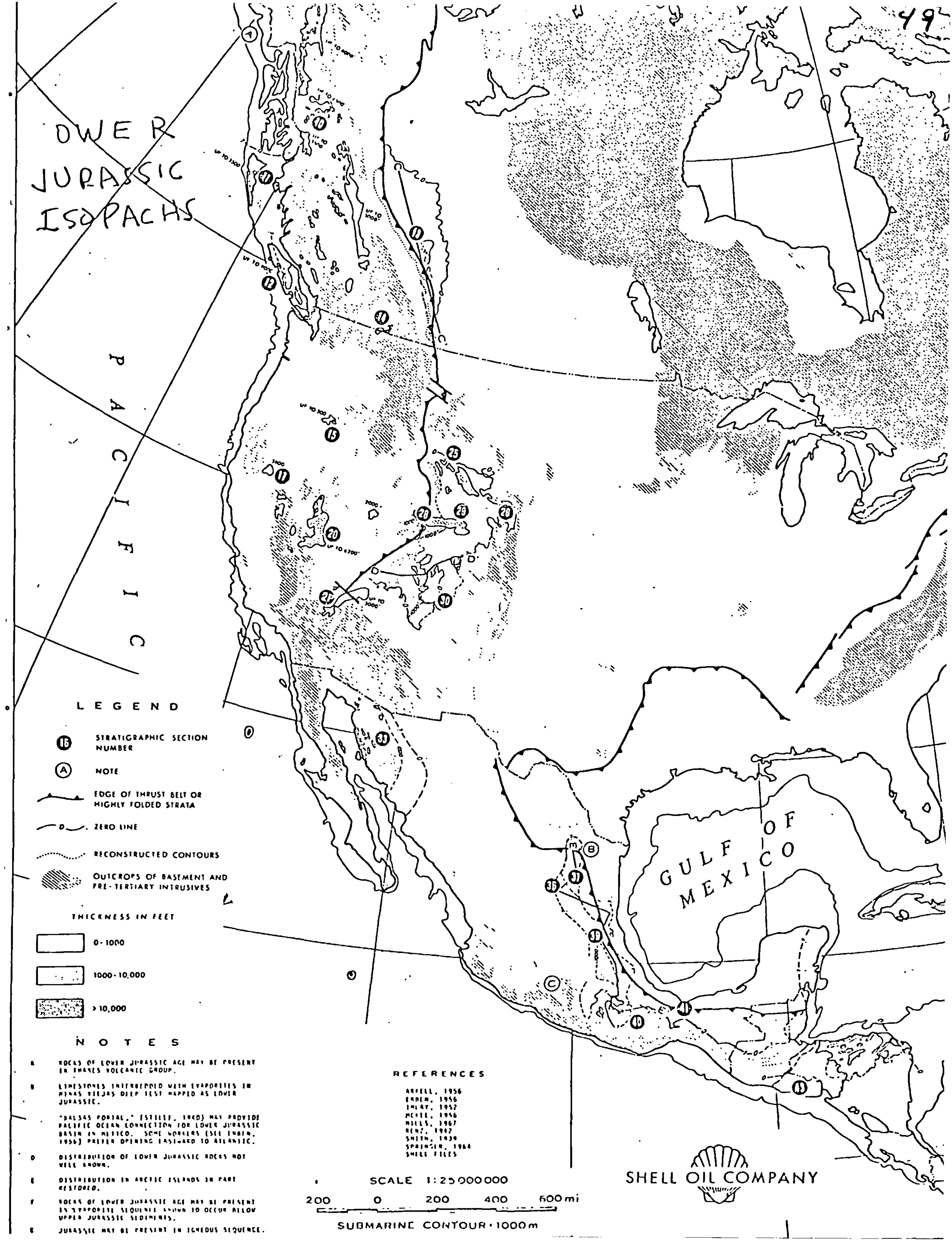
114° W.
113° W.
IDAHO
112° W.
42° N.
WYOMING
NAVAJO AND
NUGGET
111° W.
110° W.
41° N.
40° N.
NEVADA
COLORADO
39° N.
38° N.
37°
ARIZONA



West Interior
Belt Island (Morrison only)
COMPLEX MARINE FACIES
KLAMATH-OHIA
MESOCORDILLERAN HIGH
JURASSIC
NO KNOWN
EVAPORITES
EDGE
EROSIONAL-TECTONIC EDGE
Edge later marine invasion
Later marine invasion
Mixed
Morrison
Marginal Marine
Area of Ancestral Rockies mainly Prec
Deposits here are equivalent to last marine invasion only
Approximate Deposits
Edge earlier marine invasion
Marginal Marine
THE GREAT SAND PILE
Area of Todilto Limestone (Lake)
EROSIONAL EDGE
Elko
100 0 100 Miles
WESTERN UNITED STATES
U of U Geology Dept.
W. L. STOKES
1971

MORRISON FORMATION
STOKES '75


JURASSIC
OUTCROPS
PACIFIC
OCEAN
GULF OF
MEXICO
NOTE
JURASSIC MAY BE PRESENT IN IGNEOUS SEQUENCE.
PATTERN INCLUDES "MESOZOIC ROCKS" WHICH
PART MAY BE OLDER OR YOUNGER THAN JURASSIC.
LEGEND
EDGE OF THRUST BELT
OR HIGHLY FOLDED STRATA
ZERO EDGE
RECONSTRUCTED CONTOURS
OUTCROPS BASEMENT AND
PRE-TERTIARY INTRUSIVES
OUTCROPS
SHELL OIL COMPANY
SCALE 1:25000000
200 0 200 400 600 mi
SUBMARINE CONTOUR: 1000 m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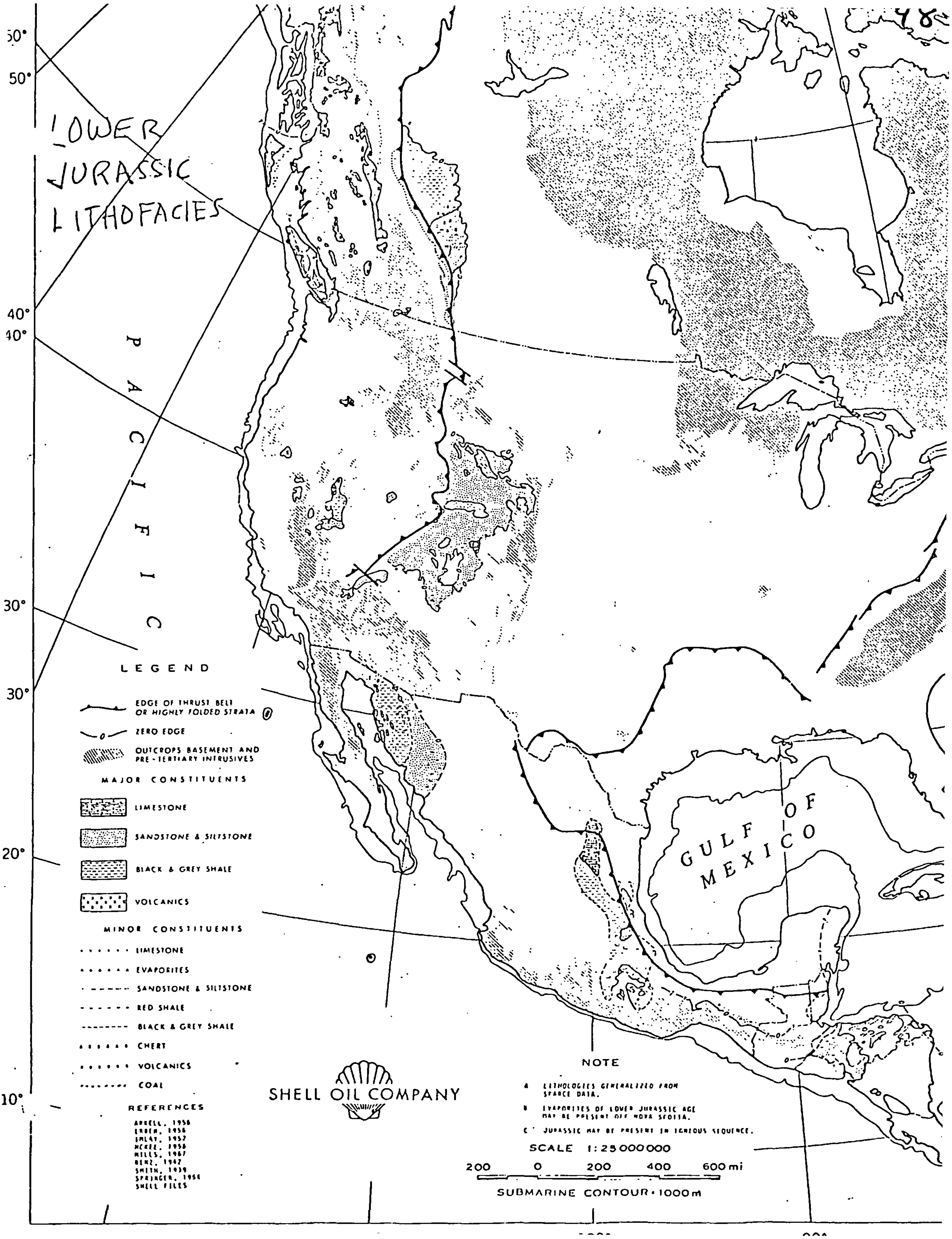
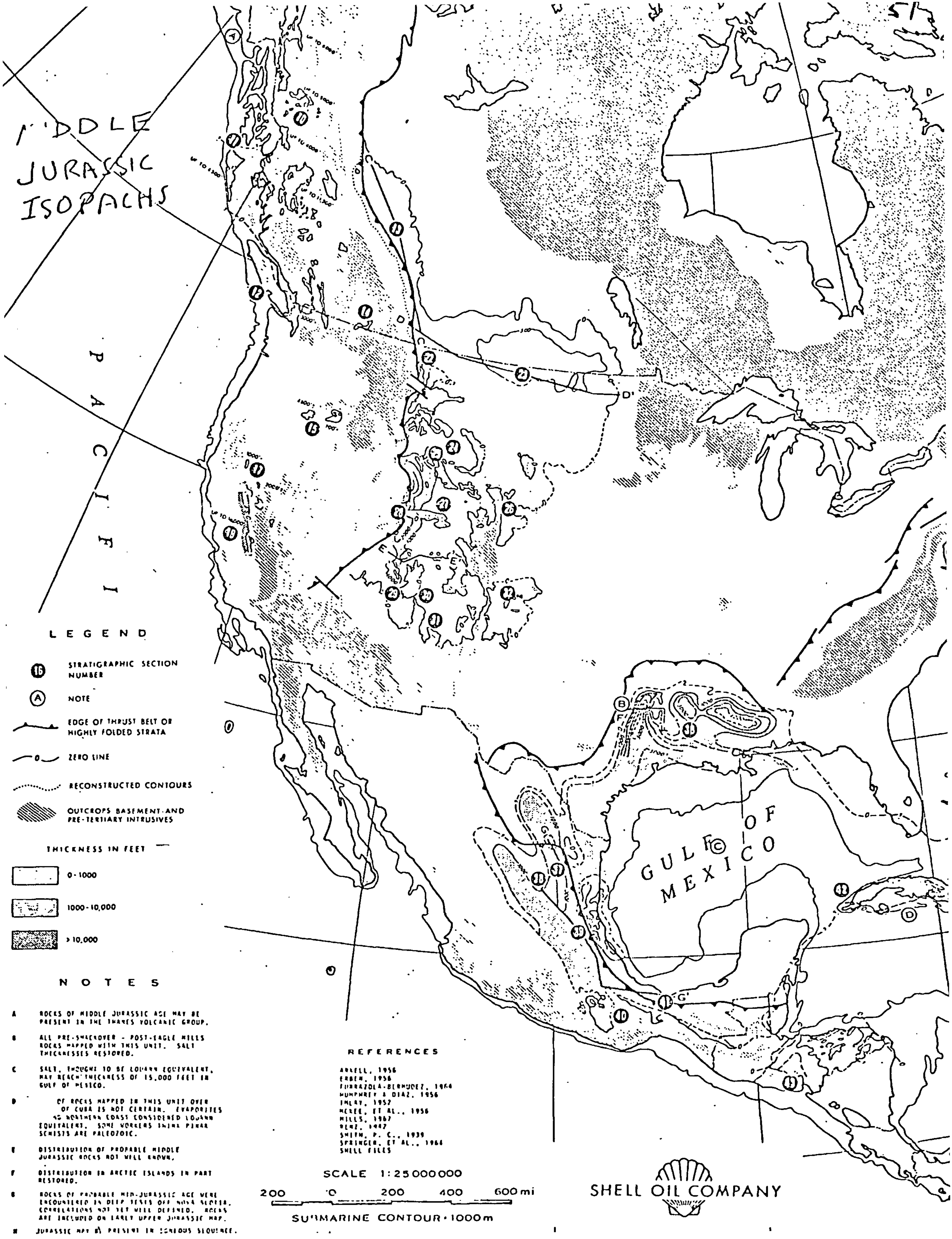
LOWER JURASSIC ISOPACHS
PACIFIC
GULF OF MEXICO
LEGEND
STRATIGRAPHIC SECTION NUMBER
NOTE
EDGE OF THRUST BELT OR HIGHLY FOLDED STRATA
ZERO LINE
RECONSTRUCTED CONTOURS
OUTCROPS OF BASEMENT AND PRE-TERTIARY INTRUSIVES
THICKNESS IN FEET
0-1000
1000-10,000
> 10,000
NOTES
A ROCKS OF LOWER JURASSIC AGE MAY BE PRESENT IN THE VOLCANIC GROUP.
B LIMESTONES INTERBEDDED WITH EVAPORITES IN MINAS VIEJAS DEEP TEST MAPPED AS LOWER JURASSIC.
"BALSAS PORTAL," (STILLE, 1960) MAY PROVIDE PACIFIC OCEAN CONNECTION FOR LOWER JURASSIC BASIN IN MEXICO. SOME WORKERS (SEE IMBEN, 1956) PREFER OPENING EASTWARD TO ATLANTIC.
D DISTRIBUTION OF LOWER JURASSIC ROCKS NOT WELL KNOWN.
E DISTRIBUTION IN ARCTIC ISLANDS IN PART RESTORED.
F ROCKS OF LOWER JURASSIC AGE MAY BE PRESENT IN EVAPORITE SEQUENCE KNOWN TO OCCUR BELOW UPPER JURASSIC SEDIMENTS.
G JURASSIC MAY BE PRESENT IN IGNEOUS SEQUENCE.
REFERENCES
SHELL FILES
SCALE 1:25000000
200 0 200 400 600 mi
SUBMARINE CONTOUR · 1000 m
SHELL OIL COMPANY


LOWER JURASSIC LITHOFACIES
PACIFIC
GULF OF MEXICO
50°
40°
30°
20°
10°
LEGEND
EDGE OF THRUST BELT OR HIGHLY FOLDED STRATA
ZERO EDGE
OUTCROPS BASEMENT AND PRE-TERTIARY INTRUSIVES
MAJOR CONSTITUENTS
LIMESTONE
SANDSTONE & SILTSTONE
BLACK & GREY SHALE
VOLCANICS
MINOR CONSTITUENTS
LIMESTONE
EVAPORITES
SANDSTONE & SILTSTONE
RED SHALE
BLACK & GREY SHALE
CHERT
VOLCANICS
COAL
REFERENCES
SHELL FILES
SHELL OIL COMPANY
NOTE
A LITHOLOGIES GENERALIZED FROM SPARCE DATA.
B EVAPORITES OF LOWER JURASSIC AGE MAY BE PRESENT OFF NOVA SCOTIA.
C JURASSIC MAY BE PRESENT IN IGNEOUS SEQUENCE.
SCALE 1:25 000 000
200 0 200 400 600 mi
SUBMARINE CONTOUR · 1000 m



MIDDLE
JURASSIC
ISOPACHS
PACIFIC
GULF OF MEXICO
LEGEND
STRATIGRAPHIC SECTION NUMBER
NOTE
EDGE OF THRUST BELT OR HIGHLY FOLDED STRATA
ZERO LINE
RECONSTRUCTED CONTOURS
OUTCROPS BASEMENT AND PRE-TERTIARY INTRUSIVES
THICKNESS IN FEET
0-1000
1000-10,000
> 10,000
NOTES
A ROCKS OF MIDDLE JURASSIC AGE MAY BE PRESENT IN THE THANES VOLCANIC GROUP.
B ALL PRE-SMACKOVER - POST-EAGLE MILLS ROCKS MAPPED WITH THIS UNIT. SALT THICKNESSES RESTORED.
C SALT, THOUGHT TO BE LOUANN EQUIVALENT, MAY REACH THICKNESS OF 15,000 FEET IN GULF OF MEXICO.
D OF ROCKS MAPPED IN THIS UNIT OVER OF CUBA IS NOT CERTAIN. EVAPORITES NORTHERN COAST CONSIDERED LOUANN EQUIVALENT. SOME WORKERS THINK PINAR SCHISTS ARE PALEOZOIC.
E DISTRIBUTION OF PROBABLE MIDDLE JURASSIC ROCKS NOT WELL KNOWN.
F DISTRIBUTION IN ARCTIC ISLANDS IN PART RESTORED.
G ROCKS OF PROBABLE MID-JURASSIC AGE WERE ENCOUNTERED IN DEEP TESTS OFF NOVA SCOTIA. CORRELATIONS NOT YET WELL DEFINED. ROCKS ARE INCLUDED ON EARLY UPPER JURASSIC MAP.
* JURASSIC MAY BE PRESENT IN IGNEOUS SEQUENCE.
REFERENCES
ARKELL, 1956
ERBEN, 1956
FURRAZOLA-BERMUDEZ, 1964
HUMPHREY & DIAZ, 1956
IMLAY, 1952
MCKEE, ET AL., 1956
MILLS, 1967
RENZ, 1947
SMITH, P. C., 1939
SPRINGER, ET AL., 1964
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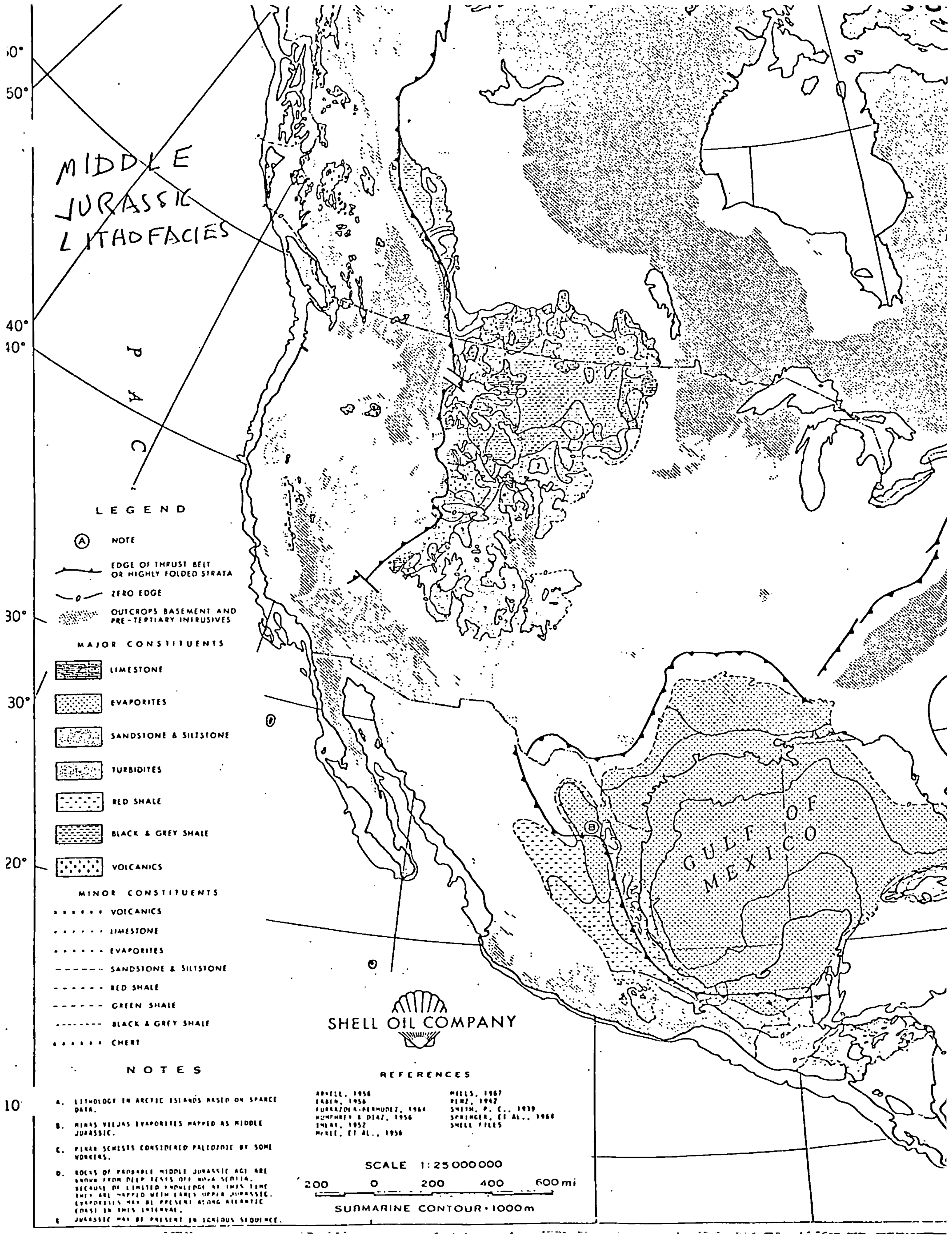
SUBMARINE CONTOUR · 1000m
SHELL OIL COMPANY


MIDDLE
JURASSIC
LITHOFACIES
P
A
C
GULF OF MEXICO
B
LEGEND
A NOTE
EDGE OF THRUST BELT OR HIGHLY FOLDED STRATA
ZERO EDGE
OUTCROPS BASEMENT AND PRE-TERTIARY INTRUSIVES
MAJOR CONSTITUENTS
LIMESTONE
EVAPORITES
SANDSTONE & SILTSTONE
TURBIDITES
RED SHALE
BLACK & GREY SHALE
VOLCANICS
MINOR CONSTITUENTS
VOLCANICS
LIMESTONE
EVAPORITES
SANDSTONE & SILTSTONE
RED SHALE
GREEN SHALE
BLACK & GREY SHALE
CHERT
NOTES
A. LITHOLOGY IN ARCTIC ISLANDS BASED ON SPARCE DATA.
B. MINAS VIEJAS EVAPORITES MAPPED AS MIDDLE JURASSIC.
C. PINAR SCHISTS CONSIDERED PALEOZOIC BY SOME WORKERS.
D. ROCKS OF PROBABLE MIDDLE JURASSIC AGE ARE KNOWN FROM DEEP TESTS OFF NOVA SCOTIA. BECAUSE OF LIMITED KNOWLEDGE AT THIS TIME THEY ARE MAPPED WITH EARLY UPPER JURASSIC. EVAPORITES MAY BE PRESENT ALONG ATLANTIC COAST IN THIS INTERVAL.
E. JURASSIC MAY BE PRESENT IN IGNEOUS SEQUENCE.
SHELL OIL COMPANY
REFERENCES
ARKELL, 1956
EARDLEY, 1956
FURRAZOLA-BERMUDEZ, 1964
HUMPHREY & DIAZ, 1956
IMLAY, 1952
McKEE, ET AL., 1956
MILLS, 1967
PENZ, 1942
SMITH, P. C., 1939
SPRINGER, ET AL., 1964
SHELL FILES
SCALE 1:25 000 000
200 0 200 400 600 mi
SUBMARINE CONTOUR · 1000m
50°
50°
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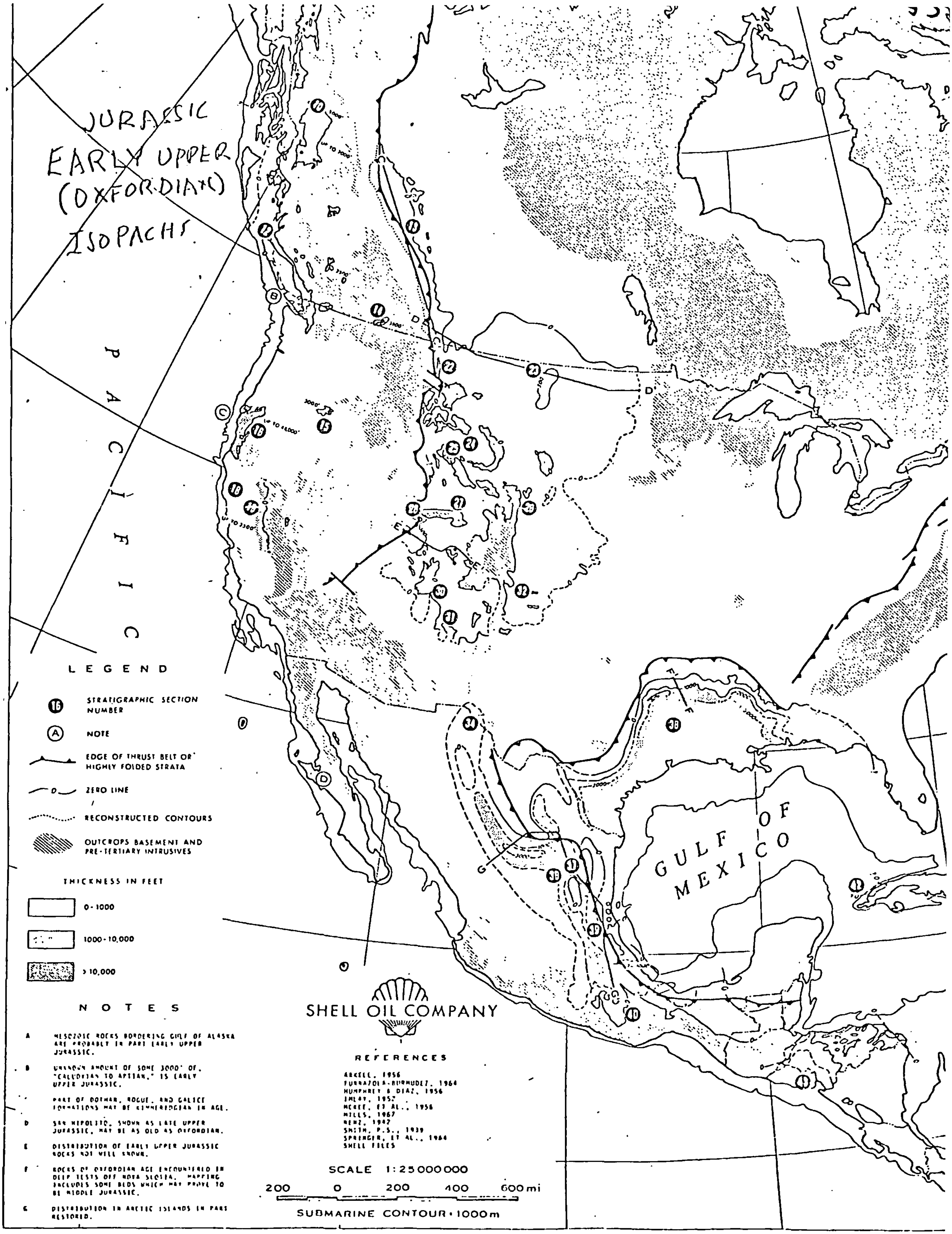
JURASSIC
EARLY UPPER
(OXFORDIAN)
ISOPACHS
PACIFIC
GULF OF MEXICO
LEGEND
STRATIGRAPHIC SECTION NUMBER
NOTE
EDGE OF THRUST BELT OR HIGHLY FOLDED STRATA
ZERO LINE
RECONSTRUCTED CONTOURS
OUTCROPS BASEMENT AND PRE-TERTIARY INTRUSIVES
THICKNESS IN FEET
0-1000
1000-10,000
> 10,000
NOTES
A MESOZOIC ROCKS BORDERING GULF OF ALASKA ARE PROBABLY IN PART EARLY UPPER JURASSIC.
B UNKNOWN AMOUNT OF SOME 3000' OF "CALLOVIAN TO APTIAN," IS EARLY UPPER JURASSIC.
C PART OF DOTHAN, ROGUE, AND GALICE FORMATIONS MAY BE KIMMERIDGIAN IN AGE.
D SAN HIPOLITO, SHOWN AS LATE UPPER JURASSIC, MAY BE AS OLD AS OXFORDIAN.
E DISTRIBUTION OF EARLY UPPER JURASSIC ROCKS NOT WELL KNOWN.
F ROCKS OF OXFORDIAN AGE ENCOUNTERED IN DEEP TESTS OFF NOVA SCOTIA. MAPPING INCLUDES SOME BEDS WHICH MAY PROVE TO BE MIDDLE JURASSIC.
G DISTRIBUTION IN ARCTIC ISLANDS IN PART RESTORED.
SHELL OIL COMPANY
REFERENCES
ARKELL, 1956
FURRAZOLA-BERMUDEZ, 1964
HUMPHREY & DIAZ, 1956
IMLAY, 1952
MCKEE, ET AL., 1956
MILLS, 1967
RENZ, 1942
SMITH, P.S., 1939
SPRINGER, ET AL., 1964
SHELL FILES
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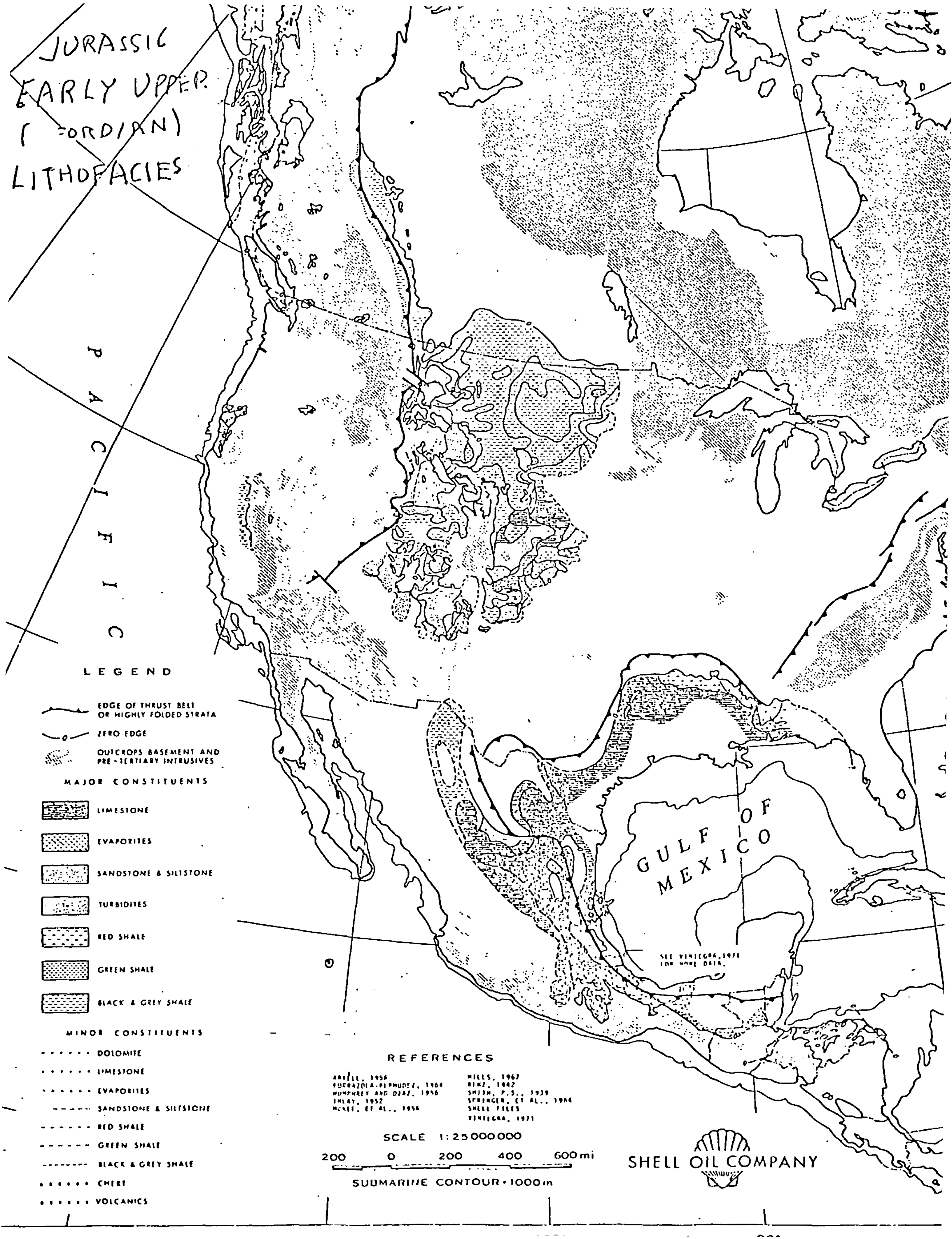
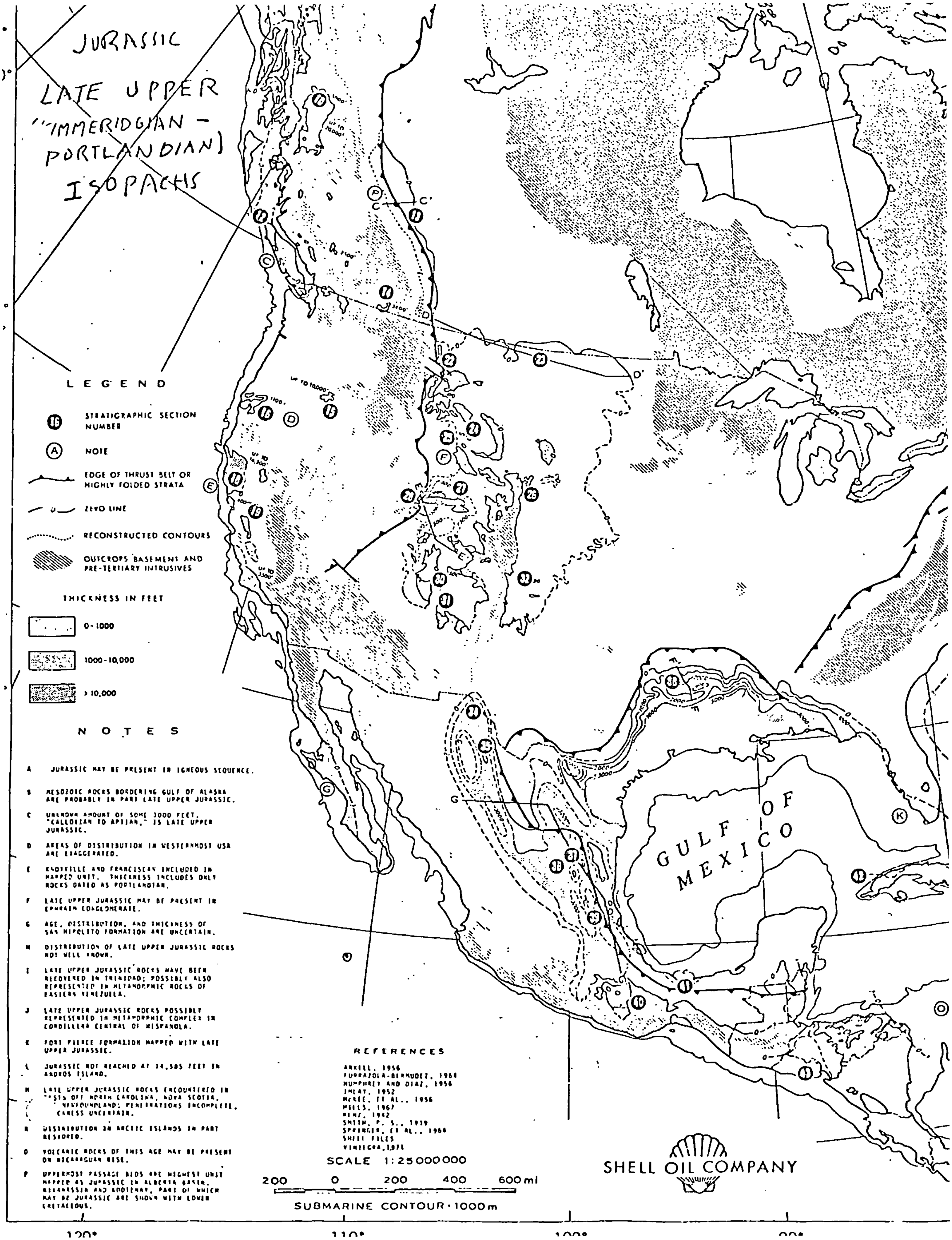
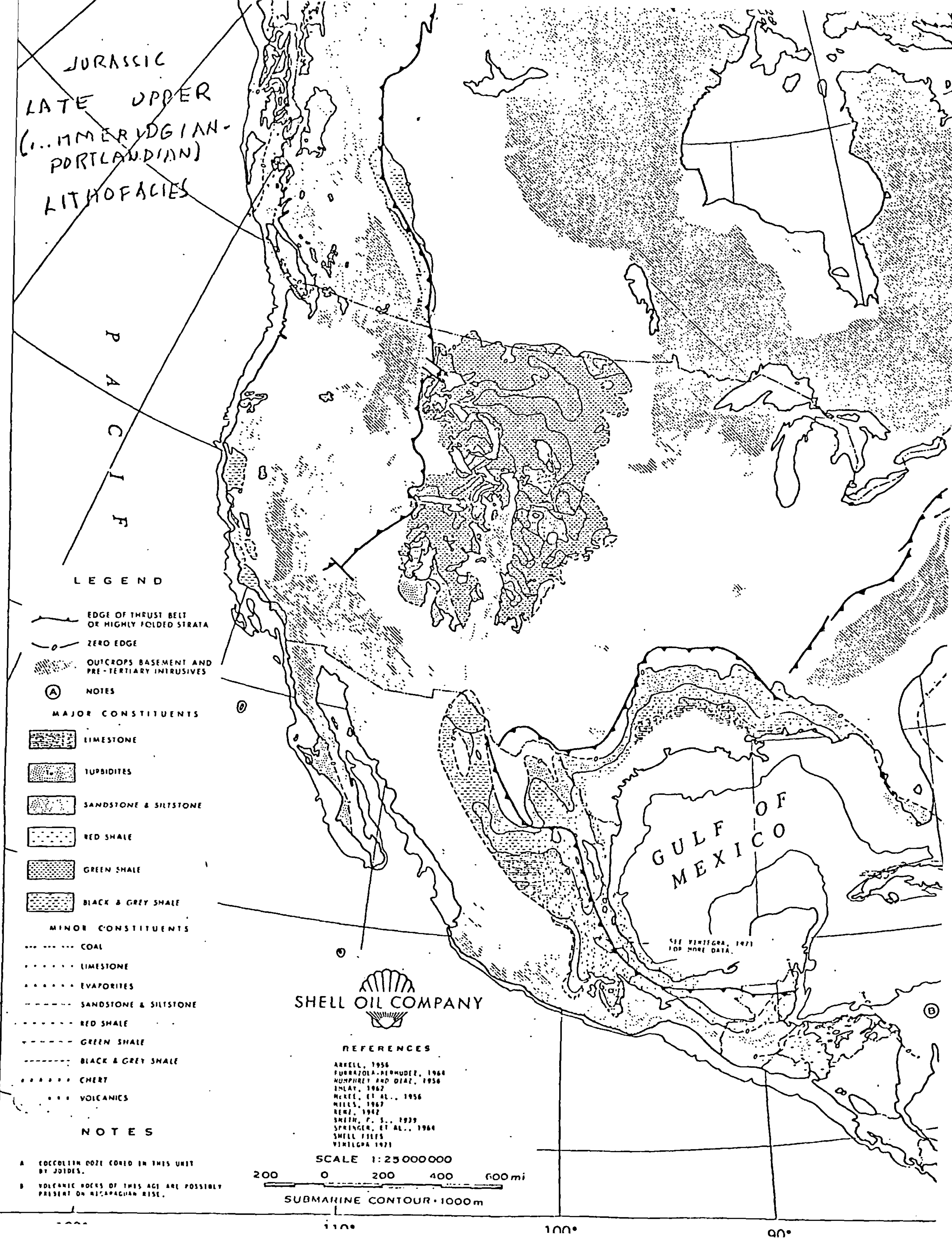
200 0 200 400 600 mi
SUBMARINE CONTOUR - 1000 m



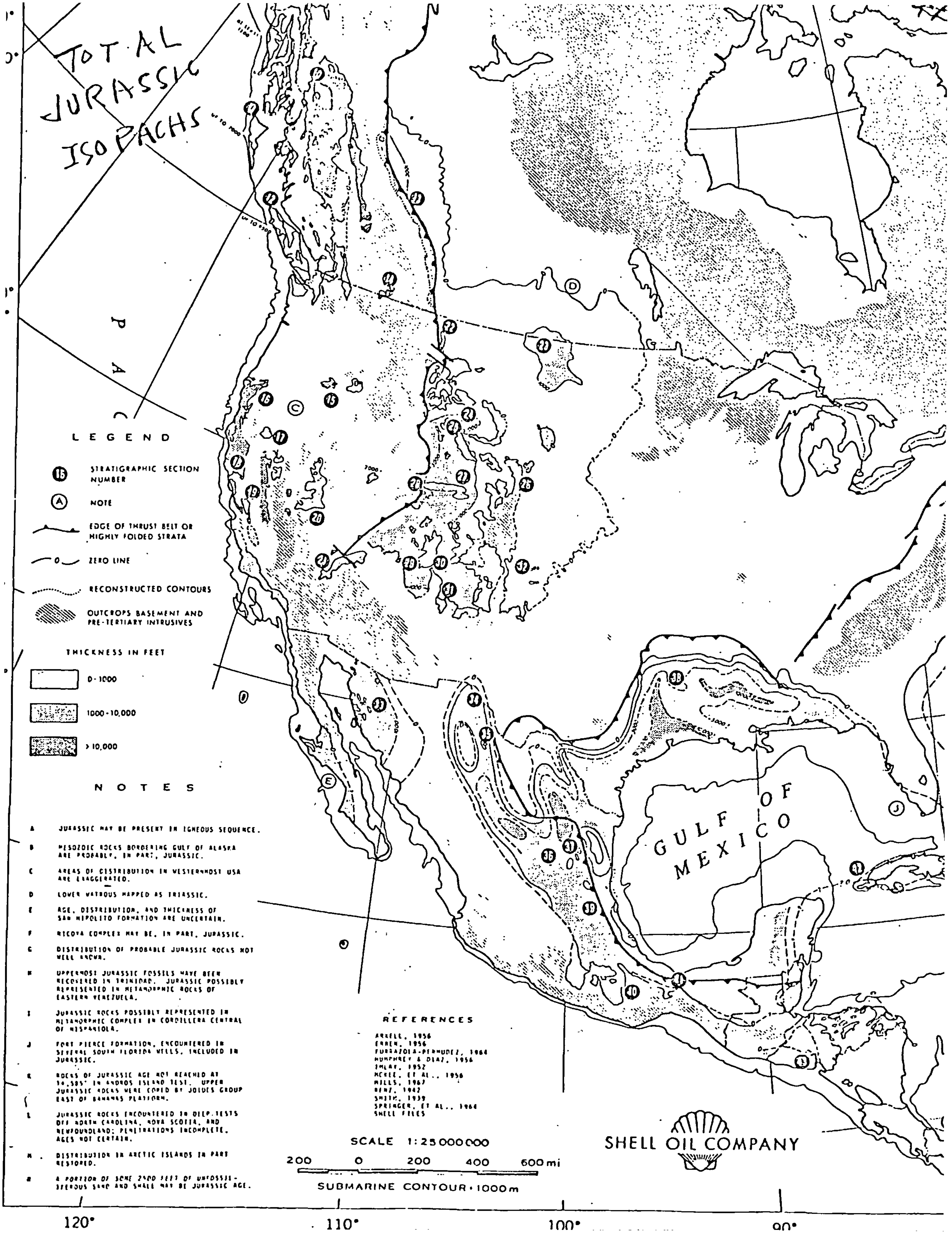
JURASSIC
EARLY UPPER
(ORDIAN)
LITHOFACIES
PACIFIC
GULF OF MEXICO
LEGEND
EDGE OF THRUST BELT OR HIGHLY FOLDED STRATA
ZERO EDGE
OUTCROPS BASEMENT AND PRE-TERTIARY INTRUSIVES
MAJOR CONSTITUENTS
LIMESTONE
EVAPORITES
SANDSTONE & SILTSTONE
TURBIDITES
RED SHALE
GREEN SHALE
BLACK & GREY SHALE
MINOR CONSTITUENTS
DOLOMITE
LIMESTONE
EVAPORITES
SANDSTONE & SILTSTONE
RED SHALE
GREEN SHALE
BLACK & GREY SHALE
CHERT
VOLCANICS
REFERENCES
HUMPHREY AND DIAZ, 1956
IMLAY, 1952
SHELL FILES
VINIEGRA, 1971
SCALE 1:25000000
200 0 200 400 600 mi
SUBMARINE CONTOUR • 1000 m
SHELL OIL COMPANY



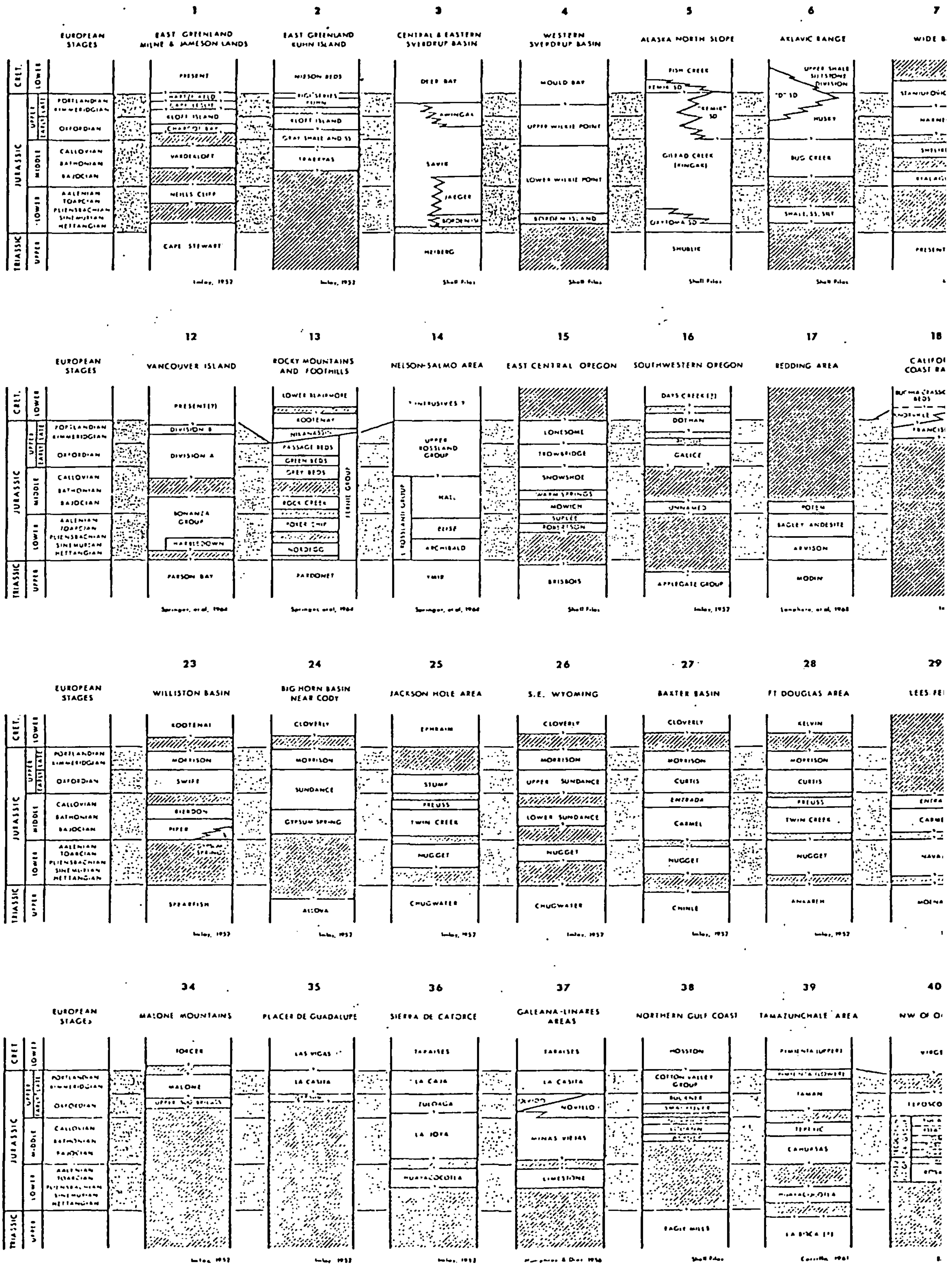
JURASSIC
LATE UPPER
("IMMERIDGIAN – PORTLANDIAN)
ISOPACHS
LEGEND
STRATIGRAPHIC SECTION NUMBER
NOTE
EDGE OF THRUST BELT OR HIGHLY FOLDED STRATA
ZERO LINE
RECONSTRUCTED CONTOURS
OUTCROPS BASEMENT AND PRE-TERTIARY INTRUSIVES
THICKNESS IN FEET
0-1000
1000-10,000
> 10,000
NOTES
A JURASSIC MAY BE PRESENT IN IGNEOUS SEQUENCE.
B MESOZOIC ROCKS BORDERING GULF OF ALASKA ARE PROBABLY IN PART LATE UPPER JURASSIC.
C UNKNOWN AMOUNT OF SOME 3000 FEET, "CALLOVIAN TO APTIAN," IS LATE UPPER JURASSIC.
D AREAS OF DISTRIBUTION IN WESTERNMOST USA ARE EXAGGERATED.
E KNOXVILLE AND FRANCISCAN INCLUDED IN MAPPED UNIT. THICKNESS INCLUDES ONLY ROCKS DATED AS PORTLANDIAN.
F LATE UPPER JURASSIC MAY BE PRESENT IN EPHRAIM CONGLOMERATE.
G AGE, DISTRIBUTION, AND THICKNESS OF SAN HIPOLITO FORMATION ARE UNCERTAIN.
H DISTRIBUTION OF LATE UPPER JURASSIC ROCKS NOT WELL KNOWN.
I LATE UPPER JURASSIC ROCKS HAVE BEEN RECOVERED IN TRINIDAD; POSSIBLY ALSO REPRESENTED IN METAMORPHIC ROCKS OF EASTERN VENEZUELA.
J LATE UPPER JURASSIC ROCKS POSSIBLY REPRESENTED IN METAMORPHIC COMPLEX IN CORDILLERA CENTRAL OF HISPANOLA.
K FORT PIERCE FORMATION MAPPED WITH LATE UPPER JURASSIC.
L JURASSIC NOT REACHED AT 14,585 FEET IN ANDROS ISLAND.
M LATE UPPER JURASSIC ROCKS ENCOUNTERED IN TESTS OFF NORTH CAROLINA, NOVA SCOTIA, NEWFOUNDLAND; PENETRATIONS INCOMPLETE, THICKNESS UNCERTAIN.
N DISTRIBUTION IN ARCTIC ISLANDS IN PART RESTORED.
O VOLCANIC ROCKS OF THIS AGE MAY BE PRESENT ON NICARAGUAN RISE.
P UPPERMOST PASSAGE BEDS ARE HIGHEST UNIT MAPPED AS JURASSIC IN ALBERTA BASIN. KILANASSIN AND KOOTENAY, PART OF WHICH MAY BE JURASSIC ARE SHOWN WITH LOWER CRETACEOUS.
REFERENCES
ARKELL, 1956
FURRAZOLA-BERMUDEZ, 1964
HUMPHREY AND DIAZ, 1956
IMLAY, 1952
McKEE, ET AL., 1956
MILLS, 1967
RINZ, 1942
SMITH, P. S., 1939
SPRINGER, ET AL., 1964
SHELL FILES
VINIEGRA, 1971
SCALE 1:25000000
200 0 200 400 600 ml
SUBMARINE CONTOUR · 1000m
SHELL OIL COMPANY
GULF OF MEXICO
UP TO 10,000'
UP TO 14,300'
UP TO 3,500'
120°
110°
100°
90°

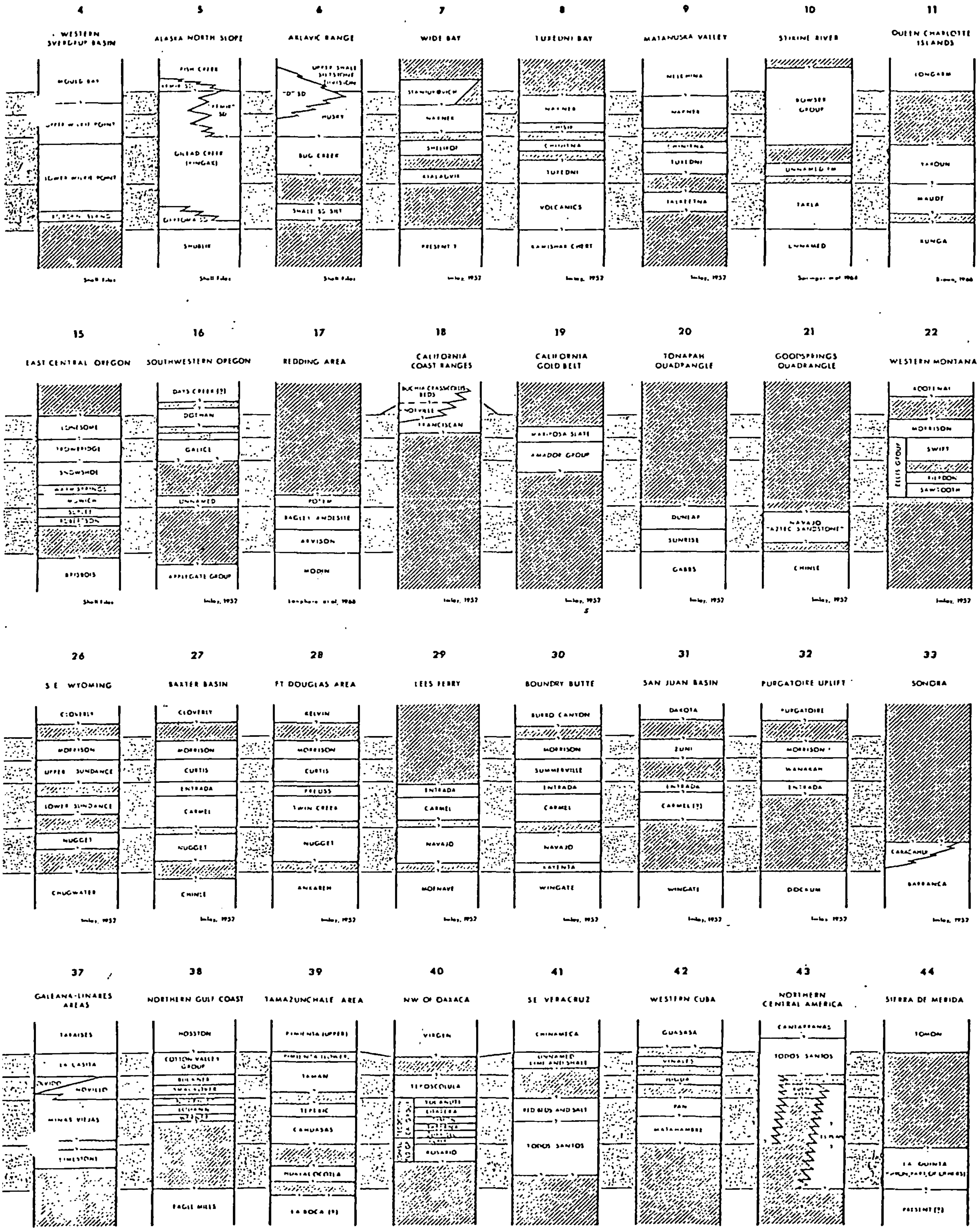


JURASSIC
LATE UPPER
(...MMERIDGIAN-
PORTLANDIAN)
LITHOFACIES
PACIF
GULF OF MEXICO
SEE VINIEGRA, 1971 FOR MORE DATA.
LEGEND
EDGE OF THRUST BELT OR HIGHLY FOLDED STRATA
ZERO EDGE
OUTCROPS BASEMENT AND PRE-TERTIARY INTRUSIVES
A NOTES
MAJOR CONSTITUENTS
LIMESTONE
TURBIDITES
SANDSTONE & SILTSTONE
RED SHALE
GREEN SHALE
BLACK & GREY SHALE
MINOR CONSTITUENTS
COAL
LIMESTONE
EVAPORITES
SANDSTONE & SILTSTONE
RED SHALE
GREEN SHALE
BLACK & GREY SHALE
CHERT
VOLCANICS
NOTES
A COCCOLITH OOZE CORED IN THIS UNIT BY JOIDES.
B VOLCANIC ROCKS OF THIS AGE ARE POSSIBLY PRESENT ON NICARAGUAN RISE.
SHELL OIL COMPANY
REFERENCES
ARKELL, 1956
FURRAZOLA-BERMUDEZ, 1964
HUMPHREY AND DIAZ, 1956
IMLAY, 1962
McKEE, ET AL., 1956
MILLS, 1967
RENZ, 1942
SMITH, F. S., 1939
SPRINGER, ET AL., 1964
SHELL FILES
VINIEGRA 1971
SCALE 1:25000000
200 0 200 400 600 mi
SUBMARINE CONTOUR · 1000 m
110°
100°
90°



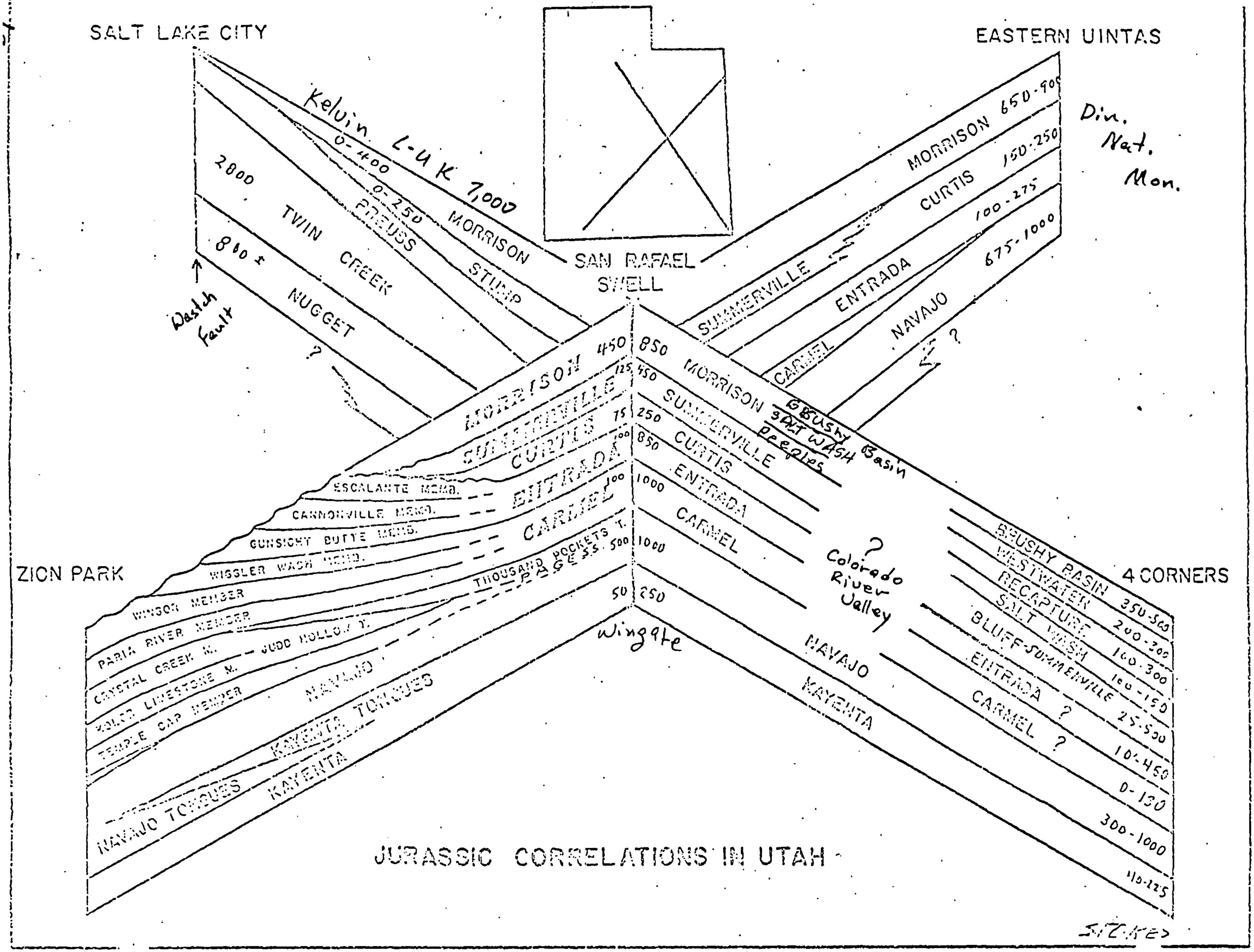
TOTAL JURASSIC ISOPACHS
GULF OF MEXICO
LEGEND
STRATIGRAPHIC SECTION NUMBER
NOTE
EDGE OF THRUST BELT OR HIGHLY FOLDED STRATA
ZERO LINE
RECONSTRUCTED CONTOURS
OUTCROPS BASEMENT AND PRE-TERTIARY INTRUSIVES
THICKNESS IN FEET
0-1000
1000-10,000
> 10,000
NOTES
A JURASSIC MAY BE PRESENT IN IGNEOUS SEQUENCE.
B MESOZOIC ROCKS BORDERING GULF OF ALASKA ARE PROBABLY, IN PART, JURASSIC.
C AREAS OF DISTRIBUTION IN WESTERNMOST USA ARE EXAGGERATED.
D LOWER WATROUS MAPPED AS TRIASSIC.
E AGE, DISTRIBUTION, AND THICKNESS OF SAN HIPOLITO FORMATION ARE UNCERTAIN.
F NICOYA COMPLEX MAY BE, IN PART, JURASSIC.
G DISTRIBUTION OF PROBABLE JURASSIC ROCKS NOT WELL KNOWN.
H UPPERMOST JURASSIC FOSSILS HAVE BEEN RECOVERED IN TRINIDAD. JURASSIC POSSIBLY REPRESENTED IN METAMORPHIC ROCKS OF EASTERN VENEZUELA.
I JURASSIC ROCKS POSSIBLY REPRESENTED IN METAMORPHIC COMPLEX IN CORDILLERA CENTRAL OF HISPANIOLA.
J FORT PIERCE FORMATION, ENCOUNTERED IN SEVERAL SOUTH FLORIDA WELLS, INCLUDED IN JURASSIC.
K ROCKS OF JURASSIC AGE NOT REACHED AT 14,585' IN ANDROS ISLAND TEST. UPPER JURASSIC ROCKS WERE CORED BY JOIDES GROUP EAST OF BAHAMAS PLATFORM.
L JURASSIC ROCKS ENCOUNTERED IN DEEP TESTS OFF NORTH CAROLINA, NOVA SCOTIA, AND NEWFOUNDLAND; PENETRATIONS INCOMPLETE. AGES NOT CERTAIN.
M DISTRIBUTION IN ARCTIC ISLANDS IN PART RESTORED.
N A PORTION OF SOME 2500 FEET OF UNFOSSILIFEROUS SAND AND SHALE MAY BE JURASSIC AGE.
REFERENCES
ARKELL, 1956
ERBEN, 1956
FURRAZOLA-BERMUDEZ, 1964
HUMPHREY & DIAZ, 1956
IMLAY, 1952
MCKEE, ET AL., 1956
MILLS, 1967
RENZ, 1942
SMITH, 1939
SPRINGER, ET AL., 1964
SHELL FILES
SCALE 1:25 000 000
200 0 200 400 600 mi
SUBMARINE CONTOUR · 1000m
SHELL OIL COMPANY
120°
110°
100°
90°


EUROPEAN STAGES
1
EAST GREENLAND MILNE & JAMESON LANDS
2
EAST GREENLAND KUHN ISLAND
3
CENTRAL & EASTERN SVERDRUP BASIN
4
WESTERN SVERDRUP BASIN
5
ALASKA NORTH SLOPE
6
AKLAVIC RANGE
12
VANCOUVER ISLAND
13
ROCKY MOUNTAINS AND FOOTHILLS
14
NELSON-SALMO AREA
15
EAST CENTRAL OREGON
16
SOUTHWESTERN OREGON
17
REDDING AREA
23
WILLISTON BASIN
24
BIG HORN BASIN NEAR CODY
25
JACKSON HOLE AREA
26
S.E. WYOMING
27
BAXTER BASIN
28
FT DOUGLAS AREA
34
MALONE MOUNTAINS
35
PLACER DE GUADALUPE
36
SIERRA DE CATORCE
37
GALEANA-LINARES AREAS
38
NORTHERN GULF COAST
39
TAMAZUNCHALE AREA
CRET.
LOWER
JURASSIC
UPPER
MIDDLE
TRIASSIC
PORTLANDIAN
KIMMERIDGIAN
OXFORDIAN
CALLOVIAN
BATHONIAN
BAJOCIAN
AALENIAN
TOARCIAN
PLIENSBACHIAN
SINEMURIAN
HETTANGIAN
CAPE STEWART
DEER BAY
SAVIK
JAEGER
HEIBERG
MOULD BAY
UPPER WILKIE POINT
LOWER WILKIE POINT
BORDEN ISLAND
FISH CREEK
SHUBLIK
BUG CREEK
HUSKY
DIVISION B
DIVISION A
BONANZA GROUP
HARBLEDOWN
PARSON BAY
KOOTENAY
PASSAGE BEDS
GREEN BEDS
GREY BEDS
ROCK CREEK
POKER CHIP
NORDEGG
FERNIE GROUP
PARDONET
UPPER ROSSLAND GROUP
HALL
ELISE
ARCHIBALD
YMIR
LONESOME
TROWBRIDGE
SNOWSHOE
WARM SPRINGS
MOWICH
SUPLEE
ROBERTSON
BRISBOIS
DAYS CREEK (?)
DOTHAN
GALICE
UNNAMED
APPLEGATE GROUP
POTEM
BAGLEY ANDESITE
ARVISON
MODIN
KOOTENAI
MORRISON
SWIFT
RIERDON
PIPER
SPEARFISH
CLOVERLY
SUNDANCE
GYPSUM SPRING
ALCOVA
EPHRAIM
STUMP
PREUSS
TWIN CREEK
NUGGET
CHUGWATER
UPPER SUNDANCE
LOWER SUNDANCE
CURTIS
ENTRADA
CARMEL
CHINLE
KELVIN
ANKAREH
MALONE
TORCER
LAS VIGAS
LA CASITA
LA CAJA
ZULOAGA
LA JOYA
TARAISES
NOVILLO
MINAS VIEJAS
HUAYACOCOTLA
HOSSTON
COTTON VALLEY GROUP
EAGLE MILLS
TAMAN
TEPEXIC
CAHUASAS
HUAYACOCOTLA
LA BOCA (?)
Imlay, 1952
Shell Files
Springer, et al, 1964
Lanphere, et al, 1968
Carrillo, 1961



4
WESTERN SVERDRUP BASIN
5
ALASKA NORTH SLOPE
6
AKLAVIK RANGE
7
WIDE BAY
8
TUXEDNI BAY
9
MATANUSKA VALLEY
10
STIKINE RIVER
11
QUEEN CHARLOTTE ISLANDS
15
EAST CENTRAL OREGON
16
SOUTHWESTERN OREGON
17
REDDING AREA
18
CALIFORNIA COAST RANGES
19
CALIFORNIA GOLD BELT
20
TONAPAH QUADRANGLE
21
GOODSPRINGS QUADRANGLE
22
WESTERN MONTANA
26
S E WYOMING
27
BAXTER BASIN
28
FT DOUGLAS AREA
29
LEES FERRY
30
BOUNDRY BUTTE
31
SAN JUAN BASIN
32
PURGATOIRE UPLIFT
33
SONORA
37
GALEANA-LINARES AREAS
38
NORTHERN GULF COAST
39
TAMAZUNCHALE AREA
40
NW OF OAXACA
41
SE VERACRUZ
42
WESTERN CUBA
43
NORTHERN CENTRAL AMERICA
44
SIERRA DE MERIDA

1


SALT LAKE CITY
EASTERN UINTAS
SAN RAFAEL SWELL
ZION PARK
4 CORNERS
Kelvin L-U K 7,000
0-400
0-250
PREUSS
MORRISON
STUMP
2800
TWIN CREEK
800 ±
NUGGET
Wasatch Fault
MORRISON 650-900
CURTIS 150-250
100-275
675-1000
SUMMERVILLE
ENTRADA
CARMEL
NAVAJO
Din. Nat. Mon.
MORRISON 450
SUMMERVILLE 125
CURTIS 75
ENTRADA 100
CARMEL 100
500
50
ESCALANTE MEMB.
CANNONVILLE MEMB.
GUNSIGHT BUTTE MEMB.
WIGGLER WASH MEMB.
WINSOR MEMBER
PARIA RIVER MEMBER
CRYSTAL CREEK M.
JUDD HOLLOW T.
KOLOB LIMESTONE M.
TEMPLE CAP MEMBER
THOUSAND POCKETS T.
PAGE S.S.
NAVAJO
KAYENTA TONGUES
KAYENTA
NAVAJO TONGUES
KAYENTA
850 MORRISON
450 SUMMERVILLE
250 CURTIS
850 ENTRADA
1000 CARMEL
1000
250
Wingate
NAVAJO
KAYENTA
Brushy Basin
SALT WASH
Peeples
Colorado River Valley
BRUSHY BASIN 350-500
WESTWATER 200-300
RECAPTURE 100-300
SALT WASH 100-150
BLUFF-SUMMERVILLE 25-500
ENTRADA 100-450
CARMEL 0-150
300-1000
150-225
JURASSIC CORRELATIONS IN UTAH
STOKES

West Uinta Range

From Rocky Mountain Assoc of Geol. 1976 Symposium

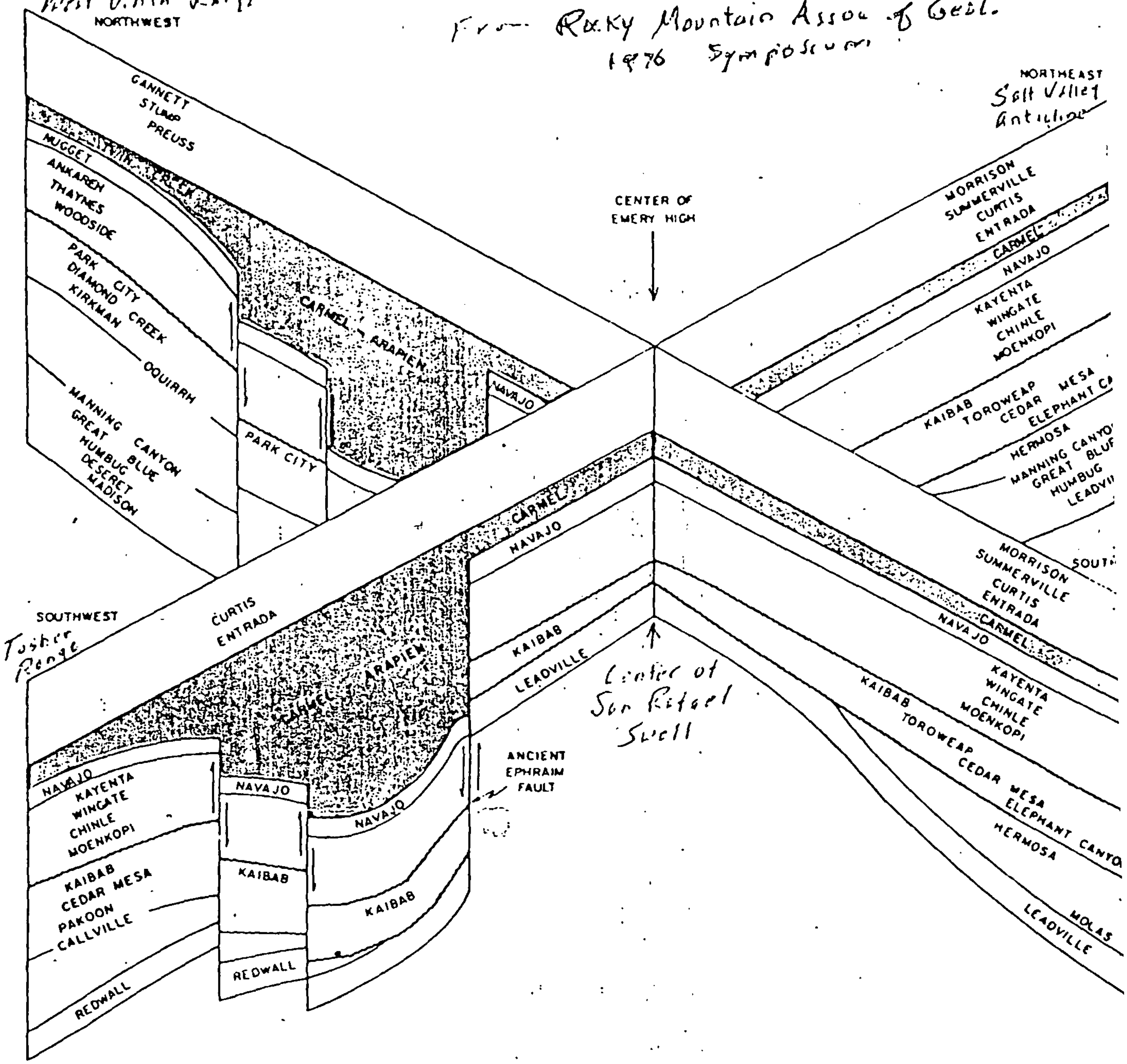


Fig. 14 — Generalized panel cross section showing the "Emery High" after Jurassic deposition.

orogenic belt was elevated in western Utah and southeastern Nevada. The tectonic instability which caused this elongate rift occurred mostly during the Jurassic Carmel-Arapien sedimentary cycle. The presence of salt and anhydrite within a dominantly marine shale sequence indicates that downwarping of the depressed area exceeded the supply of available clastics.

As shown in this panel cross section, the Entrada was not significantly affected by the previous depression.

Some faults, including the ancient Ephraim fault, *have not* had any significant movement since Middle Jurassic time. Structural traps, fault traps and facies traps could be preserved below and within the Carmel-Arapien sediments.

Figures 15, 16, 17 and 18 show the relationship of the Jurassic evaporite basin to the older formations of the Emery uplift. The northwest portion of the uplift is covered by thick evaporites, which have provided an impermeable seal over pre-Jurassic sedimentary rocks.

The Carmel carbonates are oil stained and have had gas shows in some wells in the Sanpete-Sevier rift. Thick oolitic limestone sequences in the Carmel have encountered oil staining and could be important objectives in some areas.

STRUCTURE

The present structural relationship of the eastern part of the study area is shown by a schematic cross section

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Mesozoic-Cenozoic Stratigraphy

Jurassic Topics

Follow general instructions given for the Triassic. Do number one, one from 2,3, and 4, and one from the remainder.

1. Describe the relationship of Jurassic rocks to the Ancestral Rockies and evaluate the uplifts as sediment sources for the period. Make a paleogeographic map on a standard page-size map showing outcrops of the AR at the end of the Jurassic. Add to this the depositional edge of the Morrison Formation with directional arrows showing the probable source of that formation.

2. Describe the Jurassic rocks of western Nevada. What, if any, physical relationships do these have with the Jurassic rocks of the Interior (east of the Wasatch Line)? On what evidence are Late Jurassic rocks of the Pacific margin correlated with those of Utah and Wyoming?

3. Describe the Pre-Morrison Jurassic rocks of the eastern flanks of the Front Range and comparable sections at this longitude in New Mexico. Compare these by means of typical sections with rocks of the same age in central Utah. Explain the differences. (2 sections, one east and one west, will do.)

4. Describe the supposed correlatives of the Navajo Sandstone in southern Arizona, Nevada, and California. What does the distribution of these outliers tell about the origin and dispersal of the Navajo and equivalents? Show your findings on a sketch map (standard map used in this class).

5. Study U.S.G.S. Prof. Paper 556 on the petrology of the Morrison Formation. Comment on this paper as if you were reviewing it for professional colleagues. Look for weak spots. Do you think this is or should be the last word on the subject ?

6. Explain in paleogeographic and sedimentological terms the mesning of the "Great Sand Pile".

7. Summarize in tabular form the evidence for a sub-aqueous origin v. a sub-aerial origin for the Navajo-Nugget Sandstone. List the evidence in parallel columns using headings such as primary structures, paleontology, paleogeography, etc. Mention authors who advocate particular items of evidence.

8. Show by a sequence of columnar sections the evolution of thought about the placement of the Triassic-Jurassic boundary in the Colorado Plateau. Use the format of the hand-out for the Triassic section near Salt Lake City.

9. Study U.S.G.S. Miscellaneous Geological Investigations Map I-175, Paleotectonic Maps of the Jurassic Period. Comment on the manner in which the System is divided for purposes of description and map preparation. Can you suggest improvements? Has anything been added to knowledge of the Jurassic of the united States since this folio was published? Tell what, if anything.

10. Describe the Jurassic rocks of the Black Hills.

W. L. Stokes, 1980

2/14/80

Genera..zations

CRETACEOUS STRATIGRAPHY OF WESTERN INTERIOR

I Historical:

a) Formally named by J. J. d'Omalius d'Halloy, (Belgian), 1822.

b) Widely exposed throughout the United States along Atlantic Coast, Gulf Coast, Western Interior and Pacific Coast. The Western Interior deposits were made known chiefly by early governmental surveys.

II Tectonics and Paleogeography:

a) Increasing importance of Mesocordilleral Highland as a source area for drainage and sediment. This is the dominating tectonic feature of the Cretaceous of the Western Interior. Produced a prodiguous quantity of silty and sandy sediments. Final emergence of all of Nevada and the Sierra Nevada. Pacific border was essentially along the west flanks of the Sierras.

AAPG Vol. 4 #71

b) Emergence the Sevier Orotgenic Belt along the east margins of the Mesocordilleran Highland, immediately west of present Wasatch Lime.

c) No clear evidence of Transcontinental Arch.

d) Final burial of Ancestral Rockies.

e) Development of many basins in the region where Rockes will appear.

f) Offshoots of the Mexican geosyncline extend into Arizona. Calcareous facies.

III Sedimentation:

a) Three major environmental phases: i) early fluvial phase, ii) middle marine phas and iii) final fluvial-swamp phase. Evidently all sediment came from the west.

b) Chief sediment type is siltstone such as constitutes the Mancos Shale, Lewis Shale, Pierre Shale, etc. Origin of this type of sediment is a major problem.

c) Numerous wedges of sandstone representing deltas with superposed offshore bars are characteristic. These thicken and coarsen westward. Their interrelationships constitute the number 1 stratigraphic problem.

d) Coal beds are common in association with the shifting shorelines. The coals overlap but are not cyclothemic except perhaps locally.

e) Carbonate rocks are rare except far from shore in Kansas-Nebraska-South Dakota.

f) Coarse, very thick orogenic conglomerates along western edge of geosyncline. Source easily determined.

g) Bentonites common, chiefly as thin beds, many of which are undisturbed components of the shallow-water deposits.

2/14/80

IV Boundaries

a) Jurassic-Cretaceous in continental beds

b) Lower-Upper is in both marine and non-marine beds. Diverse placement according to groups selected as guides.

c) Cretaceous-Tertiary boundary is a classic problem.

V Economic Geology

a) Coal is the number 1 resource - present in all western states.

b) Gas and oil abundant - especially in sand facies such as the Mesa Verde and the lower part of the marine section.

c) Minor amounts of uranium, bentonite, clay and building stone.

d) Tar sands locally, Asphalt Ridge, near Vernal.

VI Plaeontology

a) Cephalopods are the chief guides- particularly seaphites and baculites. Pelecypods and gastropods locally plentiful but not generally good guides.

b) Microfossils are locally abundant but spotty.

c) Plants are diagnostic and plentiful in association with coal, especially along western margins. Palynology finds valuable application here.

d) Very rare are corals, echinoderms, sponges, arthropods and brachiopods. Probably reflects the non-calcareous silty bottom conditions of the western seaways.

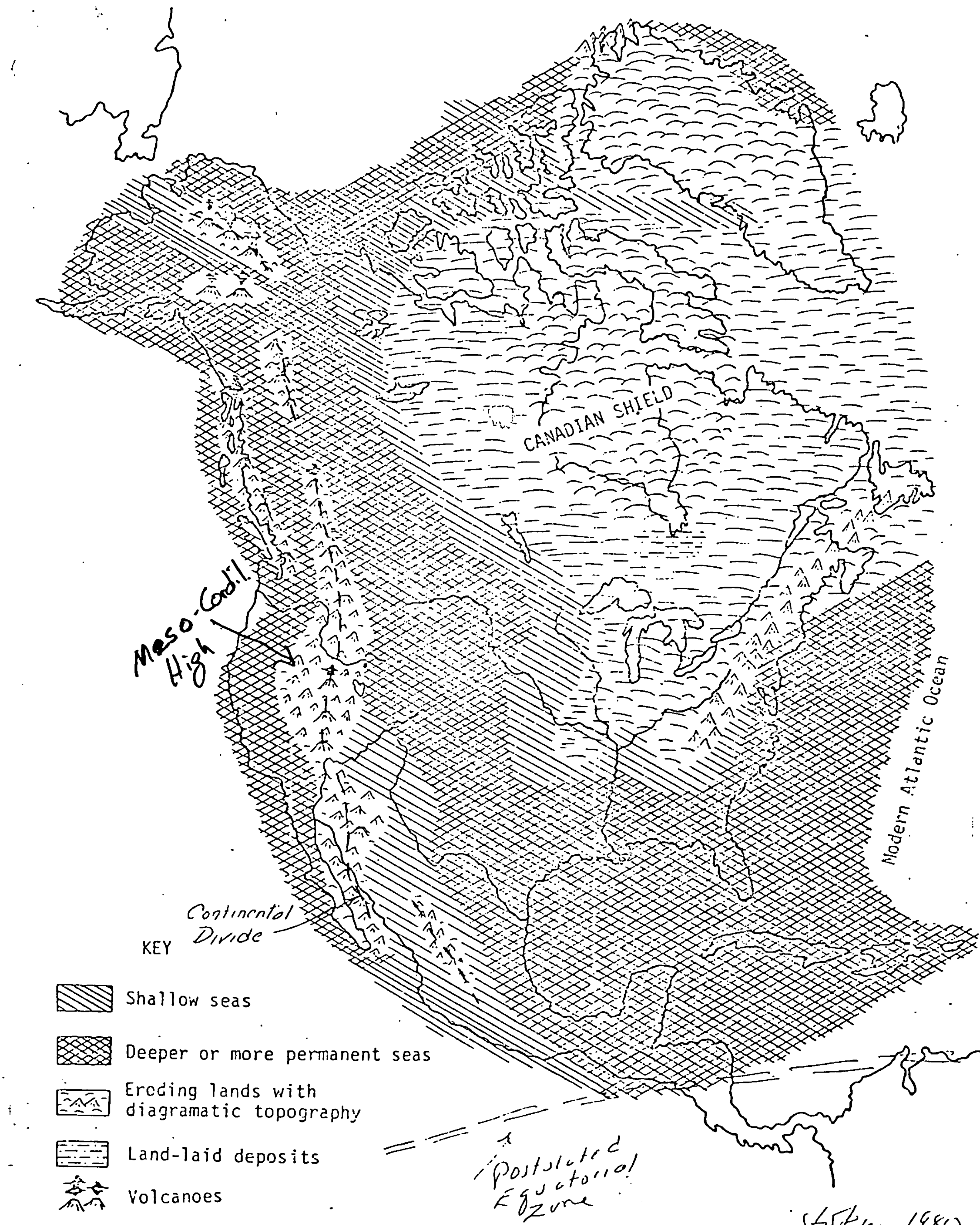
e) All classes of vertebrates are represented with fish and reptiles abundant. Two extensive dinosaur-bearing intervals, one near the beginning (Albian-Aptian) the other near the close (Maestrichtian-Danian). Primitive mammals are found in latest Cretaceous.

VII Scenic

Cretaceous is mainly drab colored. Probably best known for cliffed exposures such as Book Cliffs, and Straight Cliffs. Mesa Verde National Park has many archaeological sites in Cretaceous settings.

W. L. Stokes,

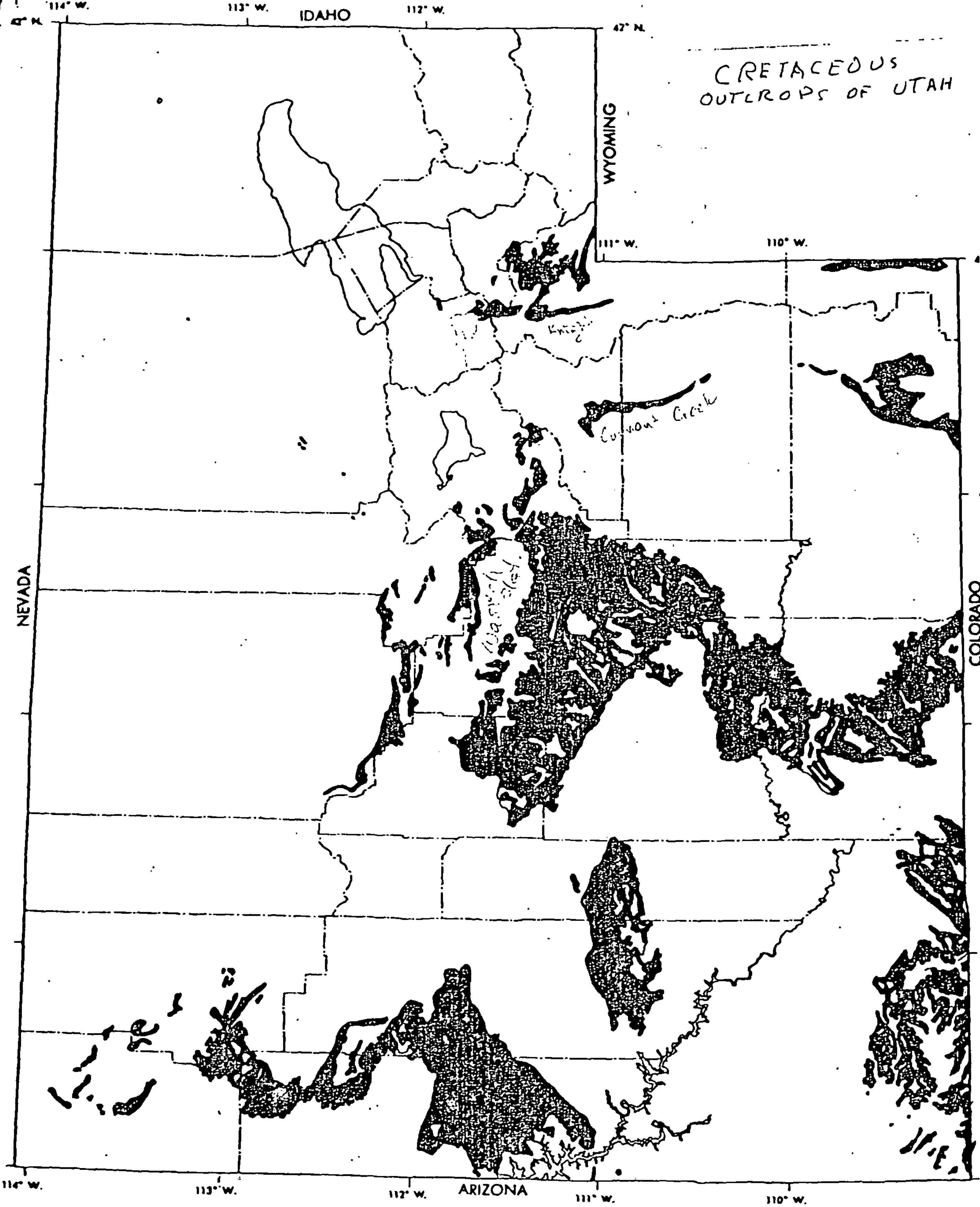
2/8/80 · PALEOGEOGRAPHY OF THE CRETACEOUS PERIOD · 7C



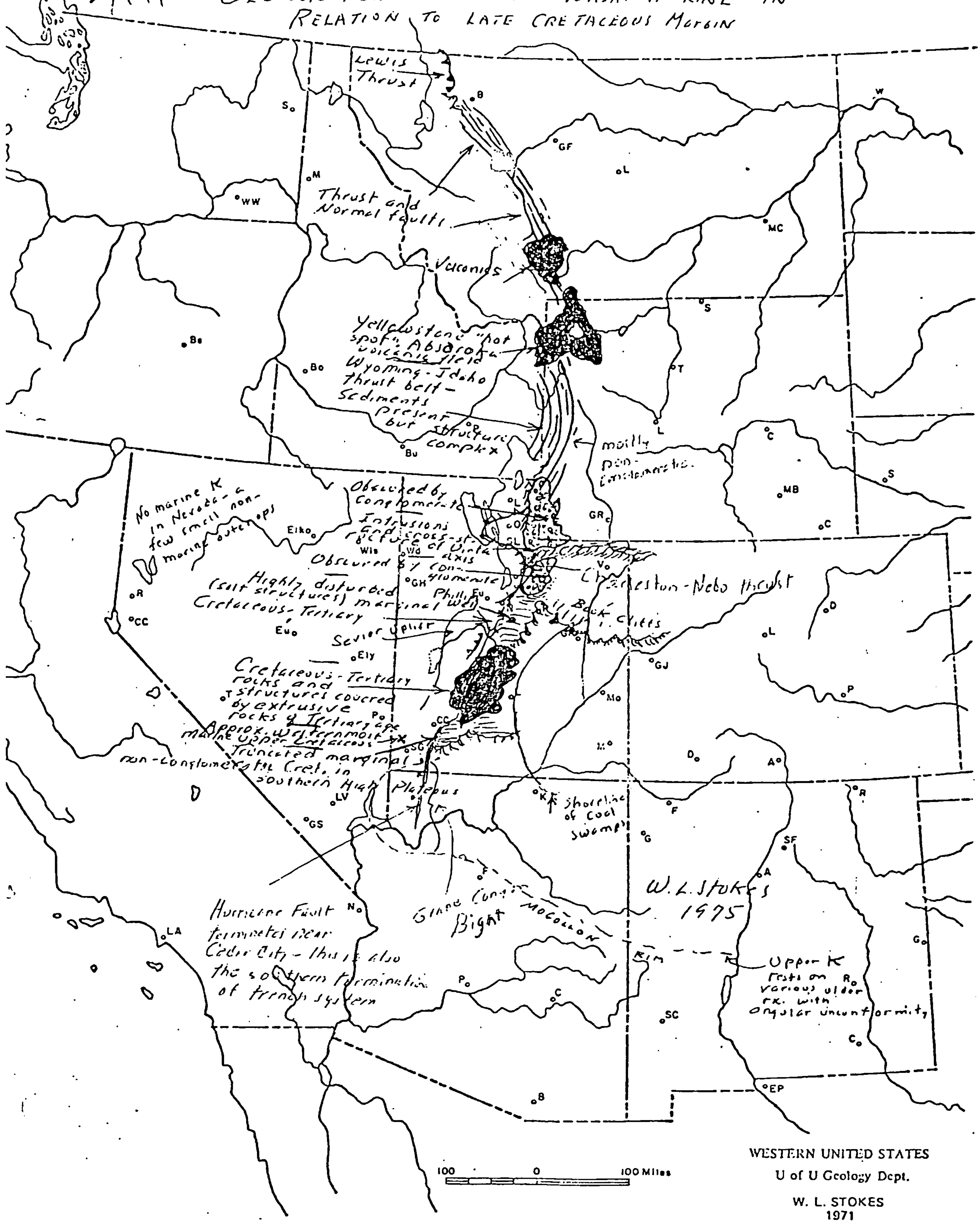
Stokes 1980

2/14/80

CRETACEOUS
OUTCROPS OF UTAH

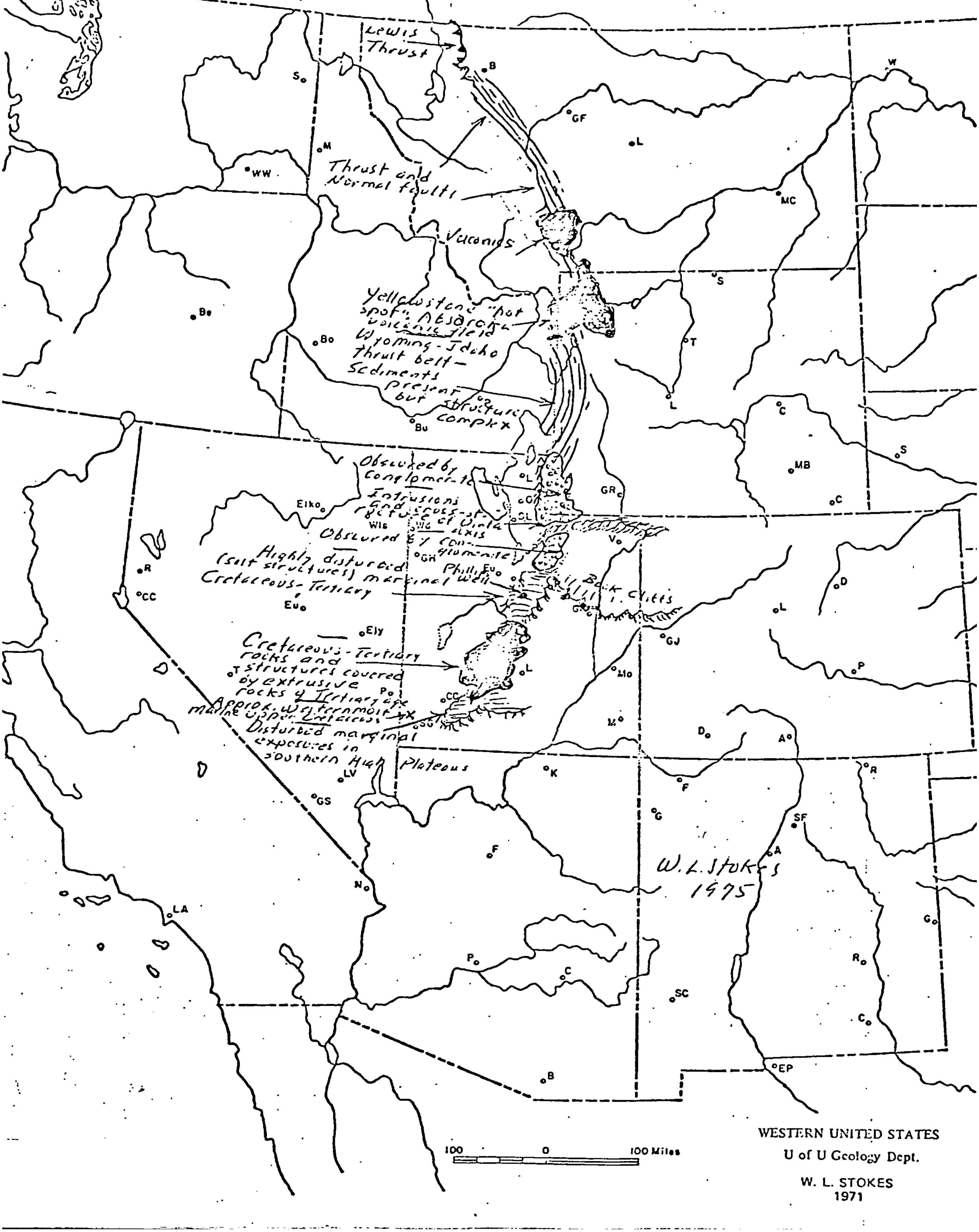


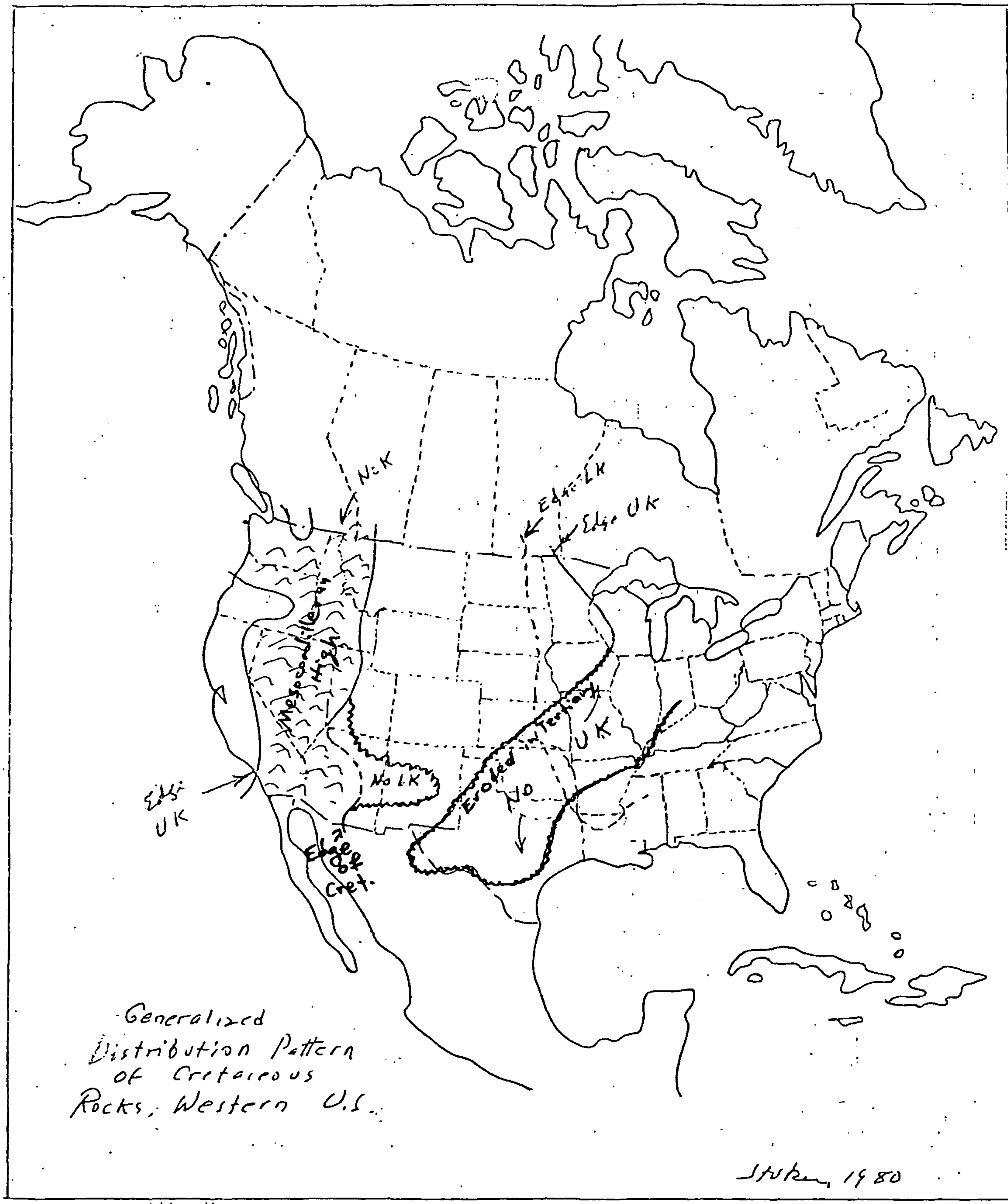
2/14/80 Geologic Features of the Wasatch Line in Relation to Late Cretaceous Margin



WESTERN UNITED STATES
U of U Geology Dept.
W. L. STOKES
1971

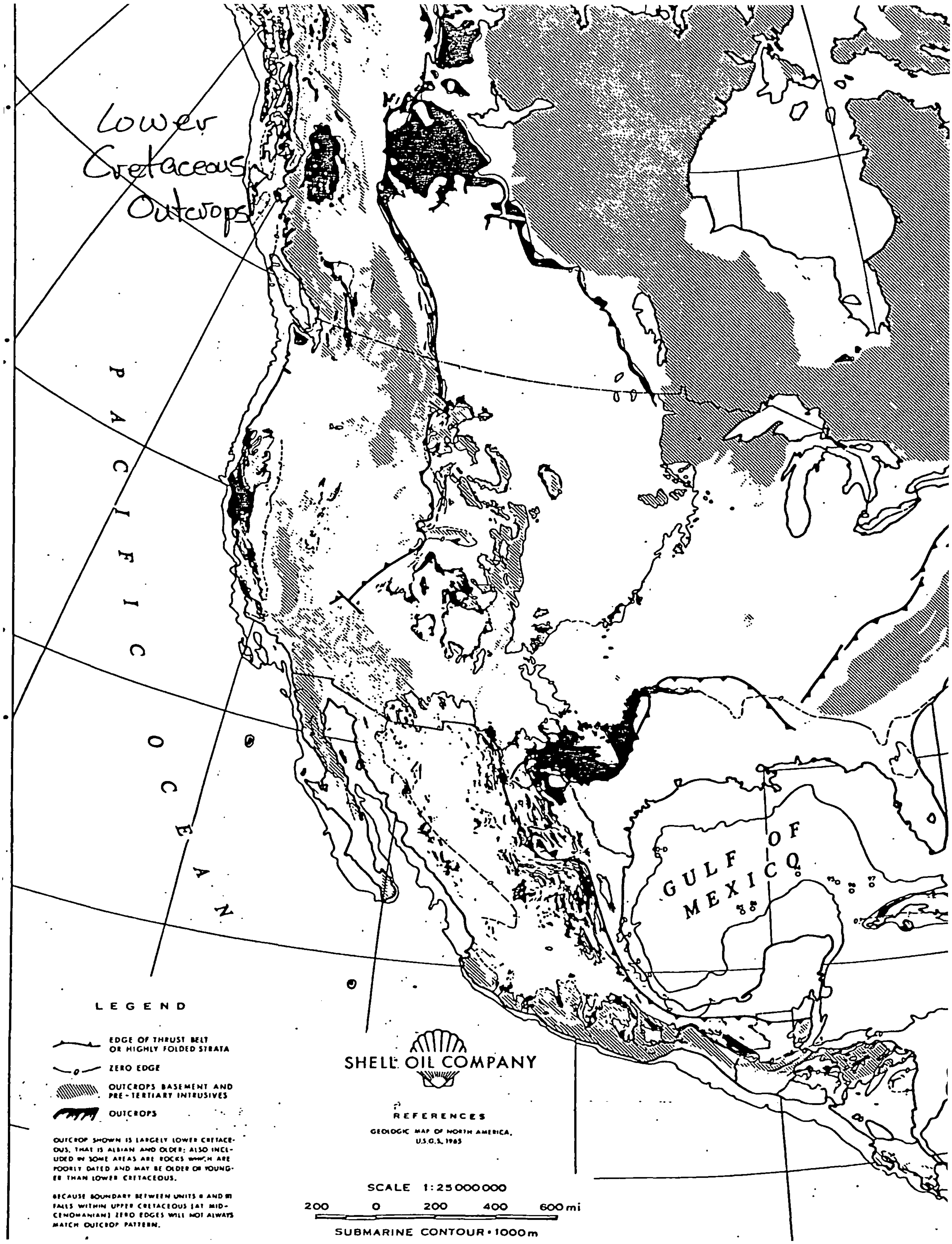
GEOLOGIC FEATURES OF THE WASATCH LINE IN RELATION TO LATE CRETACEOUS MARGIN





Generalized Distribution Pattern of Cretaceous Rocks, Western U.S.

J Stokes, 1980


Lower Cretaceous Outcrops
PACIFIC OCEAN
GULF OF MEXICO
LEGEND
EDGE OF THRUST BELT OR HIGHLY FOLDED STRATA
ZERO EDGE
OUTCROPS BASEMENT AND PRE-TERTIARY INTRUSIVES
OUTCROPS
OUTCROP SHOWN IS LARGELY LOWER CRETACEOUS, THAT IS ALBIAN AND OLDER; ALSO INCLUDED IN SOME AREAS ARE ROCKS WHICH ARE POORLY DATED AND MAY BE OLDER OR YOUNGER THAN LOWER CRETACEOUS.
BECAUSE BOUNDARY BETWEEN UNITS II AND III FALLS WITHIN UPPER CRETACEOUS (AT MID-CENOMANIAN) ZERO EDGES WILL NOT ALWAYS MATCH OUTCROP PATTERN.
SHELL OIL COMPANY
REFERENCES
GEOLOGIC MAP OF NORTH AMERICA, U.S.G.S. 1965
SCALE 1:25000000
200 0 200 400 600 mi
SUBMARINE CONTOUR • 1000m

A. REGRESSIVE SEQUENCE GENERALLY ENDS WITH WIDESPREAD SANDSTONE DEPOSITION (ISACHSEN).

B. [illegible] TURBIDITE SEQUENCE FILLS BASIN FROM SOUTH AND WEST. PROBABLY FIRST COMPLETE MESOZOIC TRANSGRESSION OVER NORTHERN LAND MASS CONNECTING ARCTIC BASIN AND NORTH SLOPE BASIN.

C. CENTRAL AND WESTERN ALASKA RECONSTRUCTED FROM SPARSE DATA.

D. MAPPING INCLUDES PART OF STANIUKOVICH FORMATION WHICH IS PARTLY JURASSIC. YOUNGEST LOWER CRETACEOUS IS WIDESPREAD LIMESTONE (HERENDEEN) PRESENT FROM CAPE DOUGLAS WESTWARD. SHUMAGIN SHALES AND SANDS PRESENT ON SHELF.

E. KENO HILL QUARTZITE MAPPED HERE; DATING UNCERTAIN.

F. DISTRIBUTION AND LITHOFACIES OF LOWER CRETACEOUS ROCKS IN WESTERN CORDILLERA ARE GENERALIZED AND UNDIFFERENTIATED.

G. INTERVAL MAY BE WIDELY PRESENT IN CENTRAL OREGON AND WASHINGTON. SPARSE DATA SUGGEST LITHOLOGY LIKE CALIFORNIA.

H. LITHOLOGY SHOWN FOR FRANCISCAN FACIES IN OUTCROP AREA. AMOUNT OF FRANCISCAN WHICH IS EQUIVALENT TO THIS INTERVAL IS QUESTIONABLE.

I. THICK SEQUENCE OF TUFFS, FLOWS, AND VOLCANIC SANDSTONES IN WESTERN NEVADA IS MOSTLY JURASSIC. PART MAY BE LOWERMOST CRETACEOUS.

J. MAPPING INCLUDES NIKANASSIN AND KOOTENAY WITH LOWER MANNVILLE; SOME PORTION MAY BE JURASSIC.

K. SAND, SHALE AND COALS OF MATTAGAMI FORMATION ASSIGNED HERE; DATING IS NOT CERTAIN.

L. WESTERN BOUNDARY OF MEXICAN GEOSYNCLINE NOT WELL CONTROLLED BECAUSE OF SPARSE OUTCROPS.

M. ALMADAN ISLAND PROBABLY EXISTS THRU THIS INTERVAL.

N. COAHUILA PENINSULA BEGINS TO FOUNDER DURING THIS INTERVAL BUT PROBABLY EXISTS, IN PART, UNTIL MID-APTIAN.

O. SUBSURFACE DATA SUGGEST GRADATION TO SHALE IN DEEP BASIN.

P. SHELF MARGIN SHOWN IN EASTERN GULF IS BASED ON ALBIAN LITHOLOGIES; WESTERN GULF MARGIN BASED ON GOOD SURFACE AND SUBSURFACE CONTROL.

Q. MANY SCATTERED EXPOSURES OF LIMESTONE AND LIMY SHALE OF THIS AGE OVERLIE GNEISS, OFTEN TRANSITIONAL OVER UPPER JURASSIC BEDS.

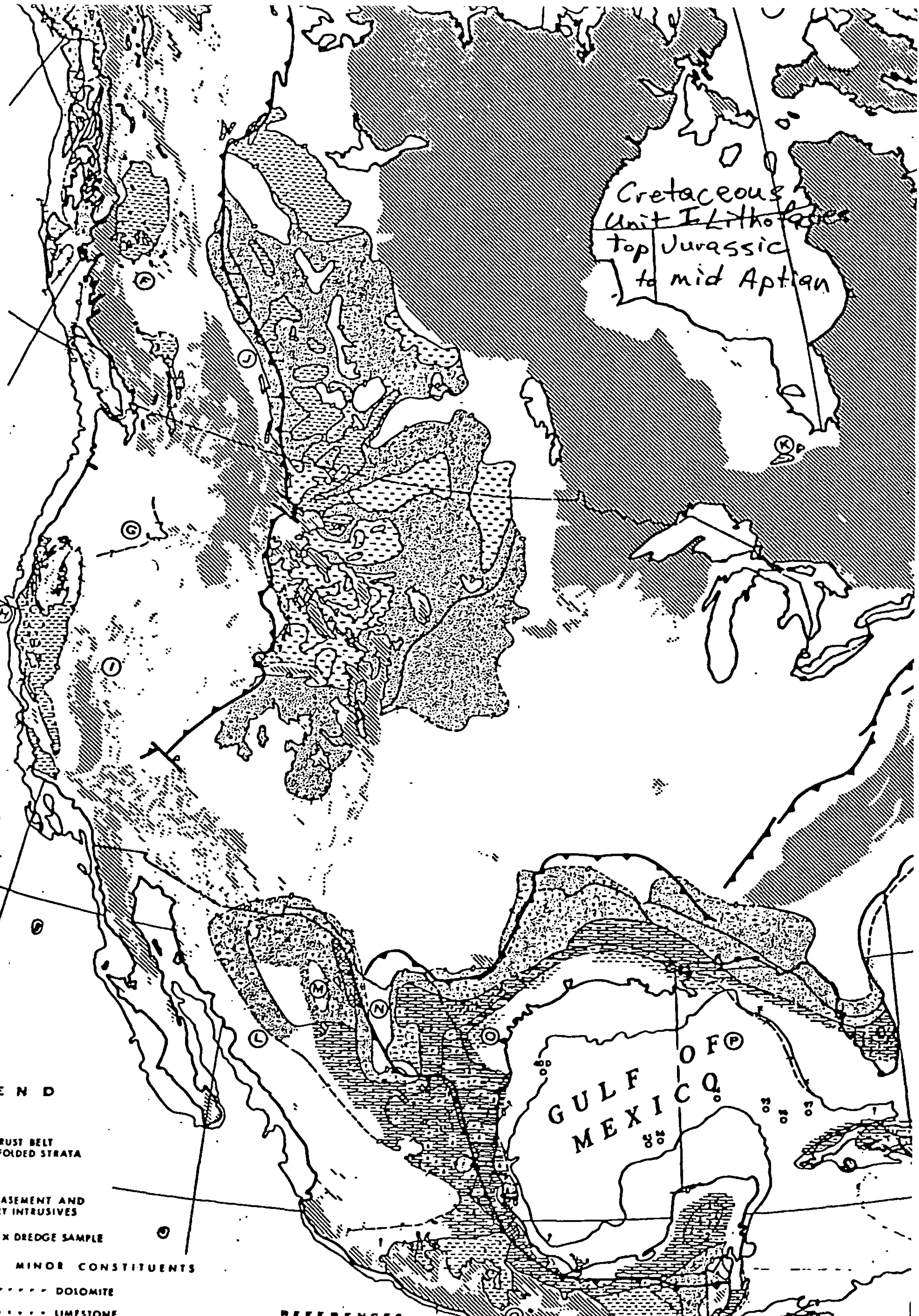
R. UPPERMOST TODOS SANTOS BEDS ARE LOCALLY DATED NEOCOMIAN AND CONTAIN GRAY SHALE AND LIMESTONE, RED SHALES, SANDS AND CONGLOMERATES. OVERLAIN BY GENERALLY THIN WAVY-BEDDED, BLACK LIMESTONE AND SHALE (CANTARRANAS).

S. EARLIEST CRETACEOUS (RIO NEGRO) DEPOSITED IN MAJOR TROUGH IN COLOMBIA WHICH BIFURCATES AND CONTINUES NORTHWARD AND NORTHEASTWARD. CLASTICS GIVE WAY UPWARD TO WIDESPREAD SHALLOW CARBONATES.

T. SHELF MARGINS PLACED FROM SAMPLES OF LATER LOWER CRETACEOUS. POSITION MAY HAVE BEEN SIMILAR IN THIS INTERVAL. JOIDES CORES J-1-4 AND J-1-5 HAVE PELAGIC SEDIMENTS IN THIS INTERVAL.

U. COAL-BEARING BEDS THOUGHT TO BE EQUIVALENT TO THOSE IN SVERDRUP BASIN.

V. RED CLAYS AND SANDS OF PROBABLE FLUVIAL ORIGIN. MAXIMUM DISTRIBUTION UNKNOWN.



LEGEND

(A) NOTES

EDGE OF THRUST BELT OR HIGHLY FOLDED STRATA

ZERO EDGE

OUTCROPS BASEMENT AND PRE-TERTIARY INTRUSIVES

CORE HOLE × DREDGE SAMPLE

MAJOR CONSTITUENTS	MINOR CONSTITUENTS
PLATFORM LIMESTONE	DOLOMITE
PELAGIC LIMESTONE	LIMESTONE
EVAPORITES	EVAPORITES
DOLOMITE	SANDSTONE & SILTSTONE
SANDSTONE & SILTSTONE	RED SHALE
RED SHALE	CONGLOMERATE
BLACK & GREY SHALE	BLACK & GREY SHALE
	CHERT
	VOLCANICS
	COAL
	TURBIDITES

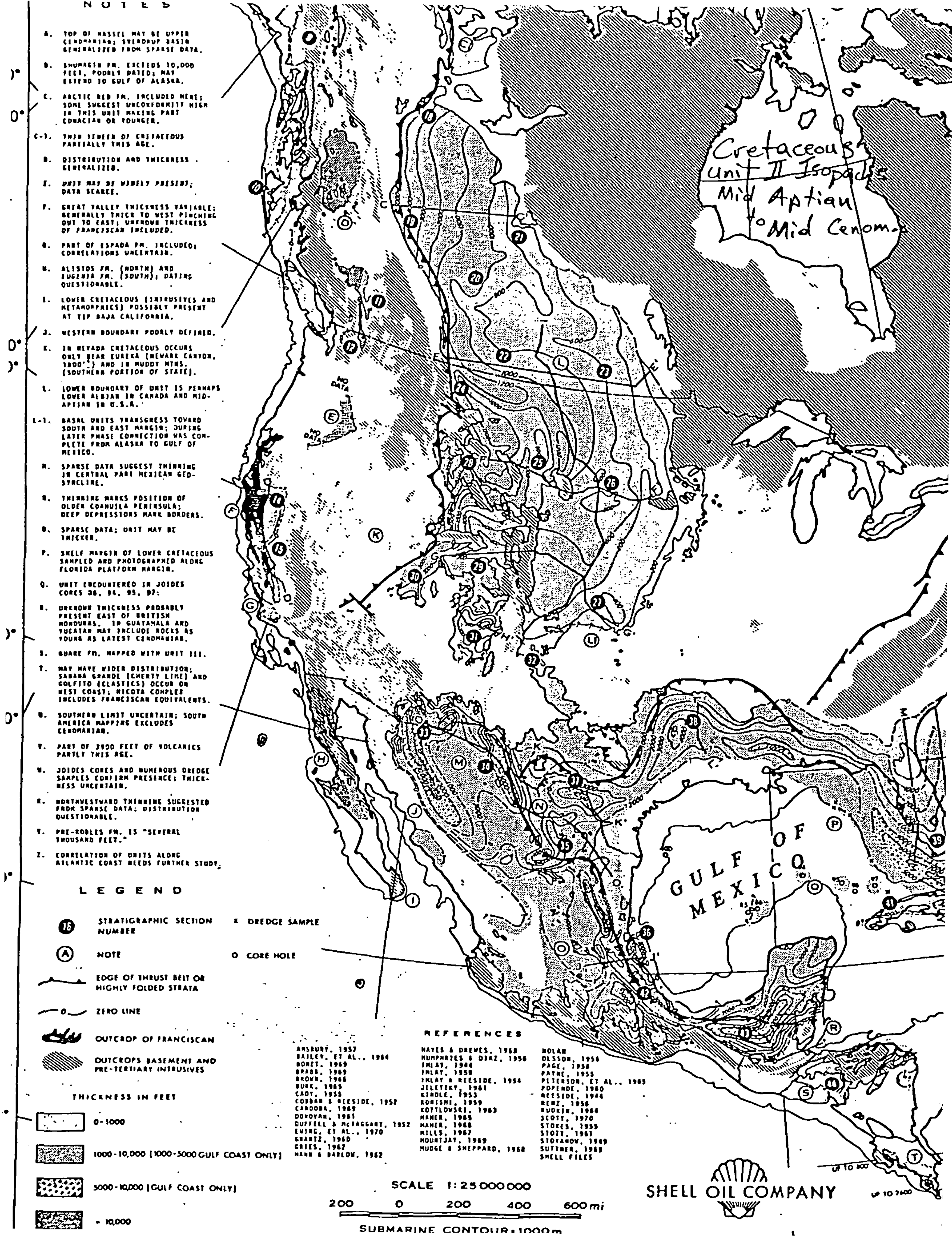
REFERENCES

ANSBURY, 1957
BAILEY, ET AL., 1964
BONET, 1969
BRABB, 1969
BROWN, 1966
BURK, 1965
CADY, 1955
COBBAN & REESIDE, 1952
CORDOBA, 1969
DONOVAN, 1961
DUFFELL & McTAGGART, 1952
EWING, ET AL., 1970
GRANTZ, 1960
GRIES, 1962
HAND & BARLOW, 1962
HAYES & DREWES, 1968
HUMPHREY & DIAZ, 1956
IMLAY, 1944
IMLAY, 1959
IMLAY & REESIDE, 1954
JELETZKY, 1961
KINDLE, 1953
KONISHI, 1959
KOTTLOWSKI, 1963
MAHER, 1965
MAHER, 1968
MILLS, 1967
MOURTJAY, 1969
MUDGE & SHEPPARD, 1968
NOLAN, ET AL., 1956
OLSSON, 1956
PAGE, 1956
PAYNE, 1955
PETERSON, ET AL., 1965
POPENOE, 1960
REESIDE, 1944
RENZ, 1956
RUDKIN, 1964
SCOTT, 1970
STOKES, 1955
STOTT, 1961
STOYANOW, 1949
SUTTNER, 1969

SHELL OIL COMPANY

SCALE 1:25 000 000

200 0 200 400 600 mi


NOTES
A. TOP OF HASSEL MAY BE UPPER CENOMANIAN; SVERDRUP BASIN GENERALIZED FROM SPARSE DATA.
B. SHUMAGIN FM. EXCEEDS 10,000 FEET, POORLY DATED; MAY EXTEND TO GULF OF ALASKA.
C. ARCTIC RED FM. INCLUDED HERE; SOME SUGGEST UNCONFORMITY HIGH IN THIS UNIT MAKING PART CONIACIAN OR YOUNGER.
C-1. THIN VENEER OF CRETACEOUS PARTIALLY THIS AGE.
D. DISTRIBUTION AND THICKNESS GENERALIZED.
E. UNIT MAY BE WIDELY PRESENT; DATA SCARCE.
F. GREAT VALLEY THICKNESS VARIABLE; GENERALLY THICK TO WEST PINCHING OUT TO EAST; UNKNOWN THICKNESS OF FRANCISCAN INCLUDED.
G. PART OF ESPADA FM. INCLUDED; CORRELATIONS UNCERTAIN.
H. ALISTOS FM. (NORTH) AND EUGENIA FM. (SOUTH); DATING QUESTIONABLE.
I. LOWER CRETACEOUS (INTRUSIVES AND METAMORPHICS) POSSIBLY PRESENT AT TIP BAJA CALIFORNIA.
J. WESTERN BOUNDARY POORLY DEFINED.
K. IN NEVADA CRETACEOUS OCCURS ONLY NEAR EUREKA (NEWARK CANYON, 1800'±) AND IN MUDDY MTNS. (SOUTHERN PORTION OF STATE).
L. LOWER BOUNDARY OF UNIT IS PERHAPS LOWER ALBIAN IN CANADA AND MID-APTIAN IN U.S.A.
L-1. BASAL UNITS TRANSGRESS TOWARD SOUTH AND EAST MARGIN; DURING LATER PHASE CONNECTION WAS COMPLETE FROM ALASKA TO GULF OF MEXICO.
M. SPARSE DATA SUGGEST THINNING IN CENTRAL PART MEXICAN GEOSYNCLINE.
N. THINNING MARKS POSITION OF OLDER COAHUILA PENINSULA; DEEP DEPRESSIONS MARK BORDERS.
O. SPARSE DATA; UNIT MAY BE THICKER.
P. SHELF MARGIN OF LOWER CRETACEOUS SAMPLED AND PHOTOGRAPHED ALONG FLORIDA PLATFORM MARGIN.
Q. UNIT ENCOUNTERED IN JOIDES CORES 36, 94, 95, 97.
R. UNKNOWN THICKNESS PROBABLY PRESENT EAST OF BRITISH HONDURAS. IN GUATAMALA AND YUCATAN MAY INCLUDE ROCKS AS YOUNG AS LATEST CENOMANIAN.
S. QUARE FM. MAPPED WITH UNIT III.
T. MAY HAVE WIDER DISTRIBUTION; SABANA GRANDE (CHERTY LIME) AND GOLFITO (CLASTICS) OCCUR ON WEST COAST; NICOYA COMPLEX INCLUDES FRANCISCAN EQUIVALENTS.
U. SOUTHERN LIMIT UNCERTAIN; SOUTH AMERICA MAPPING EXCLUDES CENOMANIAN.
V. PART OF 3900 FEET OF VOLCANICS PARTLY THIS AGE.
W. JOIDES CORES AND NUMEROUS DREDGE SAMPLES CONFIRM PRESENCE; THICKNESS UNCERTAIN.
X. NORTHWESTWARD THINNING SUGGESTED FROM SPARSE DATA; DISTRIBUTION QUESTIONABLE.
Y. PRE-ROBLES FM. IS "SEVERAL THOUSAND FEET."
Z. CORRELATION OF UNITS ALONG ATLANTIC COAST NEEDS FURTHER STUDY.
LEGEND
STRATIGRAPHIC SECTION NUMBER
NOTE
DREDGE SAMPLE
CORE HOLE
EDGE OF THRUST BELT OR HIGHLY FOLDED STRATA
ZERO LINE
OUTCROP OF FRANCISCAN
OUTCROPS BASEMENT AND PRE-TERTIARY INTRUSIVES
THICKNESS IN FEET
0-1000
1000-10,000 (1000-5000 GULF COAST ONLY)
5000-10,000 (GULF COAST ONLY)
> 10,000
Cretaceous Unit II Isopach
Mid Aptian to Mid Cenom.
NO DATA
GULF OF MEXICO
REFERENCES
AMSBURY, 1957
BAILEY, ET AL., 1964
BONET, 1969
BRABB, 1969
BROWN, 1966
BURK, 1965
CADY, 1955
COBBAN & REESIDE, 1952
CARDOBA, 1969
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MOUNTJAY, 1969
MUDGE & SHEPPARD, 1968
NOLAN
OLSSON, 1956
PAGE, 1956
PAYNE, 1955
PETERSON, ET AL., 1965
POPENOE, 1960
REESIDE, 1944
RENZ, 1956
RUDKIN, 1964
SCOTT, 1970
STOKES, 1955
STOTT, 1961
STOYANOW, 1949
SUTTNER, 1969
SHELL FILES
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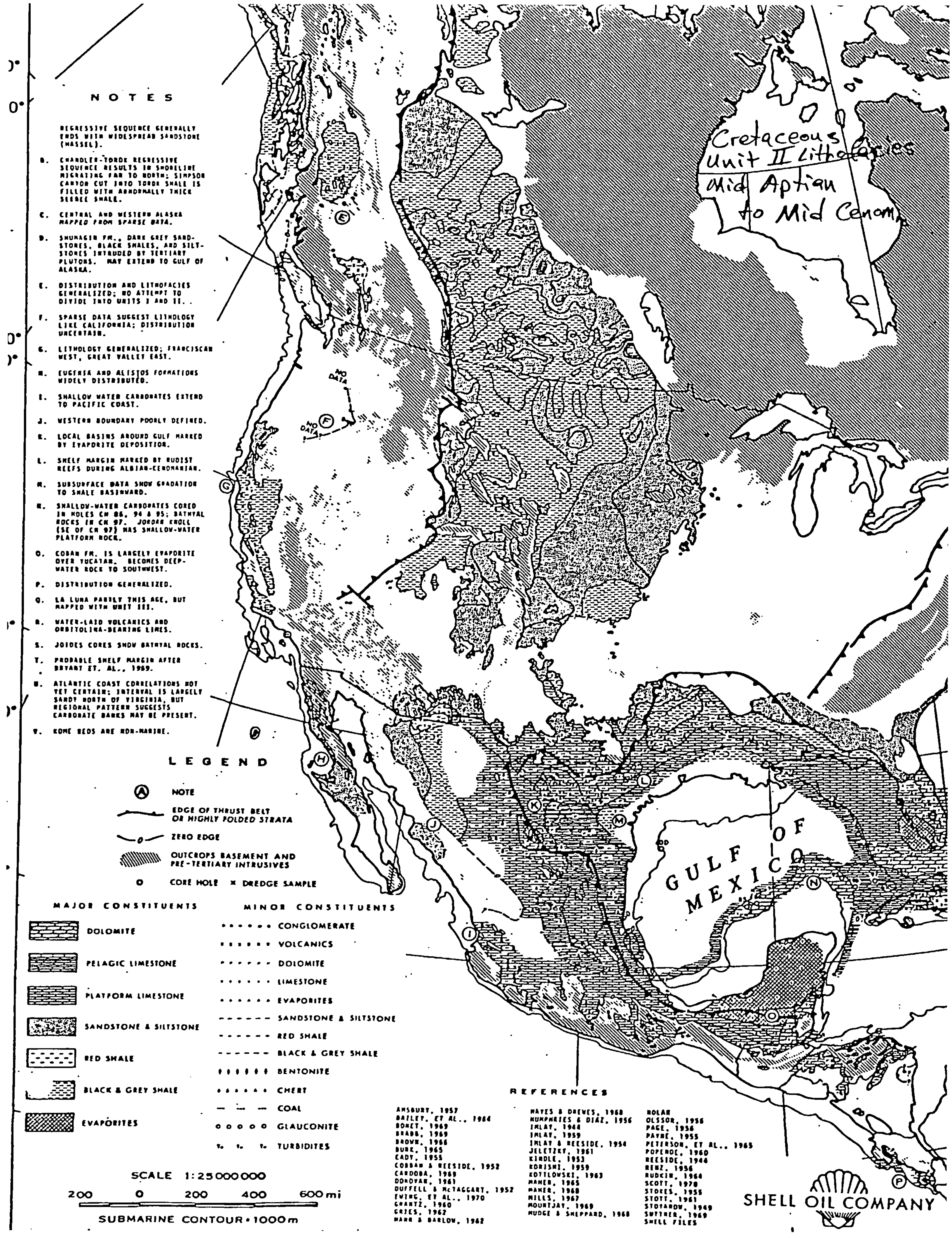
200 0 200 400 600 mi
SUBMARINE CONTOUR: 1000m
SHELL OIL COMPANY
UP TO 800
UP TO 2600


Cretaceous Unit II Lithofacies
Mid Aptian to Mid Cenom.
NOTES
REGRESSIVE SEQUENCE GENERALLY ENDS WITH WIDESPREAD SANDSTONE (NASSEL).
B. CHANDLER-TORDK REGRESSIVE SEQUENCE RESULTS IN SHORELINE MIGRATING FAR TO NORTH; SIMPSON CANYON CUT INTO TORDK SHALE IS FILLED WITH ABNORMALLY THICK SEABEE SHALE.
C. CENTRAL AND WESTERN ALASKA MAPPED FROM SPARSE DATA.
D. SHUMAGIN FM., DARK GREY SANDSTONES, BLACK SHALES, AND SILTSTONES INTRUDED BY TERTIARY PLUTONS. MAY EXTEND TO GULF OF ALASKA.
E. DISTRIBUTION AND LITHOFACIES GENERALIZED; NO ATTEMPT TO DIVIDE INTO UNITS I AND II.
F. SPARSE DATA SUGGEST LITHOLOGY LIKE CALIFORNIA; DISTRIBUTION UNCERTAIN.
G. LITHOLOGY GENERALIZED; FRANCISCAN WEST, GREAT VALLEY EAST.
H. EUGENIA AND ALISTOS FORMATIONS WIDELY DISTRIBUTED.
I. SHALLOW WATER CARBONATES EXTEND TO PACIFIC COAST.
J. WESTERN BOUNDARY POORLY DEFINED.
K. LOCAL BASINS AROUND GULF MARKED BY EVAPORITE DEPOSITION.
L. SHELF MARGIN MARKED BY RUDIST REEFS DURING ALBIAN-CENOMANIAN.
M. SUBSURFACE DATA SHOW GRADATION TO SHALE BASINWARD.
N. SHALLOW-WATER CARBONATES CORED IN HOLES CH 86, 94 & 95; BATHYAL ROCKS IN CH 97. JORDAN KNOLL (SE OF CH 97) HAS SHALLOW-WATER PLATFORM ROCK.
O. COBAN FM. IS LARGELY EVAPORITE OVER YUCATAN. BECOMES DEEP-WATER ROCK TO SOUTHWEST.
P. DISTRIBUTION GENERALIZED.
Q. LA LUNA PARTLY THIS AGE, BUT MAPPED WITH UNIT III.
R. WATER-LAID VOLCANICS AND ORBITOLINA-BEARING LIMES.
S. JOIDES CORES SHOW BATHYAL ROCKS.
T. PROBABLE SHELF MARGIN AFTER BRYANT ET. AL., 1969.
U. ATLANTIC COAST CORRELATIONS NOT YET CERTAIN; INTERVAL IS LARGELY SANDY NORTH OF VIRGINIA, BUT REGIONAL PATTERN SUGGESTS CARBONATE BANKS MAY BE PRESENT.
V. SOME BEDS ARE NON-MARINE.
NO DATA
NO DATA
GULF OF MEXICO
LEGEND
NOTE
EDGE OF THRUST BELT OR HIGHLY FOLDED STRATA
ZERO EDGE
OUTCROPS BASEMENT AND PRE-TERTIARY INTRUSIVES
CORE HOLE × DREDGE SAMPLE
MAJOR CONSTITUENTS
DOLOMITE
PELAGIC LIMESTONE
PLATFORM LIMESTONE
SANDSTONE & SILTSTONE
RED SHALE
BLACK & GREY SHALE
EVAPORITES
MINOR CONSTITUENTS
CONGLOMERATE
VOLCANICS
DOLOMITE
LIMESTONE
EVAPORITES
SANDSTONE & SILTSTONE
RED SHALE
BLACK & GREY SHALE
BENTONITE
CHERT
COAL
GLAUCONITE
TURBIDITES
SCALE 1:25000000
200 0 200 400 600 mi
SUBMARINE CONTOUR • 1000 m
REFERENCES
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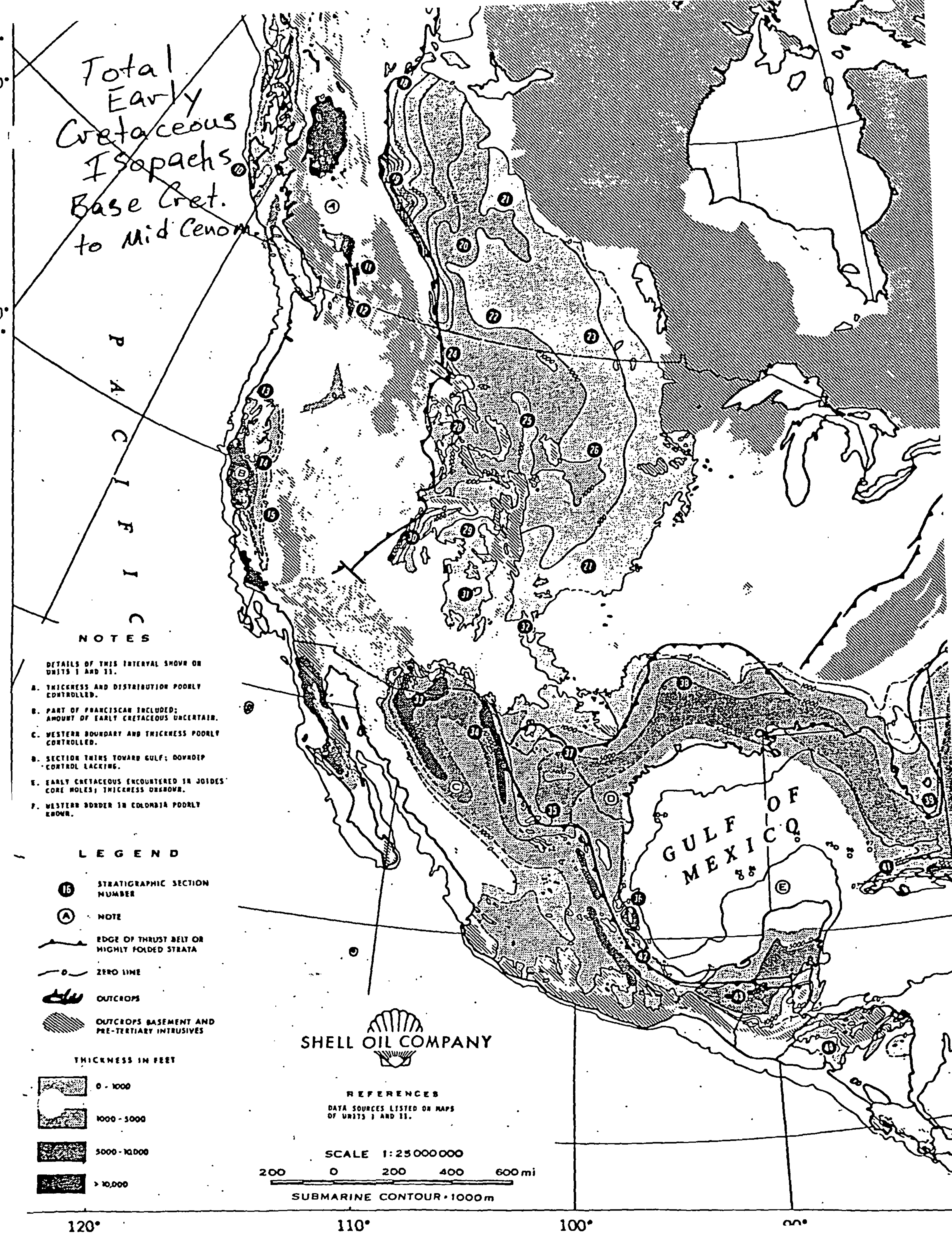
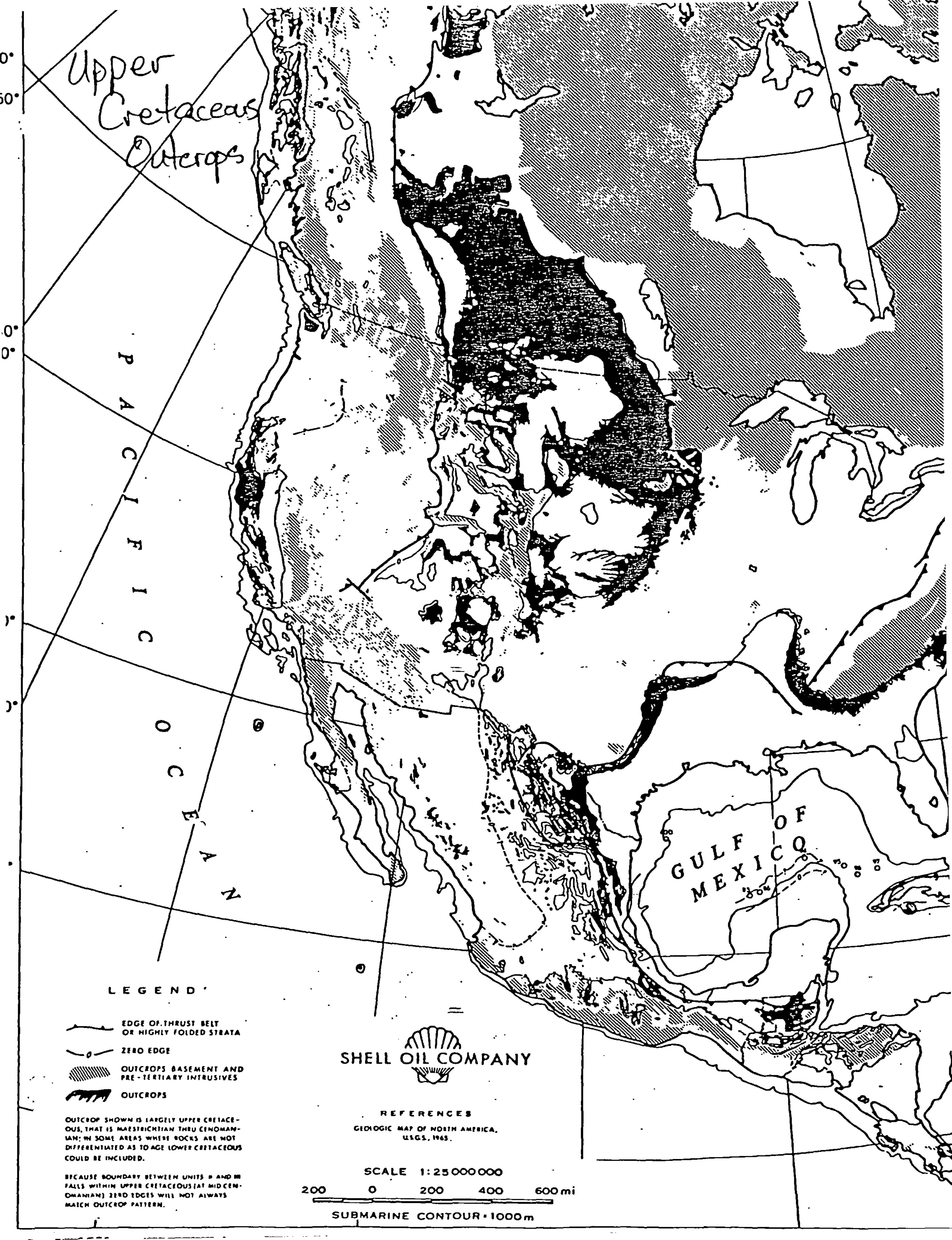


Total
Early
Cretaceous
Isopachs
Base Cret.
to Mid Cenom.
PACIFIC
GULF OF MEXICO
NOTES
DETAILS OF THIS INTERVAL SHOWN ON UNITS I AND II.
A. THICKNESS AND DISTRIBUTION POORLY CONTROLLED.
B. PART OF FRANCISCAN INCLUDED; AMOUNT OF EARLY CRETACEOUS UNCERTAIN.
C. WESTERN BOUNDARY AND THICKNESS POORLY CONTROLLED.
D. SECTION THINS TOWARD GULF; DOWNDIP CONTROL LACKING.
E. EARLY CRETACEOUS ENCOUNTERED IN JOIDES CORE HOLES; THICKNESS UNKNOWN.
F. WESTERN BORDER IN COLOMBIA POORLY KNOWN.
LEGEND
STRATIGRAPHIC SECTION NUMBER
NOTE
EDGE OF THRUST BELT OR HIGHLY FOLDED STRATA
ZERO LINE
OUTCROPS
OUTCROPS BASEMENT AND PRE-TERTIARY INTRUSIVES
THICKNESS IN FEET
0 - 1000
1000 - 5000
5000 - 10,000
> 10,000
SHELL OIL COMPANY
REFERENCES
DATA SOURCES LISTED ON MAPS OF UNITS I AND II.
SCALE 1:25 000 000
200 0 200 400 600 mi
SUBMARINE CONTOUR • 1000 m
120°
110°
100°


Upper
Cretaceas
Outcrops
PACIFIC OCEAN
GULF OF MEXICO
LEGEND
EDGE OF THRUST BELT OR HIGHLY FOLDED STRATA
ZERO EDGE
OUTCROPS BASEMENT AND PRE-TERTIARY INTRUSIVES
OUTCROPS
OUTCROP SHOWN IS LARGELY UPPER CRETACEOUS, THAT IS MAESTRICHTIAN THRU CENOMANIAN; IN SOME AREAS WHERE ROCKS ARE NOT DIFFERENTIATED AS TO AGE LOWER CRETACEOUS COULD BE INCLUDED.
BECAUSE BOUNDARY BETWEEN UNITS II AND III FALLS WITHIN UPPER CRETACEOUS (AT MID CENOMANIAN) ZERO EDGES WILL NOT ALWAYS MATCH OUTCROP PATTERN.
SHELL OIL COMPANY
REFERENCES
GEOLOGIC MAP OF NORTH AMERICA, U.S.G.S., 1965.
SCALE 1:25000000
200 0 200 400 600 mi
SUBMARINE CONTOUR = 1000 m

Cret. Unit III Isopachs
Mid Cenom. to Top Turonian
NOTES
A. OCCURRENCE SPOTTY, THICKNESS UNCERTAIN; MUCH OF INTERVAL MAY BE MISSING.
B. ALL U.K. INCLUDED IN KANOUK FM.; NO ATTEMPT TO DIVIDE BETWEEN UNITS III, IV, V.
C. KUSKOKWIM GEOSYNCLINE GENERALIZED FROM SPARSE DATA.
D. SHUMAGIN FM. EXCEEDS 10,000'; WIDELY DISTRIBUTED, POORLY DATED; NO ATTEMPT TO ALLOCATE TO MAP UNITS.
E. SUBDIVISION OF U.K. NOT POSSIBLE; MAXIMUM THICKNESS 5,000'; PARTS OF EAGLE PLAINS AND MOOSE CHANNEL FORMATIONS AND U.K. SHALE DIVISION INCLUDED; SOME BEDS NO DOUBT YOUNGER THAN THIS INTERVAL.
F. LITTLE BEAR FM. INCLUDED; AGE UNCERTAIN.
G. THIN VENEER OF CRETACEOUS MAY BE PARTLY THIS AGE.
H. SIFTON, SUTSUT, BEADY, AND SOPHIE MOUNTAIN FORMATIONS IN PART THIS AGE; NO ATTEMPT TO SUBDIVIDE.
I. UNIT MAY BE WIDELY PRESENT, MASKED BY YOUNG VOLCANICS. NEAR ROSEBURG, WHITSETT LST. DATED CENOMANEAN-TURONIAN BY FORAMINIFERS.
J. OUTCROP OF FRANCISCAN SHOWN IN BLACK-AMOUNT THIS AGE IS UNCERTAIN.
K. SOME 750' RED CONGLOMERATE IN SANTA MONICA MTNS. THOUGHT TO BE TURONIAN.
L. DISTRIBUTION UNCERTAIN UNDER VOLCANICS.
M. U.K. EXPOSURES SCATTERED; WESTERN BORDER QUESTIONABLE.
N. EXPOSURES RARE; MUCH VOLCANIC COVER; RUDIST-BEARING LST. GENERALLY LESS THAN 1000' THICK.
O. EAST OF 92° AND NORTH OF CENTRAL GUATEMALA THE U.K. IS NOT DIVISIBLE; MAP INCLUDES ALL ROCKS TURONIAN THRU MAESTRICHTIAN.
P. GAURE FM. INCLUDES BEDS OLDER THAN MID-CENOMANEAN.
Q. JOIDES CORE HOLES FAILED TO ENCOUNTER ROCKS THIS AGE.
R. "CRETACEOUS" ROCKS IN COSTA RICA AND PANAMA PROBABLY YOUNGER THAN TURONIAN; "BASEMENT ROCKS," INCLUDING METAMORPHICS AND IGNEOUS GENERALLY TERMED NICOYA COMPLEX, MAY INCLUDE ROCKS THIS AGE.
S. MAPPING IN SO. AMERICA INCLUDES ALL OF CENOMANIAN PLUS TURONIAN.
T. DABEIBA FM., POORLY DEFINED, INCLUDES 2800-3800 M. BLACK SHALE AND LST. INTRUDED BY ANDESITE DIKES; OLSSON (1956) DATES AS TURONIAN, VARIOUS OTHERS WOULD EXTEND RANGE BOTH OLDER AND YOUNGER.
U. DISTRIBUTION EXAGGERATED IN JAMAICA, HISPANOLA, PUERTO RICO, AND VIRGIN ISLANDS.
V. VOLCANIC SEQUENCE POORLY DATED.
W. PROBABLY THICKENS NORTHWESTWARD.
X. MAP INCLUDES BASAL SAND (DAKOTA) AND PART OF MANCOS SHALE.
Y. UNIT III NOT RECOGNIZED IN MINNESOTA, BUT COULD BE PRESENT IN NON-MARINE FACIES
LEGEND
CORE HOLE
STRATIGRAPHIC SECTION NUMBER
NOTE
EDGE OF THRUST BELT OR HIGHLY FOLDED STRATA
ZERO LINE
OUTCROP OF FRANCISCAN
OUTCROPS BASEMENT AND PRE-TERTIARY INTRUSIVES
THICKNESS IN FEET
0-1000
1000-10,000
> 10,000
REFERENCES
ASQUITH, 1970
BAILEY, ET AL., 1964
BOLIN & PETSCH, 1954
BONET, 1969
BOVIN, 1969
BROWN, 1968
BRYANT, ET AL., 1969
CADY, 1955
CLARK, 1915
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DONOVAN, 1961
GILL, ET AL., 1970
GRANTZ, 1960
HALE, 1961
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HUMPHREY & DIAZ, 1956
IMLAY, 1944
IMLAY & REESIDE, 1954
KOTTLOWSKI, 1963
MANER, 1965
MANER, 1963
MILLS, ET AL., 1967
MOUNTJOY, 1969
OLSON, 1956
PAGE, 1966
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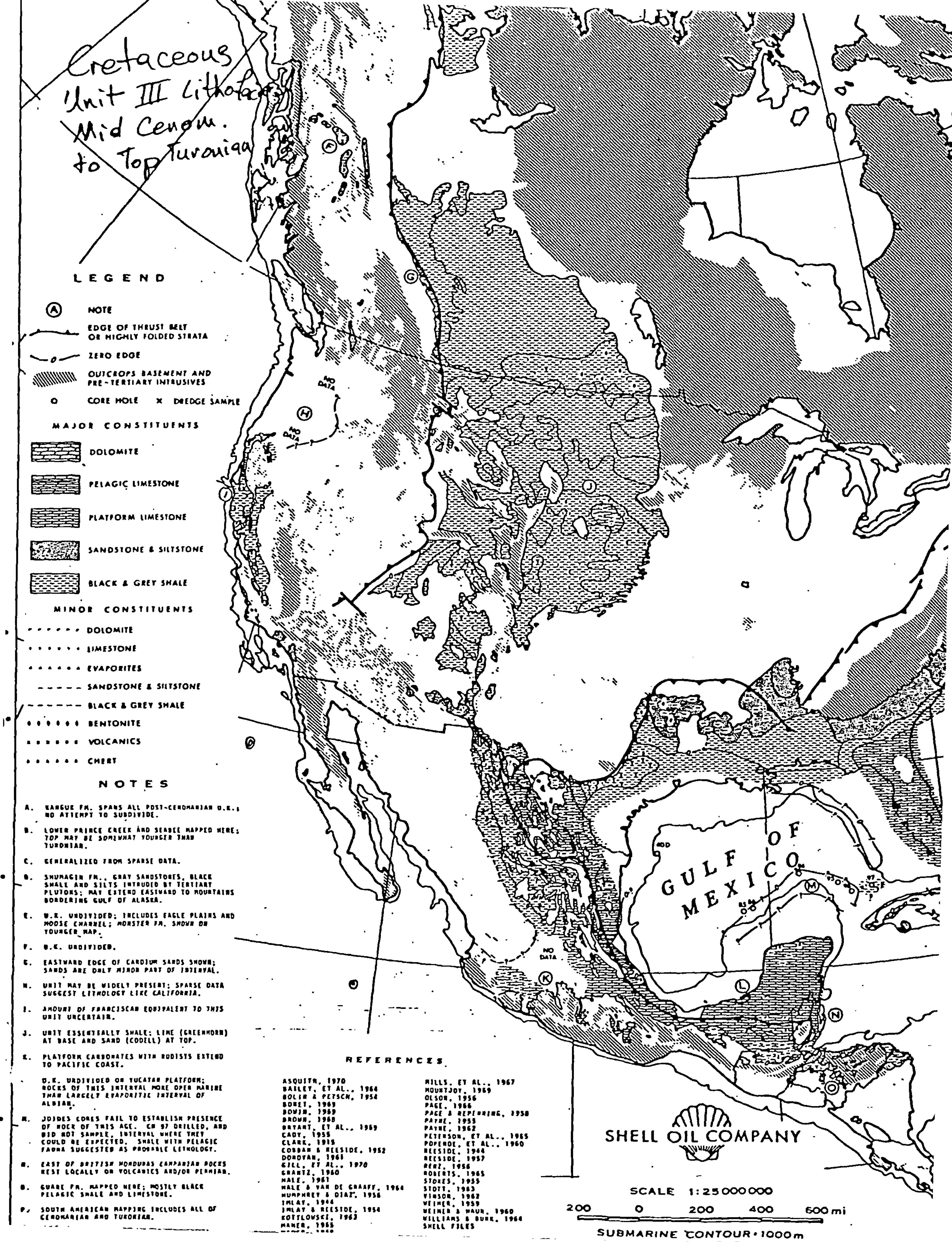
PAYNE, 1955
PAYNE, 1962
PETERSON, ET AL., 1965
POPENOE, ET AL., 1960
REESIDE, 1944
REESIDE, 1957
RENZ, 1956
ROBERTS, 1965
STOKES, 1955
STOTT, 1963
VINSON, 1962
WEIMER, 1959
WEIMER & HAUN, 1960
WILLIAMS & BURK, 1964
SHELL FILES
SCALE 1:25000000
200 0 200 400 600 mi
SUBMARINE CONTOUR · 1000 m
SHELL OIL COMPANY
GULF OF MEXICO
NO DATA
120°
110°
100°
90°



NOTES

A. KANGUK FM. SPANS ALL POST-CENOMANIAN U.K.; NO ATTEMPT TO SUBDIVIDE.

B. LOWER PRINCE CREEK AND SEABEE MAPPED HERE; *TOP MAY BE SOMEWHAT YOUNGER THAN TURONIAN.*

C. GENERALIZED FROM SPARSE DATA.

D. SHUMAGIN FM., GRAY SANDSTONES, BLACK SHALE AND SILTS INTRUDED BY TERTIARY PLUTONS; MAY EXTEND EASTWARD TO MOUNTAINS BORDERING GULF OF ALASKA.

E. U.K. UNDIVIDED; INCLUDES EAGLE PLAINS AND MOOSE CHANNEL; MONSTER FM. SHOWN ON YOUNGER MAP.

F. U.K. UNDIVIDED.

G. EASTWARD EDGE OF CARDIUM SANDS SHOWN; SANDS ARE ONLY MINOR PART OF INTERVAL.

H. UNIT MAY BE WIDELY PRESENT; SPARSE DATA SUGGEST LITHOLOGY LIKE CALIFORNIA.

I. AMOUNT OF FRANCISCAN EQUIVALENT TO THIS UNIT UNCERTAIN.

J. UNIT ESSENTIALLY SHALE; LIME (GREENHORN) AT BASE AND SAND (CODELL) AT TOP.

K. PLATFORM CARBONATES WITH RUDISTS EXTEND TO PACIFIC COAST.

U.K. UNDIVIDED ON YUCATAN PLATFORM; ROCKS OF THIS INTERVAL MORE OPEN MARINE THAN LARGELY EVAPORITIC INTERVAL OF ALBIAN.

M. JOIDES CORES FAIL TO ESTABLISH PRESENCE OF ROCK OF THIS AGE. CH 97 DRILLED, AND DID NOT SAMPLE, INTERVAL WHERE THEY COULD BE EXPECTED. SHALE WITH PELAGIC FAUNA SUGGESTED AS PROBABLE LITHOLOGY.

N. EAST OF BRITISH HONDURAS CAMPANIAN ROCKS REST LOCALLY ON VOLCANICS AND/OR PERMIAN.

O. GUARE FM. MAPPED HERE; MOSTLY BLACK PELAGIC SHALE AND LIMESTONE.

P. SOUTH AMERICAN MAPPING INCLUDES ALL OF CENOMANIAN AND TURONIAN.

REFERENCES

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BAILEY, ET AL., 1964
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BONET, 1969
BOWIN, 1969
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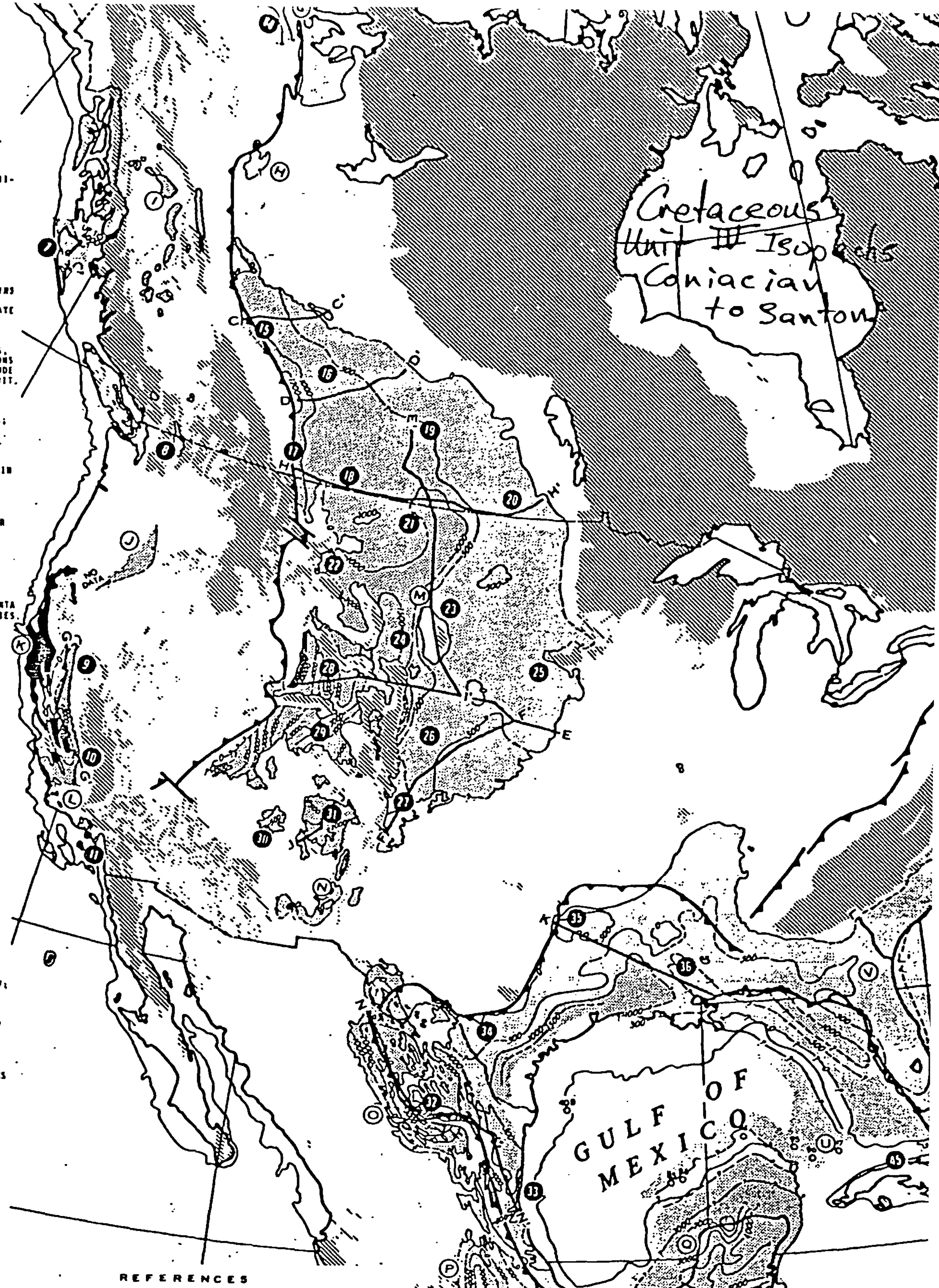
WEIMER, 1959
WEIMER & HAUN, 1960
WILLIAMS & BURK, 1964
SHELL FILES



NOTES

A. OCCURRENCE LIMITED; THICKNESS UNCERTAIN BUT LESS THAN 1000 FEET.

B. ALL U.K. INCLUDED IN KANGUK FM. (DATED TURONIAN-CAMPANIAN); NO ATTEMPT TO DIVIDE BETWEEN UNITS III-IV-V.

C. UPPER PORTION OF BEDS MAPPED (U. PRINCE CK.) PROBABLY REPRESENTS UNIT V.

D. GENERALIZED FROM SPARSE DATA.

E. SHUMAGIN FM. EXCEEDS 10,000 FEET, COVERS WIDE AREA ON SHELF, SANAK AND KODIAK ISLANDS, AND IN MOUNTAINS BORDERING GULF OF ALASKA (VALDEZ AND YAKUTAT); NO ATTEMPT TO ALLOCATE TO MAP UNITS.

F. SUBDIVISION OF U.K. NOT POSSIBLE; MAP INCLUDES PARTS OF EAGLE PLAINS, MOOSE CHANNEL AND MONSTER FORMATIONS AND U.K. SHALE DIVISION; MAY INCLUDE BEDS OLDER AND YOUNGER THAN MAP UNIT.

G. DISTRIBUTION UNCERTAIN.

H. KOOTANEELEE FM. OF FORT LIARD AREA; LESS THAN 1000'; DATED EARLY SANTONIAN.

I. SIFTON, SUTSUT, BEADY, AND SOPHIE MOUNTAIN FORMATIONS PROBABLY CONTAIN ROCKS THIS AGE; NO ATTEMPT TO SUBDIVIDE.

J. MAY BE WIDELY PRESENT IN CENTRAL OREGON AND WASHINGTON UNDER YOUNGER VOLCANICS.

K. FRANCISCAN OUTCROP SHOWN IN BLACK; AMOUNT AGE EQUIVALENT TO THIS UNIT IS UNCERTAIN.

L. ROCKS THIS AGE MAY BE MORE WIDELY PRESENT, ESPECIALLY IN VENTURA, SANTA BARBARA, AND SAN LUIS OBISPO COUNTIES.

M. LOWER MEMBER OF PIERRE (GAMMON FERRUGINEOUS MBR.) INCLUDED HERE.

N. MAP INCLUDES PART OF MANCOS SHALE, SANDY AND SHALEY UNITS REFERRED TO MESAVERDE AND A SANDY UNIT WITH VOLCANICS (McRAE).

O. U.K. SPARSELY EXPOSED; WESTERN BORDER IS QUESTIONABLE.

P. THICK CLASTIC SECTION (MEXCALA) GRADES EASTWARD TO LIMESTONE (ESCAMELA AND TILANTONGO).

Q. EAST OF 92nd MERIDIAN AND NORTH OF CENTRAL GUATEMALA THE U.K. IS NOT DIVISIBLE WITH PRESENT CONTROL; MAP INCLUDES ALL ROCKS FROM TURONIAN THRU MAESTRICHTIAN.

R. MAPPING IN HONDURAS INCLUDES ALL POST-TURONIAN U.K.

S. "BASEMENT ROCKS" INCLUDING METAMORPHICS AND IGNEOUS GENERALLY CALLED "NICOYA COMPLEX" MAY INCLUDE ROCKS THIS AGE.

T. DABEIBA FM. MAY BE PARTLY THIS AGE; TO THE NORTH CARMEN UPLIFT HAS BEDS WITH CONIACIAN FAUNA.

U. JOIDES CORE HOLE 95 ENCOUNTERED DEEP WATER SEDIMENT FROM THIS UNIT; INTERVAL DRILLED BUT UNSAMPLED IN 97; ABSENT IN CORE HOLE 94 AND PROBABLY ABSENT IN 86.

V. SUWANEE CHANNEL, A DEEP-WATER SEAWAY ACROSS NORTHERN FLORIDA IS DATED BY FORAMINIFERS AS EARLY SANTONIAN.

W. OPHIOLITE BELT OCCURS NORTH OF CORDILLERA CENTRAL; MAY INCLUDE ROCKS THIS AGE.

X. ROCKS THIS AGE MISSING OR VERY THIN IN JOIDES CORE HOLES SHOWN; ALSO IN CORE TAKEN ON BERMUDA RISE.

LEGEND

o CORE HOLE

16 STRATIGRAPHIC SECTION NUMBER

(A) NOTE

EDGE OF THRUST BELT OR HIGHLY FOLDED STRATA

ZERO LINE

OUTCROP OF FRANCISCAN

OUTCROPS BASEMENT AND PRE-TERTIARY INTRUSIVES

THICKNESS IN FEET

0-1000

1000-10,000

> 10,000

REFERENCES

ASQUITH, 1970
BAILEY, ET AL., 1964
BOLIN & PETSCH, 1954
BONET, 1969
BOWIN, 1969
BROWN, 1968
BRYANT, ET AL., 1969
CADY, 1955
CLARK, 1915
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HALE, 1961
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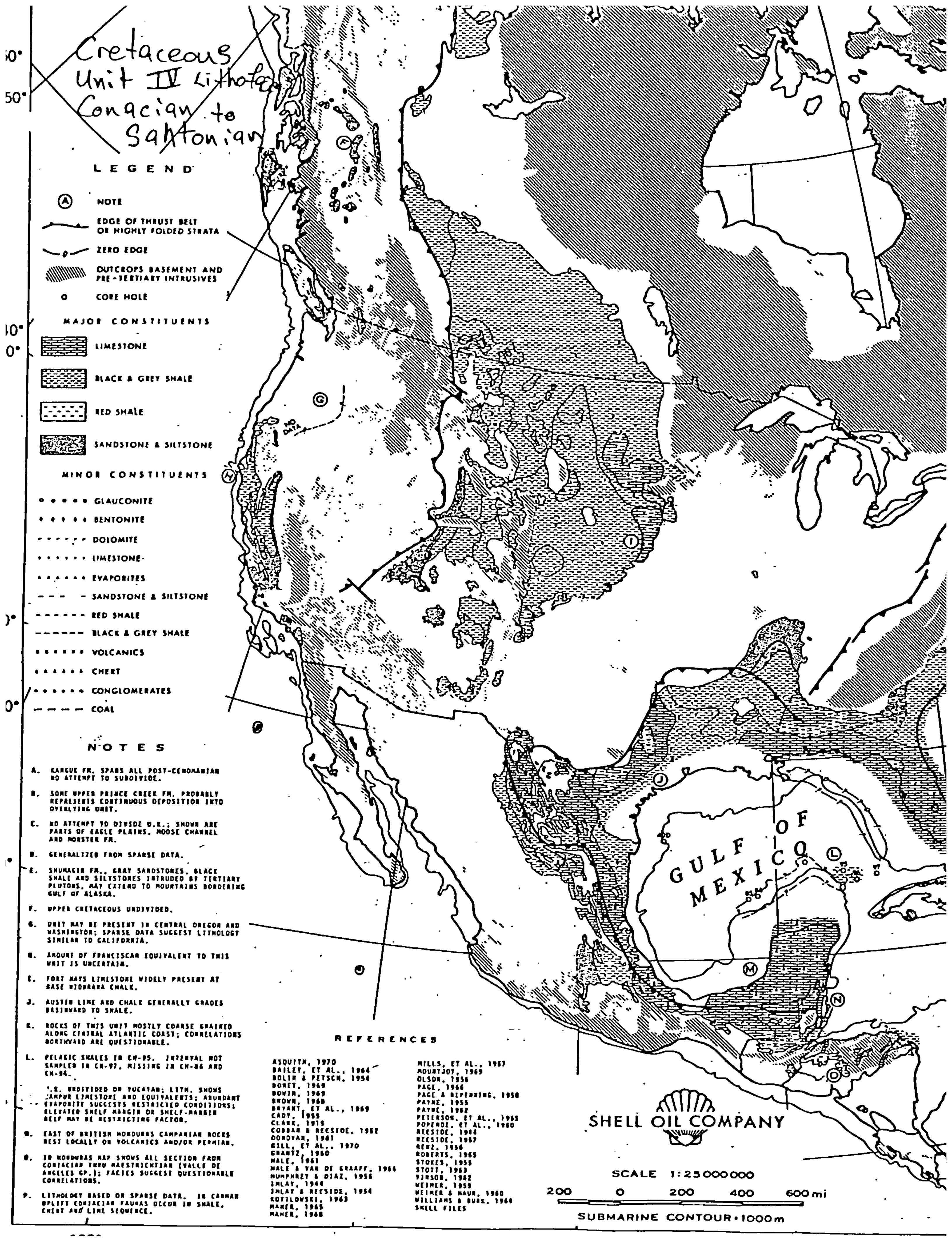
HUMPHREY & DIAZ, 1956
IMLAY, 1944
IMLAY & REESIDE, 1954
KOTTLOWSKI, 1963
MAHER, 1965
MAHER, 1968
MILLS, ET AL., 1967
MOUNTJOY, 1969
OLSON, 1956
PAGE, 1966
PAGE & REPENNING, 1958
PAYNE, 1955
PAYNE, 1962
PETERSON, ET AL., 1965
POPENOE, ET AL., 1960
REESIDE, 1944
REESIDE, 1957
RENZ, 1956
ROBERTS, 1965
STOKES, 1955
STOTT, 1963
VINSON, 1962
WEIMER, 1959
WEIMER & HAUN, 1960
WILLIAMS & BURK, 1964
SHELL FILES

SCALE 1:25 000 000

200 0 200 400 600 mi

SUBMARINE CONTOUR • 1000 m

SHELL OIL COMPANY



Cretaceous
Unit IV Lithofacies
Conacian to
Santonian
LEGEND
NOTE
EDGE OF THRUST BELT OR HIGHLY FOLDED STRATA
ZERO EDGE
OUTCROPS BASEMENT AND PRE-TERTIARY INTRUSIVES
CORE HOLE
MAJOR CONSTITUENTS
LIMESTONE
BLACK & GREY SHALE
RED SHALE
SANDSTONE & SILTSTONE
MINOR CONSTITUENTS
GLAUCONITE
BENTONITE
DOLOMITE
LIMESTONE
EVAPORITES
SANDSTONE & SILTSTONE
RED SHALE
BLACK & GREY SHALE
VOLCANICS
CHERT
CONGLOMERATES
COAL
NOTES
A. KANGUK FM. SPANS ALL POST-CENOMANIAN NO ATTEMPT TO SUBDIVIDE.
B. SOME UPPER PRINCE CREEK FM. PROBABLY REPRESENTS CONTINUOUS DEPOSITION INTO OVERLYING UNIT.
C. NO ATTEMPT TO DIVIDE U.K.; SHOWN ARE PARTS OF EAGLE PLAINS, MOOSE CHANNEL AND MONSTER FM.
D. GENERALIZED FROM SPARSE DATA.
E. SHUMAGIN FM., GRAY SANDSTONES, BLACK SHALE AND SILTSTONES INTRUDED BY TERTIARY PLUTONS, MAY EXTEND TO MOUNTAINS BORDERING GULF OF ALASKA.
F. UPPER CRETACEOUS UNDIVIDED.
G. UNIT MAY BE PRESENT IN CENTRAL OREGON AND WASHINGTON; SPARSE DATA SUGGEST LITHOLOGY SIMILAR TO CALIFORNIA.
H. AMOUNT OF FRANCISCAN EQUIVALENT TO THIS UNIT IS UNCERTAIN.
I. FORT HAYS LIMESTONE WIDELY PRESENT AT BASE NIOBRARA CHALK.
J. AUSTIN LIME AND CHALK GENERALLY GRADES BASINWARD TO SHALE.
K. ROCKS OF THIS UNIT MOSTLY COARSE GRAINED ALONG CENTRAL ATLANTIC COAST; CORRELATIONS NORTHWARD ARE QUESTIONABLE.
L. PELAGIC SHALES IN CH-95. INTERVAL NOT SAMPLED IN CH-97, MISSING IN CH-86 AND CH-94.
U.K. UNDIVIDED ON YUCATAN; LITH. SHOWS CAMPUR LIMESTONE AND EQUIVALENTS; ABUNDANT EVAPORITE SUGGESTS RESTRICTED CONDITIONS; ELEVATED SHELF MARGIN OR SHELF-MARGIN REEF MAY BE RESTRICTING FACTOR.
N. EAST OF BRITISH HONDURAS CAMPANIAN ROCKS REST LOCALLY ON VOLCANICS AND/OR PERMIAN.
O. IN HONDURAS MAP SHOWS ALL SECTION FROM CONIACIAN THRU MAESTRICHTIAN (VALLE DE ANGELES GP.); FACIES SUGGEST QUESTIONABLE CORRELATIONS.
P. LITHOLOGY BASED ON SPARSE DATA. IN CAYMAN UPLIFT CONIACIAN FAUNAS OCCUR IN SHALE, CHERT AND LIME SEQUENCE.
NO DATA
GULF OF MEXICO
REFERENCES
ASQUITH, 1970
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STOTT, 1963
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WEIMER, 1959
WEIMER & HAUN, 1960
WILLIAMS & BURK, 1964
SHELL FILES
SHELL OIL COMPANY
SCALE 1:25000000
200 0 200 400 600 mi
SUBMARINE CONTOUR • 1000m

NOTES

A. OCCURRENCE LIMITED; THICKNESS UNCERTAIN BUT NOWHERE EXCEEDS 1000'.

B. RANGE INCLUDES TURONIAN-CAMPANIAN; NO ATTEMPT TO DIVIDE BETWEEN UNITS III-IV-V.

C. MAY BE REPRESENTED BY UPPER PORTION OF BEDS MAPPED AS UNIT IV; DEPOSITION PROBABLY CONTINUOUS INTO TERTIARY BUT THIS UNIT NOT SEPARABLE.

D. SHUMAGIN FM., KODIAK AND VALDEZ WIDELY PRESENT ON SHELF.

E. VALDEZ FM. INCLUDES BEDS THIS AGE; UP TO 10,000 FT.

F. SUBDIVISION OF UPPER CRETACEOUS NOT POSSIBLE; MAXIMUM THICKNESS ABOUT 5000 FEET; INCLUDED ARE EAGLE PLAINS, MOOSE CHANNEL, AND MONSTER FORMATIONS. PRESENCE OF BEDS THIS AGE NEAR NORMAN WELLS IS QUESTIONABLE.

G. THIS INTERVAL (WAPITI) MAPPED AT FORT LIARD.

H. SEFTON, SUTSUT, BEADY, AND SOPHIE MOUNTAIN PROBABLY CONTAIN ROCKS THIS AGE; NO ATTEMPT TO SUBDIVIDE.

I. NON-MARINE BEDS (CHUCKANUT) MAY BE CRETACEOUS. MARINE ROCKS THIS AGE (LAMBERT AND NORTHUMBERLAND [PART]) OCCUR ON SAN JUAN AND VANCOUVER ISLANDS.

J. LOWER PART OF SOLEDUCK FM. MAY BE CRETACEOUS.

K. LOWER PART OF PIPESTONE CANYON FM. MAY BE CRETACEOUS.

L. OUTCROP FRANCISCAN SHOWN IN BLACK; AMOUNT THIS AGE UNCERTAIN. CAMPANIAN FOSSILS FOUND NORTH OF SAN FRANCISCO.

M. VALLE SALITRAL AND ROSARIO FORMATIONS LOCALLY MAY EXCEED 10,000 FEET.

N. MAPPING INCLUDES LANCE AND EQUIVALENT, EXCLUDES FORT UNION AND EQUIVALENT. EASTWARD THE UPPERMOST CRETACEOUS HAS BEEN REMOVED OVER WIDE AREAS.

O. WESTERN BOUNDARY POORLY DEFINED.

P. BASED ON SPARSE DATA.

Q. EAST OF 92° AND NORTH OF CENTRAL GUATEMALA UPPER CRETACEOUS IS NOT DIVISIBLE; MAP INCLUDES ALL ROCKS TURONIAN THRU MAESTRICHTIAN.

R. MAPPING IN HONDURAS INCLUDES ALL POST-TURONIAN CRETACEOUS ROCKS.

S. ROCKS OF QUESTIONABLE UPPER CRETACEOUS AGE (NICOYA, IN PART) UNDERLIE EOCENE AND PALEOCENE OVER MOST OF PANAMA AND SOUTHERN COSTA RICA.

T. THICKNESS UNCERTAIN.

U. OPHIOLITE BELT OCCURS NORTH OF CORDILLERA CENTRAL; MAY INCLUDE ROCKS OF THIS AGE.

V. THICKNESS IN VOLCANIC SEQUENCE UNCERTAIN.

W. THICKNESS IN JOIDES CORE ESTIMATED AT 200 m. MAXIMUM.

X. ROCKS OF THIS AGE MISSING IN CORE HOLES SHOWN; ALSO MISSING ON BERMUDA RISE.

Y. PRESENT IN CH 95; POSSIBLY PRESENT IN CH 97; UNSAMPLED OR ABSENT IN CH 86 AND CH 94.

LEGEND

CORE HOLE

STRATIGRAPHIC SECTION NUMBER

NOTE

EDGE OF THRUST BELT OR HIGHLY FOLDED STRATA

ZERO LINE

OUTCROP OF FRANCISCAN

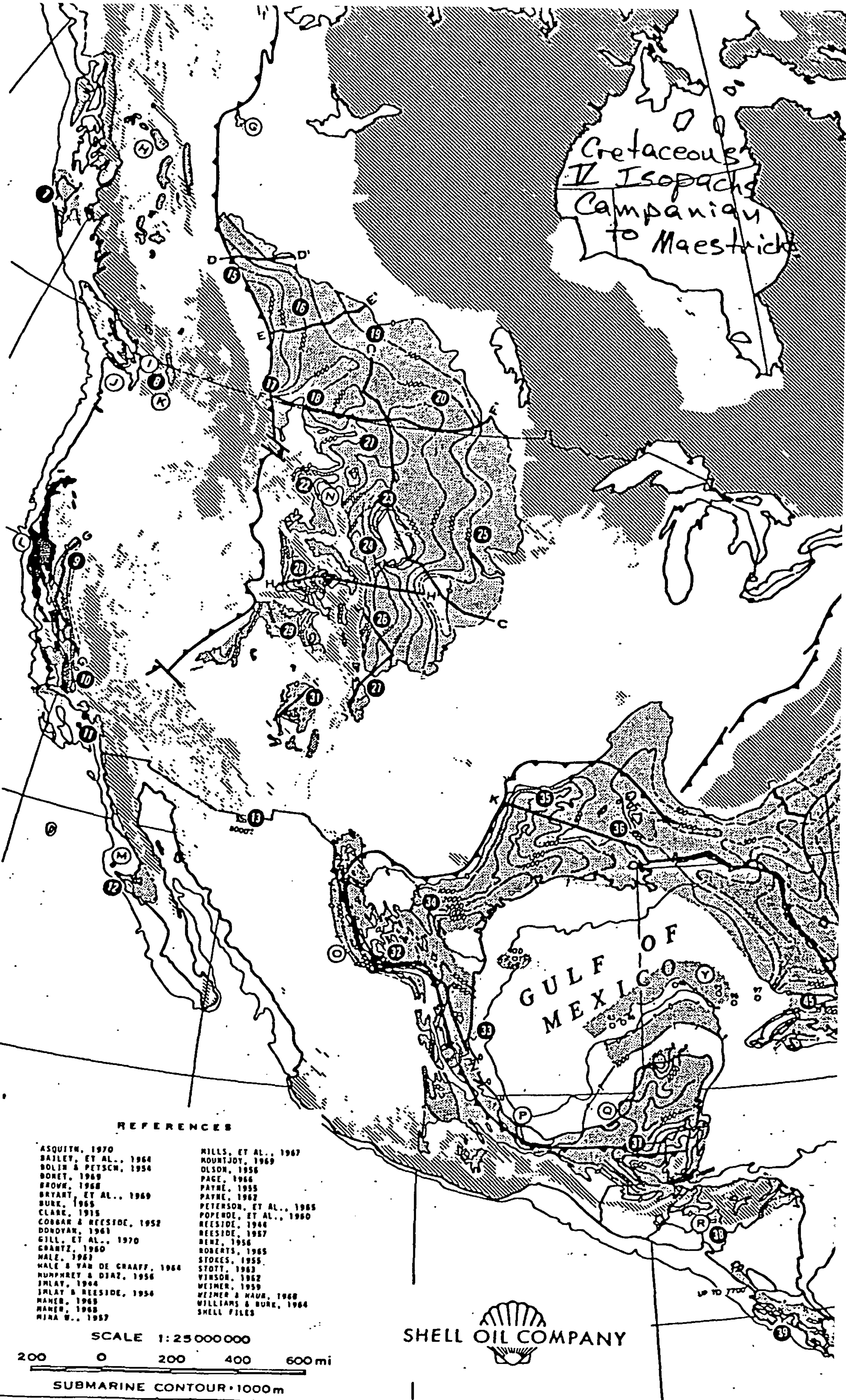
OUTCROPS BASEMENT AND PRE-TERTIARY INTRUSIVES

THICKNESS IN FEET

0-1000

1000-10,000

> 10,000



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HALE, 1961
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IMLAY, 1944
IMLAY & REESIDE, 1954
MAHER, 1963
MAHER, 1968
MINA U., 1957
MILLS, ET AL., 1967
MOUNTJOY, 1969
OLSON, 1956
PAGE, 1966
PAYNE, 1955
PAYNE, 1962
PETERSON, ET AL., 1965
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WILLIAMS & BURK, 1964
SHELL FILES

NOTES

...NGUE INCLUDES TURONIAN-...AMPANIAN; NO ATTEMPT TO DIVIDE ...ETWEEN UNITS III-IV-V.

B. SUBDIVISION OF UPPER CRETACEOUS NOT POSSIBLE; INCLUDED ARE EAGLE PLAINS, MOOSE CHANNEL, AND MONSTER FORMATIONS; PRESENCE OF BEDS THIS AGE NEAR NORMAN WELLS IS QUESTIONABLE.

C. MUCH OF COAL IN WESTERN CANADA OCCURS IN BELLY RIVER AND EDMONTON STRATA.

D. LITHOLOGY GENERALIZED; FRANCISCAN WEST AND GREAT VALLEY EAST.

E. IRON SPRINGS FM. AND PINTO SANDSTONE INCLUDED HERE.

F. MAPPING SHOWS ALL ROCKS FROM TURONIAN THRU MAESTRICHTIAN IN NORTHERN GUATEMALA AND YUCATAN. LOCALLY IN GUATEMALA AND BRITISH HONDURAS THIS UNIT IS MISSING.

G. *IN HONDURAS MAP SHOWS* ALL SECTION FROM CONIACIAN THRU MAESTRICHTIAN (VALLE DE ANGELES GROUP). CORRELATIONS ARE QUESTIONABLE.

H. LITHOLOGY NOT WELL KNOWN; FRANCISCAN-LIKE ROCKS ARE INCLUDED IN NICOYA; DISTRIBUTION UNCERTAIN.

I. LURUACO FM., ESTIMATED TO BE 8000'± THICK AND IN PART UPPER CRETACEOUS IS LARGELY SHALE WITH SILTS, SANDS AND CONGLOMERATES; *OCCURS WIDELY ALONG NORTHWESTERN* COAST OF COLOMBIA.

J. CORE HOLE 95 (JOIDES) CONTAINS FINE-GRAINED SEDIMENTS THIS AGE WITH PELAGIC FAUNAS; CORE HOLE 97 WAS NOT SAMPLED WHERE THIS INTERVAL MAY OCCUR; INTERVAL IS DEEP-WATER AT JORDON KNOLL AND MISSING *IN CH 86 AND 94.*

K. CARBONIFEROUS ROCK RECOVERED FROM SALT DOME.

L. DEEP WATER LIMESTONES FLANK RIDGE WHERE PALEOCENE RESTS ON SANTONIAN.

M. *LITHOLOGY SHOWN* SAME AS UNIT IV; SEDIMENTATION PROBABLY CONTINUOUS INTO TERTIARY BUT UPPERMOST CRETACEOUS IS NOT DIVISIBLE.

LEGEND

Ⓐ NOTE

EDGE OF THRUST BELT OR HIGHLY FOLDED STRATA

ZERO EDGE

OUTCROPS BASEMENT AND PRE-TERTIARY INTRUSIVES

CORE HOLE

MAJOR CONSTITUENTS

DOLOMITE

PELAGIC LIMESTONE

PLATFORM LIMESTONE

SANDSTONE & SILTSTONE

RED SHALE

BLACK & GREY SHALE

MINOR CONSTITUENTS

BENTONITE

CONGLOMERATES

VOLCANICS

DOLOMITE

LIMESTONE

EVAPORITES

SANDSTONE & SILTSTONE

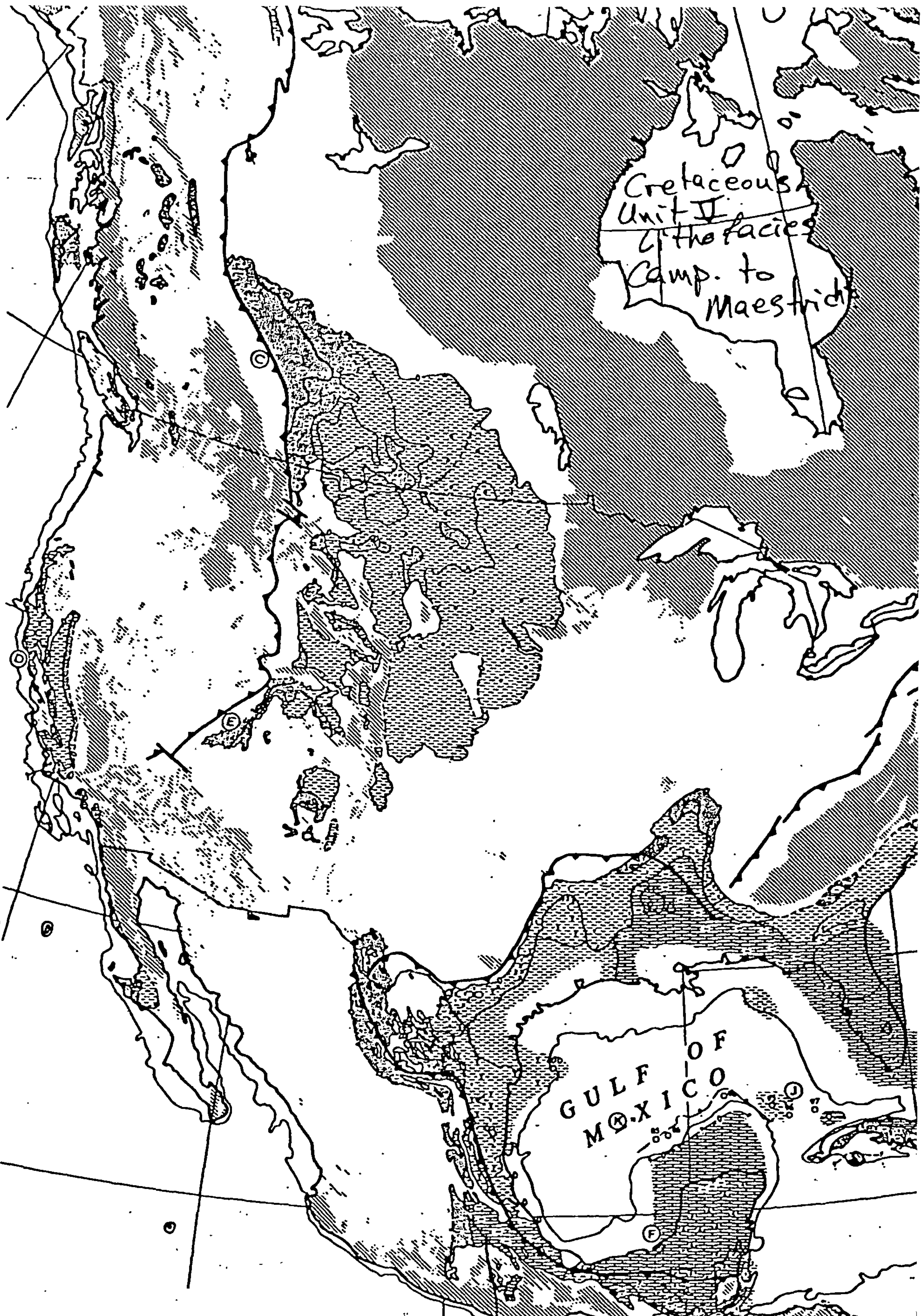
RED SHALE

GREEN SHALE

BLACK & GREY SHALE

CHERT

COAL



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HALE & VAN DE GRAAFF, 1964
HUMPHREY & DIAZ, 1956
IMLAY, 1944
IMLAY & REESIDE, 1954
MAHER, 1965
MAHER, 1968
MINA U., 1957
MILLS, ET AL., 1967
MOUNTJOY, 1969
OLSON, 1956
PAGE, 1966
PAYNE, 1955
PAYNE, 1962
PETERSON, ET AL., 1965
POPENOE, ET AL., 1960
REESIDE, 1944
REESIDE, 1957
RENZ, 1956
ROBERTS, 1965
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VINSON, 1962
WEIMER, 1959
WEIMER & HAUN, 1960
WILLIAMS & BURK, 1964
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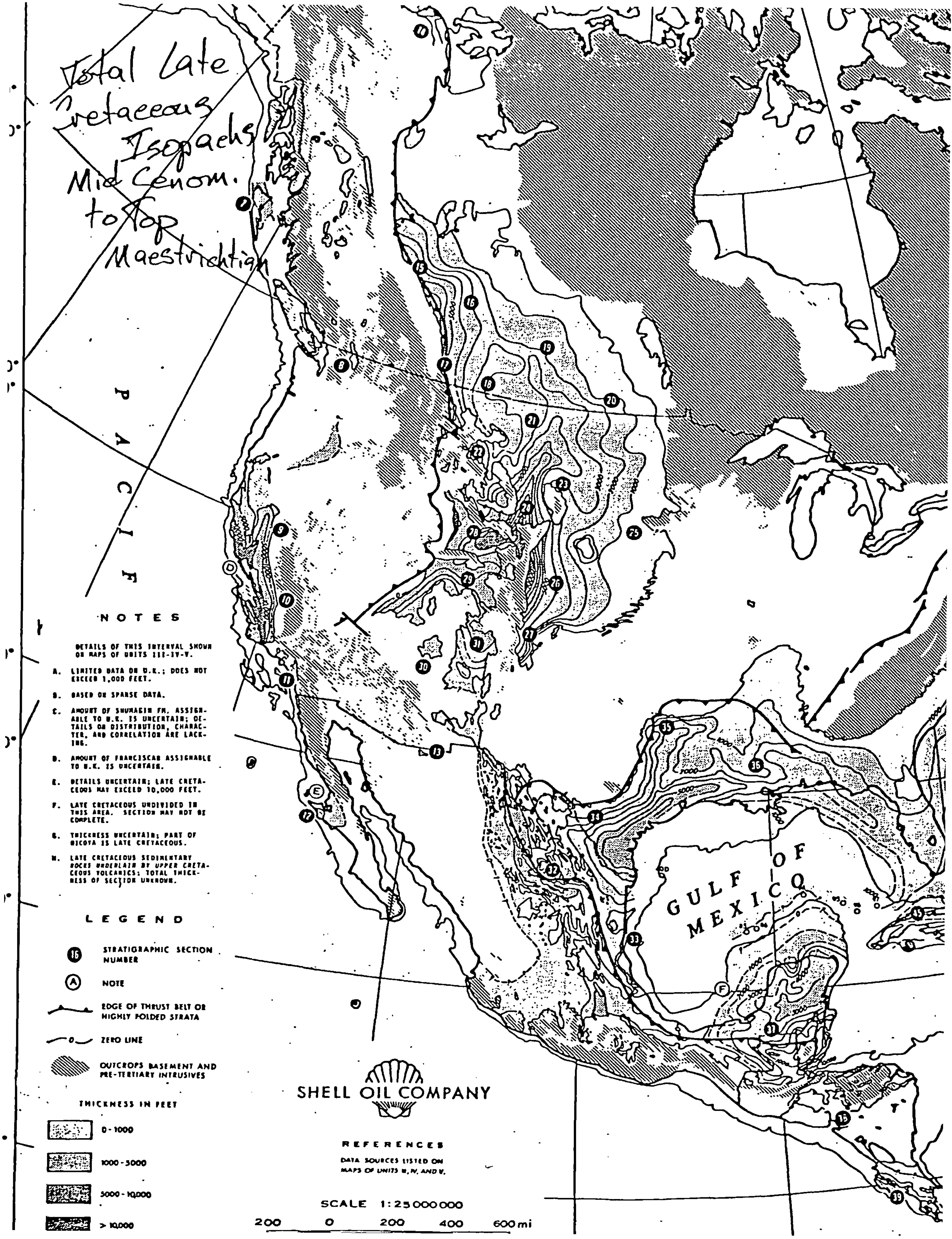
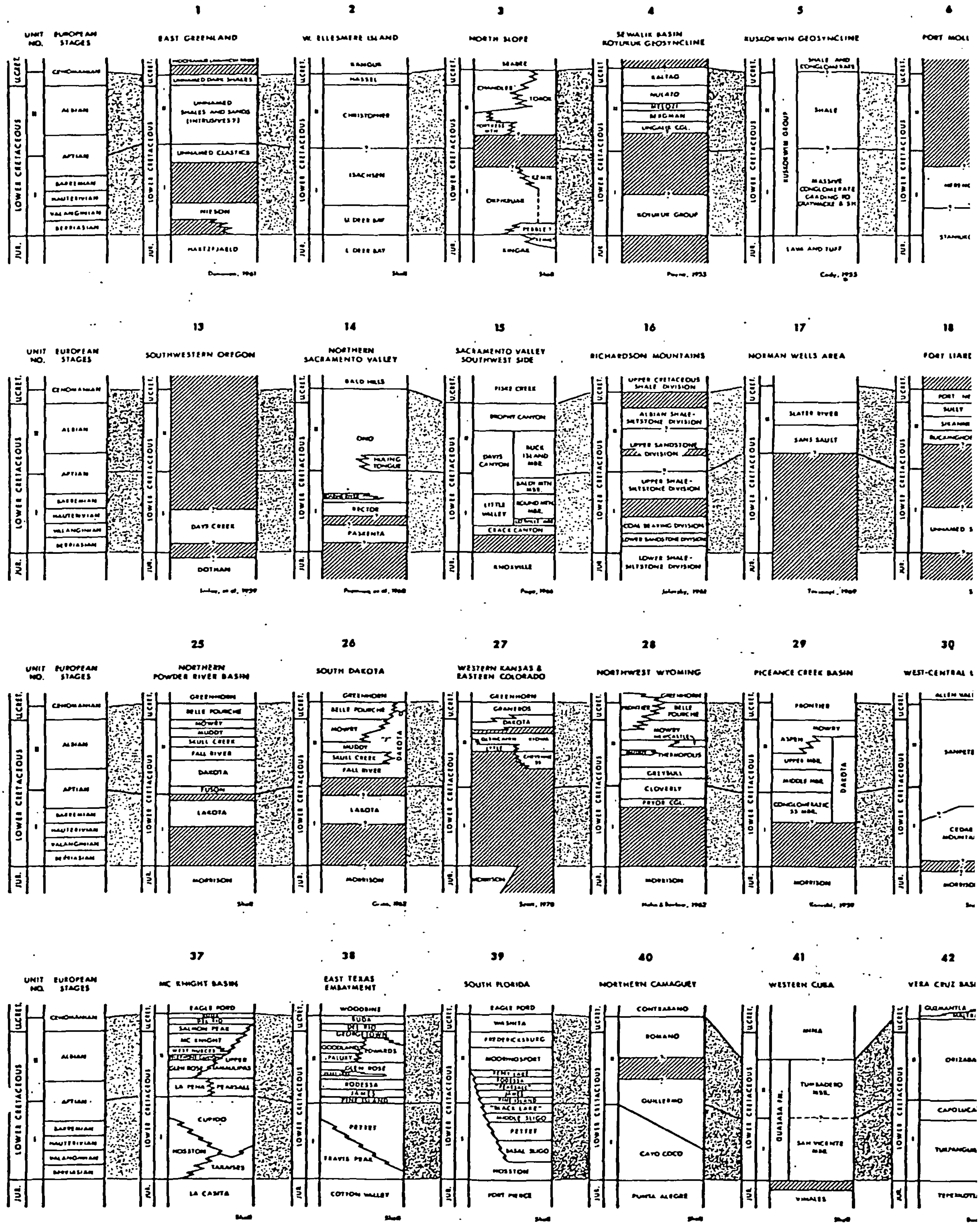


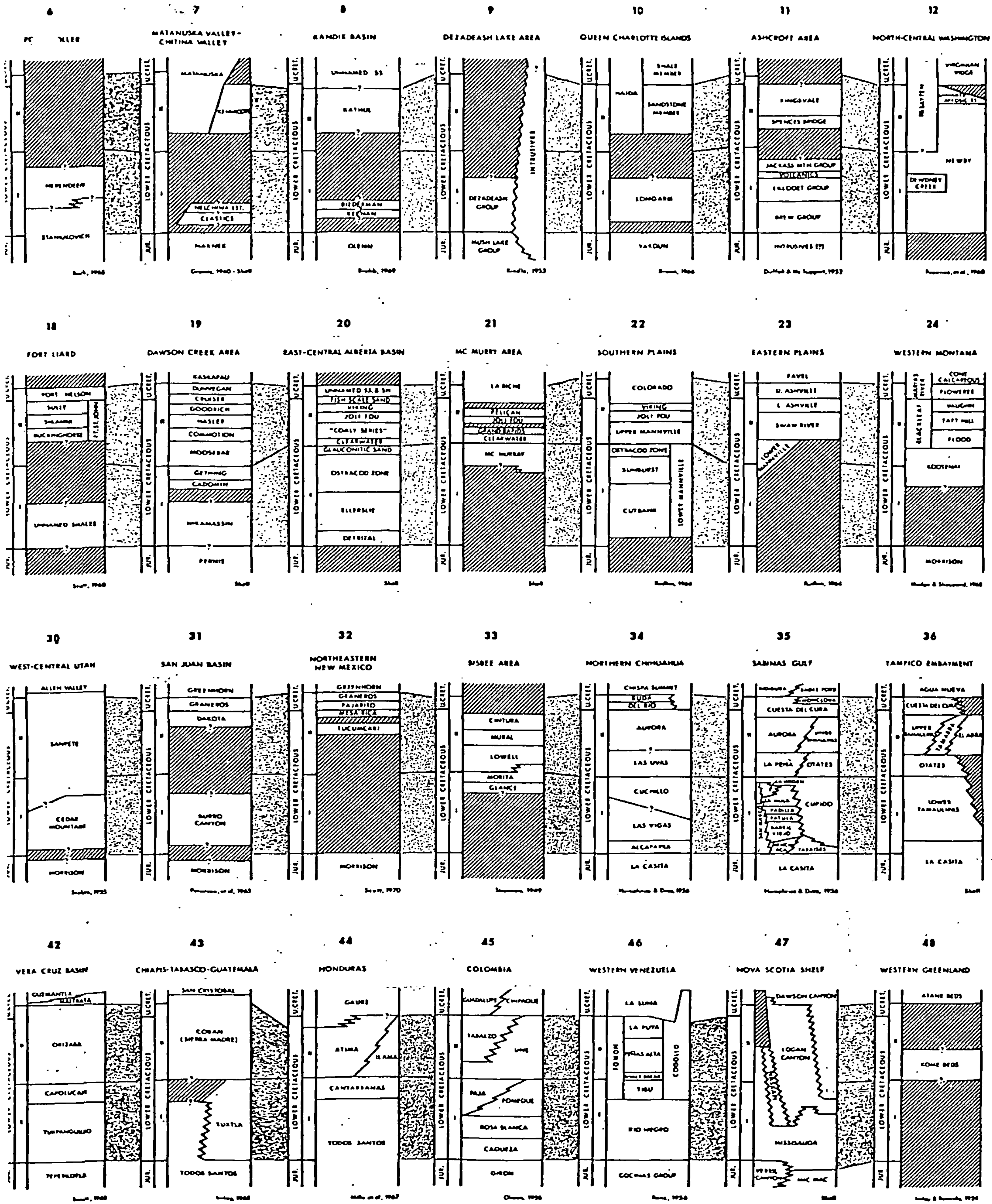
Total Late Cretaceous Isopachs Mid Cenom. to Top Maestrichtian
PACIFIC
GULF OF MEXICO
NOTES
DETAILS OF THIS INTERVAL SHOWN ON MAPS OF UNITS III-IV-V.
A. LIMITED DATA ON U.K.; DOES NOT EXCEED 1,000 FEET.
B. BASED ON SPARSE DATA.
C. AMOUNT OF SHUMAGIN FM. ASSIGNABLE TO U.K. IS UNCERTAIN; DETAILS ON DISTRIBUTION, CHARACTER, AND CORRELATION ARE LACKING.
D. AMOUNT OF FRANCISCAN ASSIGNABLE TO U.K. IS UNCERTAIN.
E. DETAILS UNCERTAIN; LATE CRETACEOUS MAY EXCEED 10,000 FEET.
F. LATE CRETACEOUS UNDIVIDED IN THIS AREA. SECTION MAY NOT BE COMPLETE.
G. THICKNESS UNCERTAIN; PART OF NICOYA IS LATE CRETACEOUS.
H. LATE CRETACEOUS SEDIMENTARY ROCKS UNDERLAIN BY UPPER CRETACEOUS VOLCANICS; TOTAL THICKNESS OF SECTION UNKNOWN.
LEGEND
STRATIGRAPHIC SECTION NUMBER
NOTE
EDGE OF THRUST BELT OR HIGHLY FOLDED STRATA
ZERO LINE
OUTCROPS BASEMENT AND PRE-TERTIARY INTRUSIVES
THICKNESS IN FEET
0 - 1000
1000 - 5000
5000 - 10,000
> 10,000
SHELL OIL COMPANY
REFERENCES
DATA SOURCES LISTED ON MAPS OF UNITS III, IV, AND V.
SCALE 1:25000000
200 0 200 400 600 mi

(BASE CRETACEOUS TO MI



STRATIGRAPHIC SECTIONS

S TO MID-CENOMANIAN)



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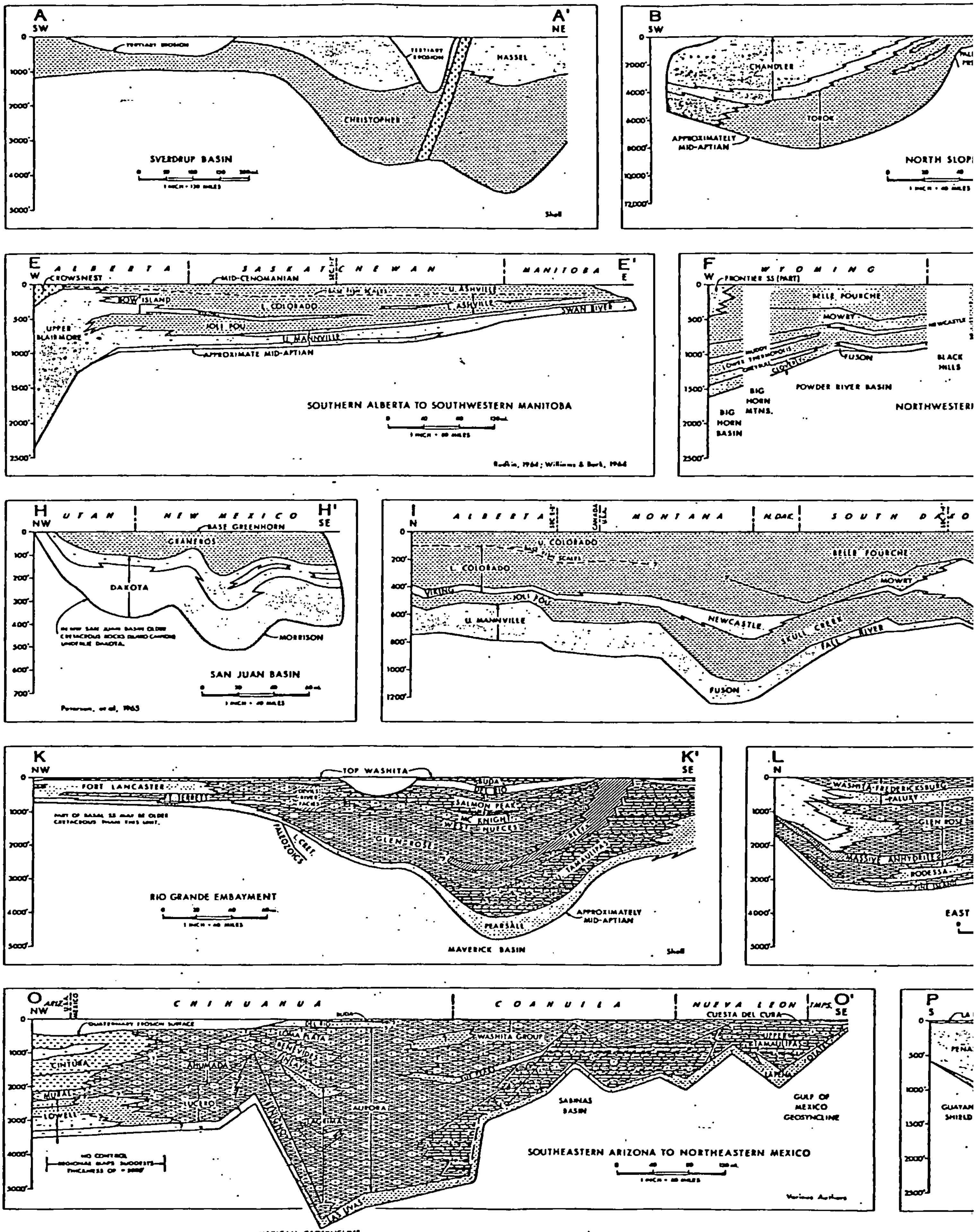
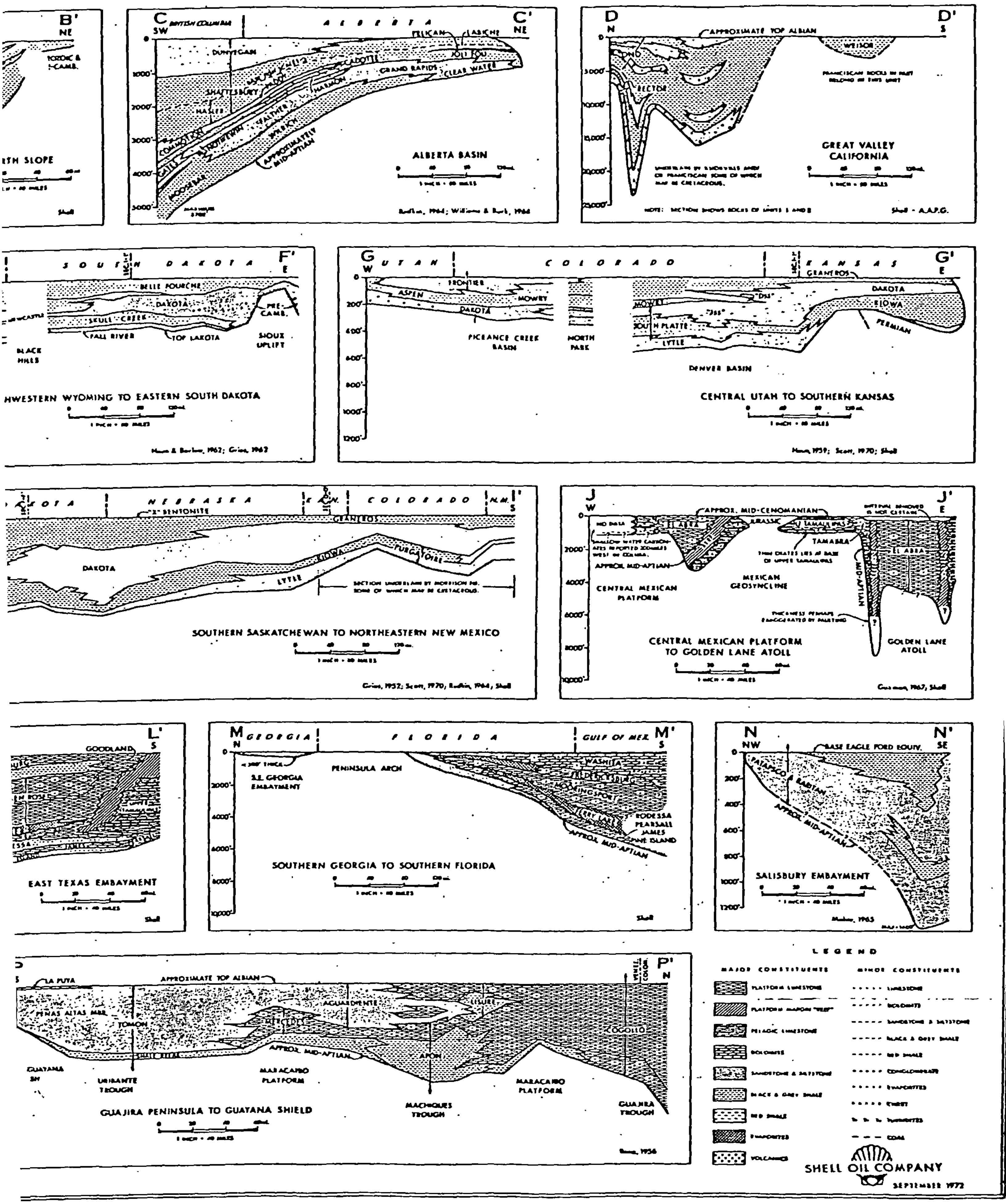
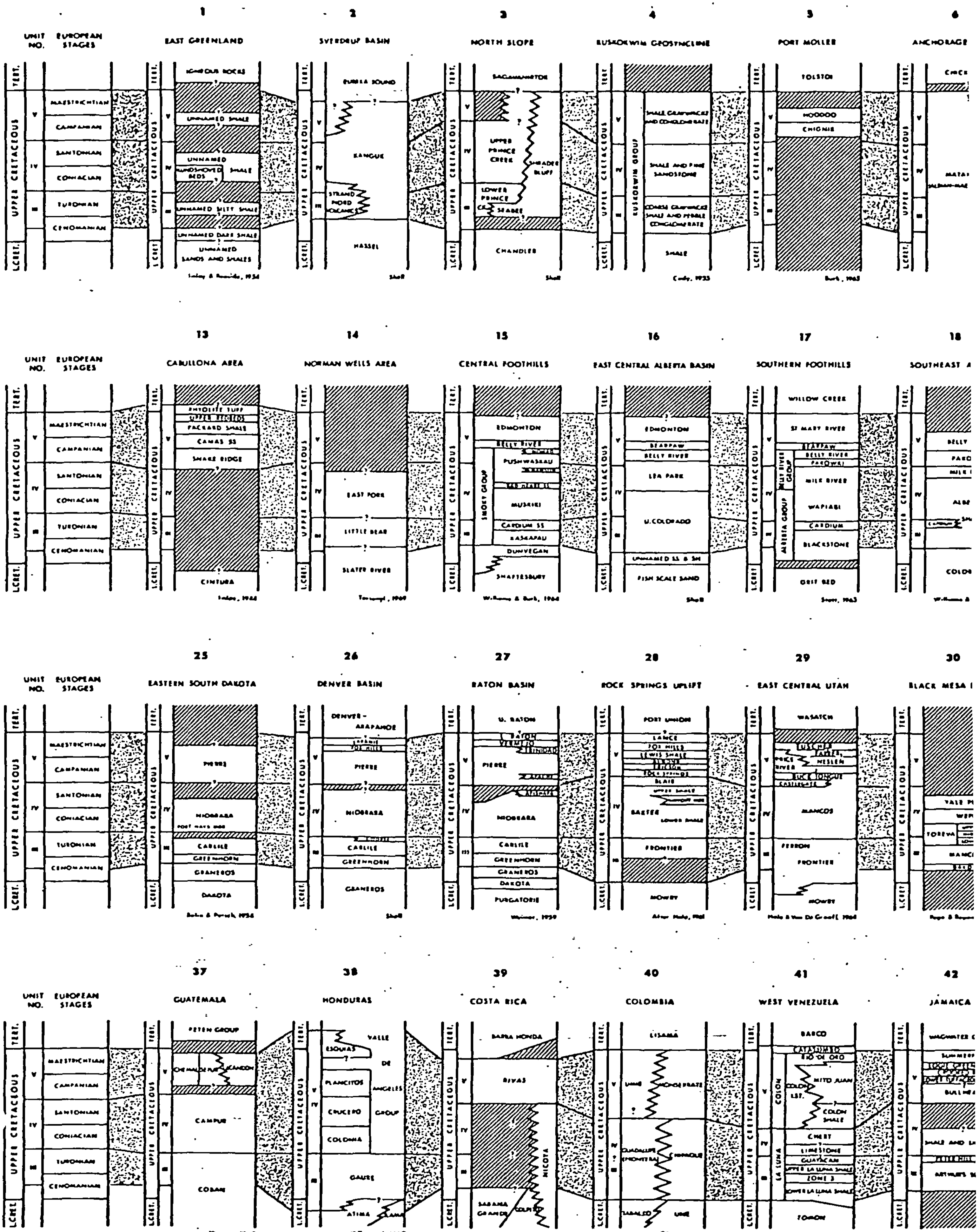
(MID APTIAN - MID



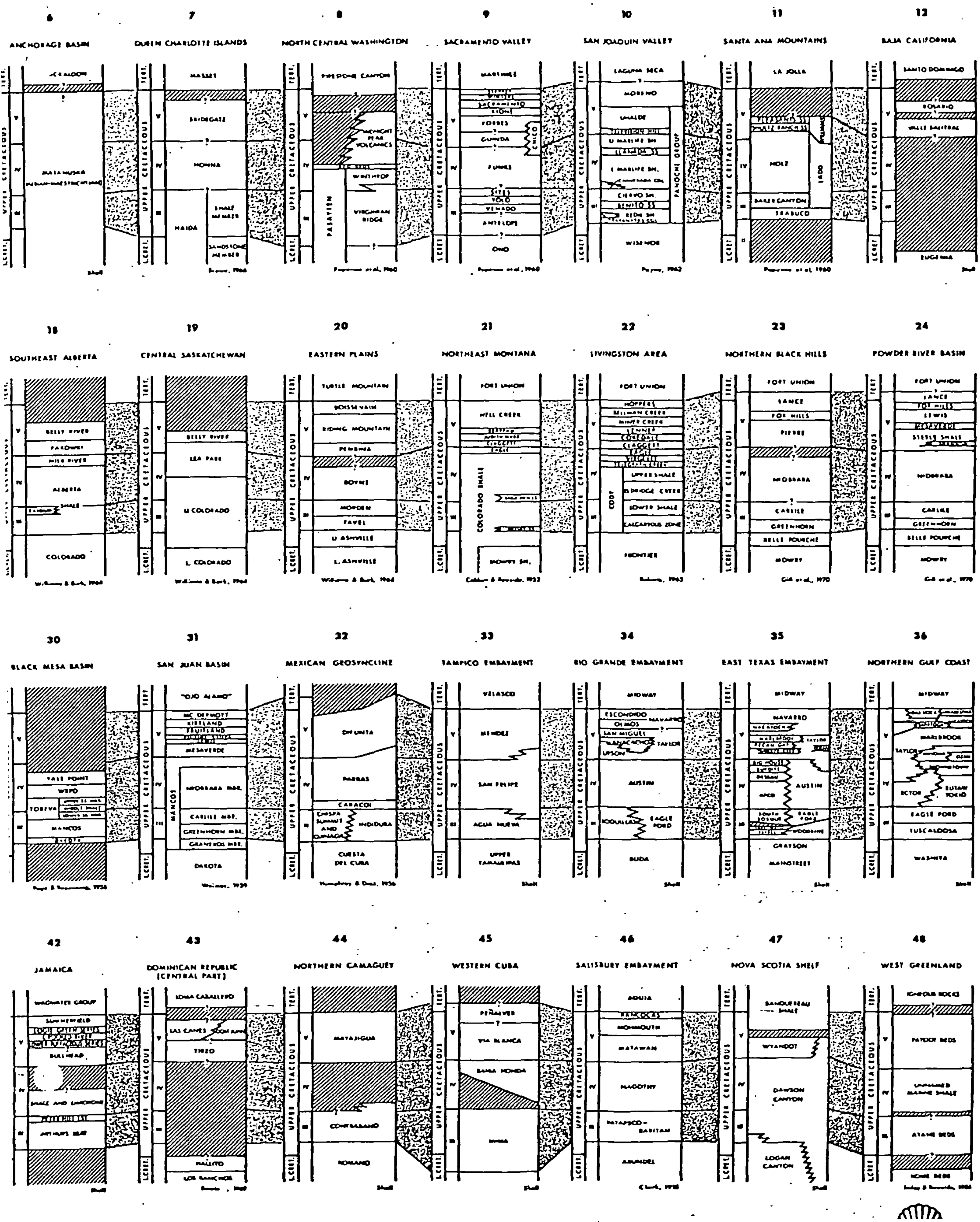
EOUS UNIT II
I - MID CENOMANIAN)



(MID-CENOMANIAN TO



AN TO TOP MAESTRICHTIAN)



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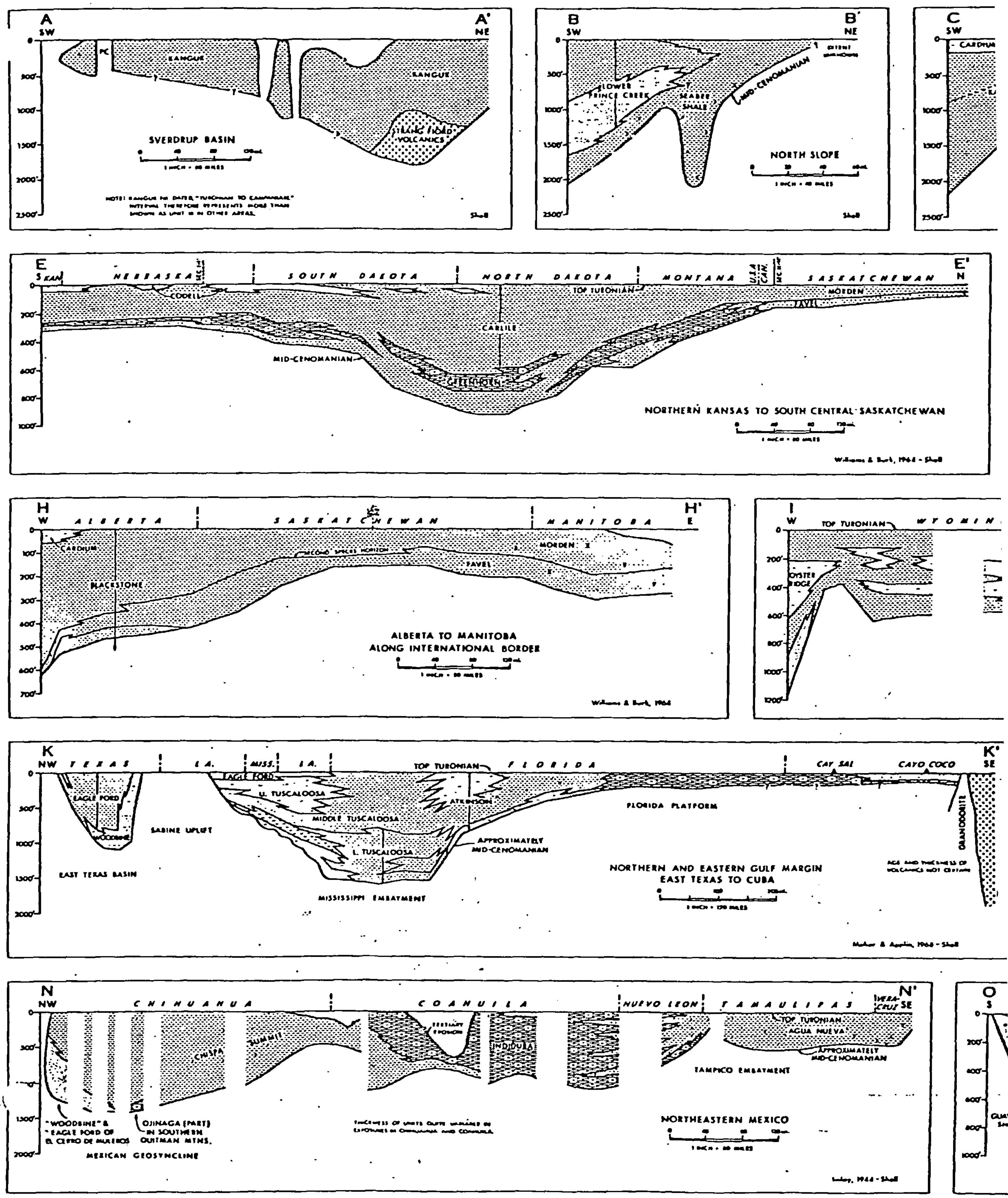
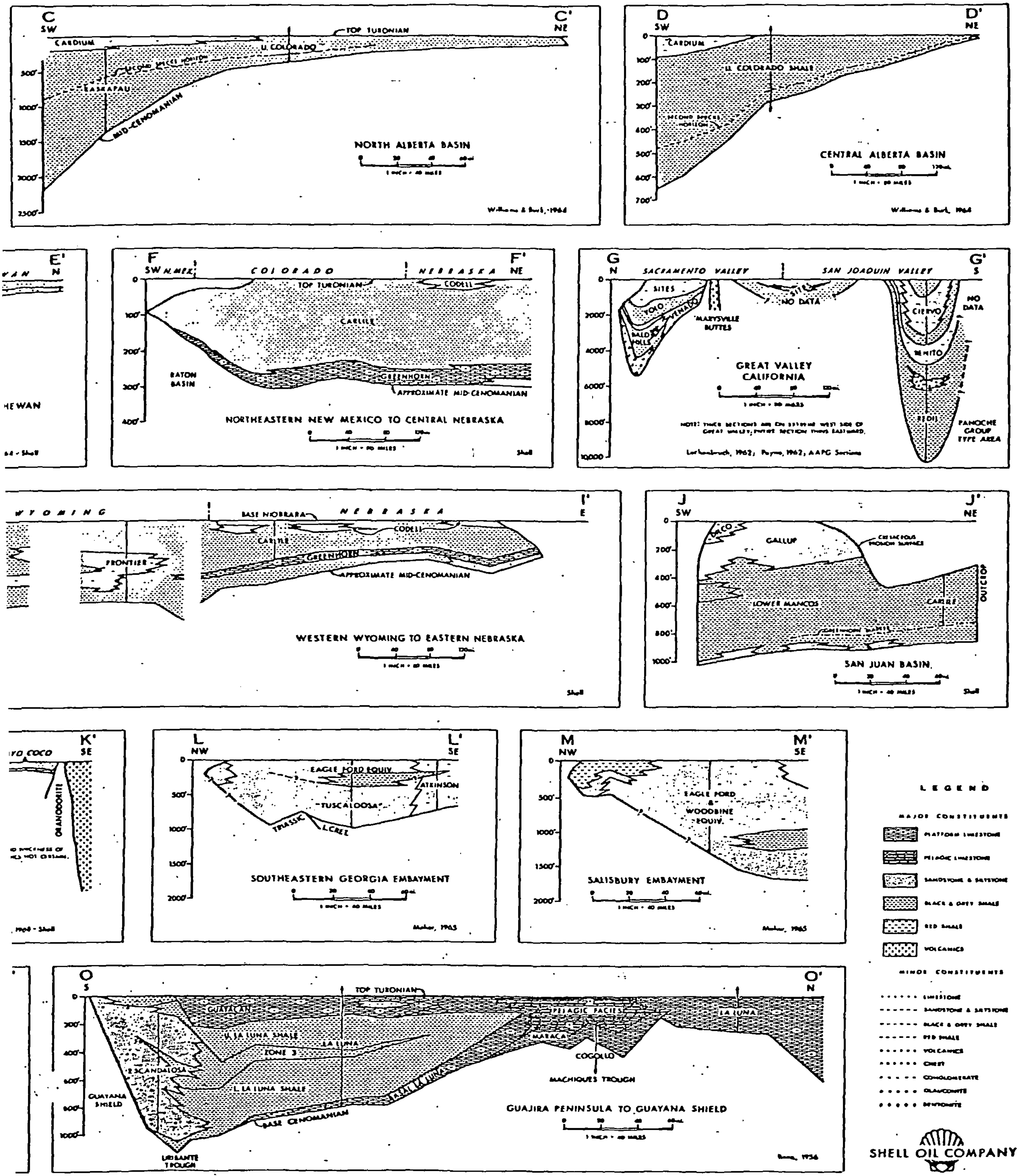
(MID CENOMANIAN - T



\CEOUS UNIT III
MANIAN - TOP TURONIAN)



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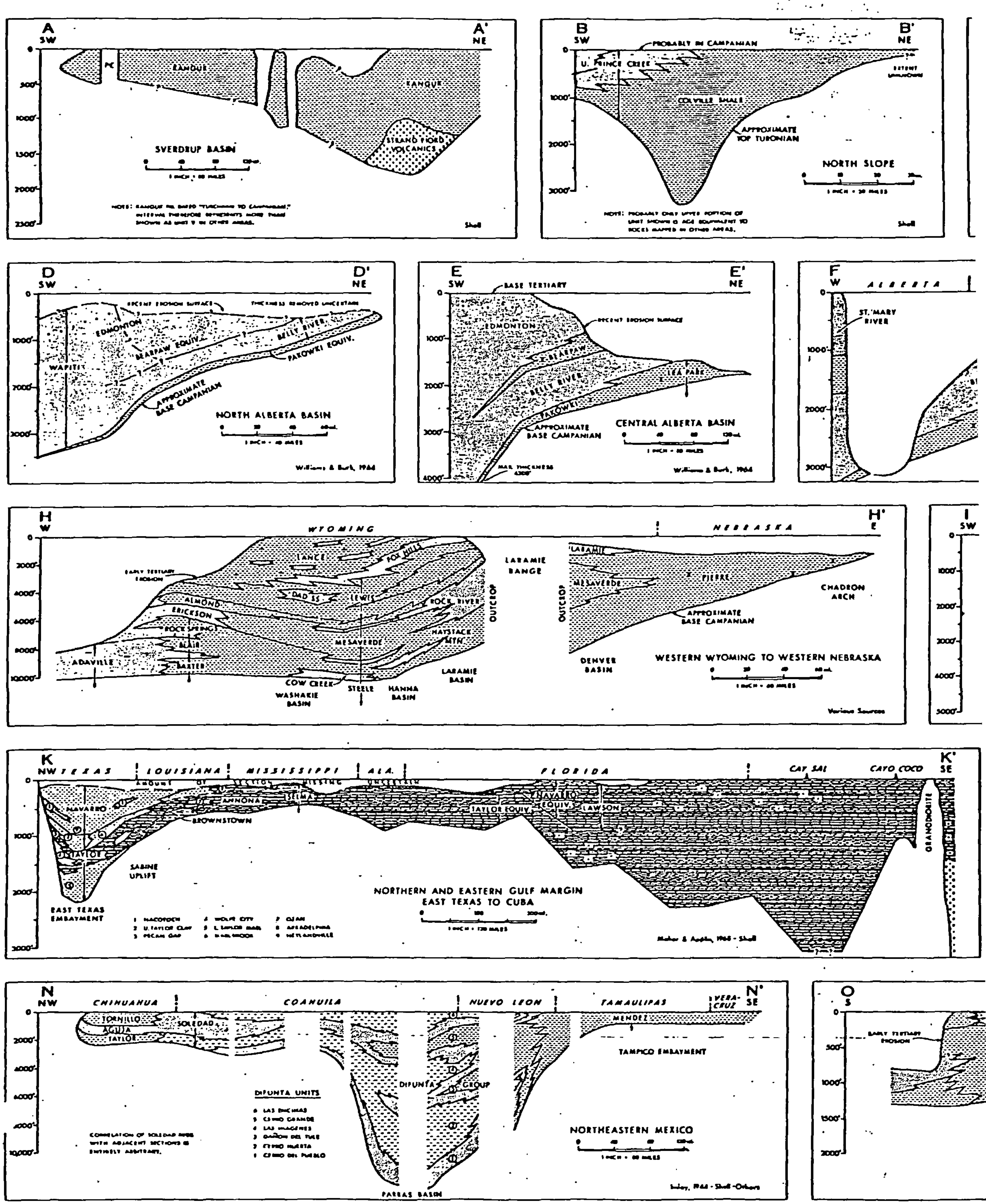
(CAMPANIAN - MAES



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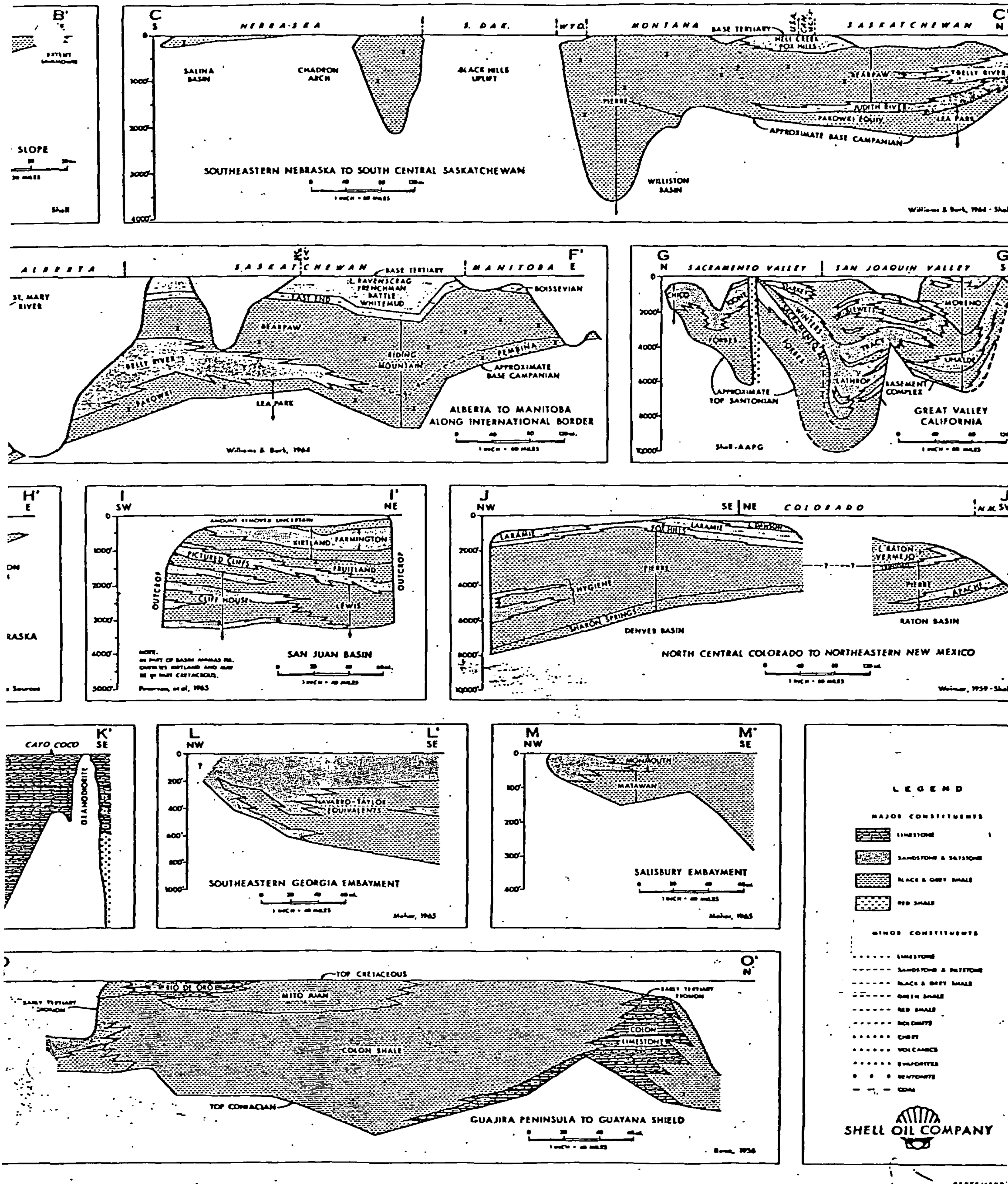
ETACEOUS UNIT V
PANIAN - MAESTRICHTIAN)



Environment No.*	Environment	Table No.**
1	Marine	3-1
	Deltaic	
	Slope	
2	Prodelta	3-2
3	Delta Front	3-3
	Distributary	
4	Distributary Channel	3-4
5	Distributary Mouth Bar	3-5
6	Levee	3-6
7	Crevasse Splay	3-7
	Interdistributary	
8	Swamp	3-8
9	Marsh	3-9
10	Bay	3-10
11	Lake-Pond	3-11
12	Tidal Channel	3-12
	Interdeltaic	
	Shallow Marine	
13	Offshore	3-13
14	Shoreface	3-14
15	Foreshore	3-15
	Beach	
16	Backshore	3-16
17	Tidal Inlet	3-17
18	Washover Fan	3-18
19	Chenier	3-19
	Interdeltaic (Con't.)	
	Lagoon	
20	Lagoon	3-20
21	Lagoonal Swamp-Marsh	3-21
22	Tidal Flat-Salt Marsh	3-22
23	Gilbert Delta	3-23
24	Estuary	3-24
	Lowland Plain	
25	Swamp	3-8
26	Marsh	3-9
27	Lake-Pond	3-11
28	Tidal Channel	3-12
29	Distributary Channel	3-4
30	Levee	3-6
31	Crevasse Splay	3-7
	Alluvial	
	Fluvial	
32	Braided	3-25
33	Transitional	3-26
34	Meandering	3-27
	Floodplain	
35	Overbank	3-28
36	Lake	3-29
37	Swamp	3-30

*Keyed to environments illustrated on Figu: 54 and 58
**Individual tables in Appendix

From. G. E. Vaninetti
"Coal Stratigraphy of the John Henry Member of the Straight Cliffs Fm. Kaiparowits Plateau, Utah.
M.S. thesis. Univ of Utah

2/21/80

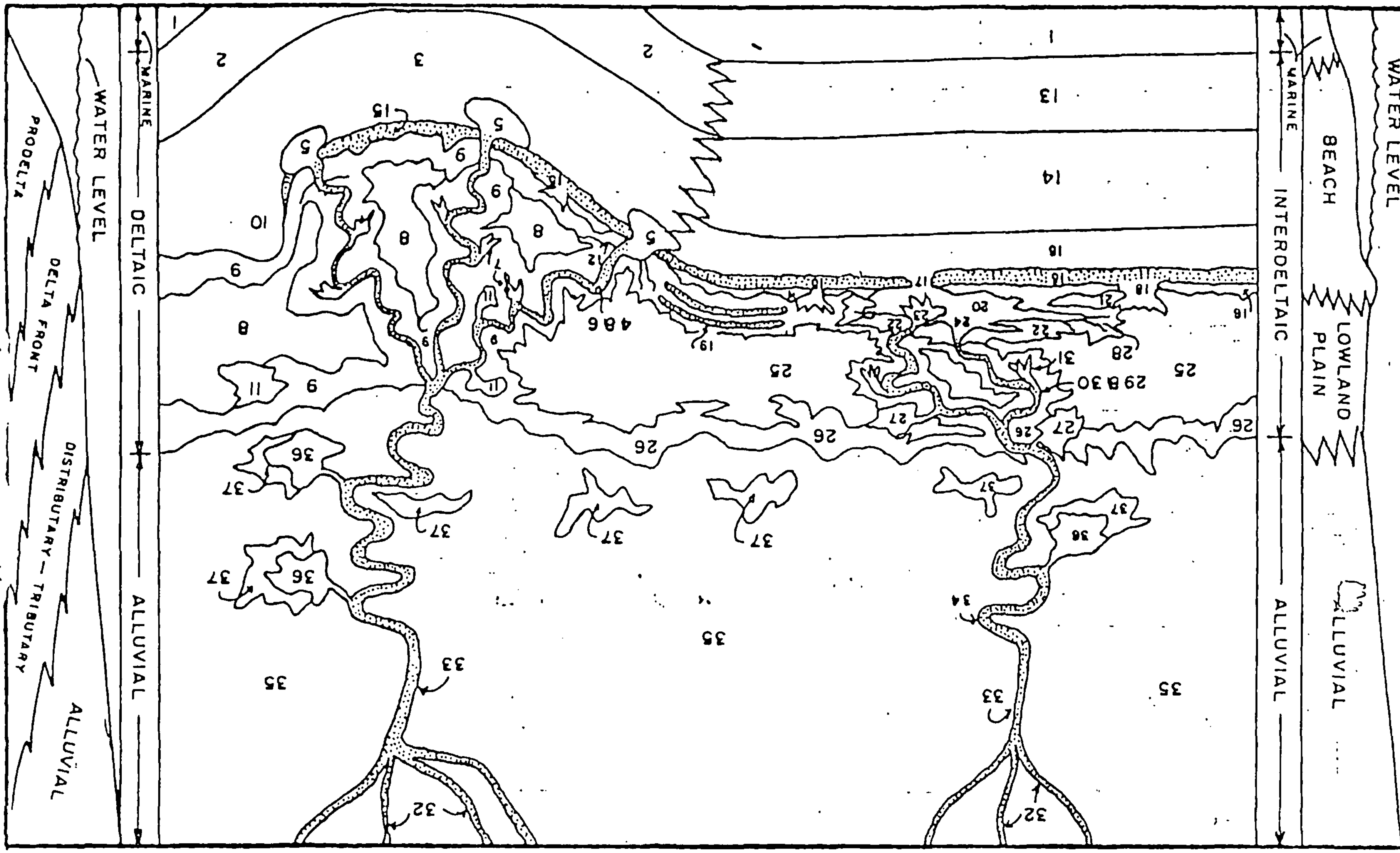


Figure 54. Plan-view of the depositional environments used in the environmental classification proposed in this study.

"Coal Stratigraphy of the John Henry Member of the Straight Cliffs Fm, Kaiparowits Plateau, Utah: M.S. thesis, Univ. of Utah

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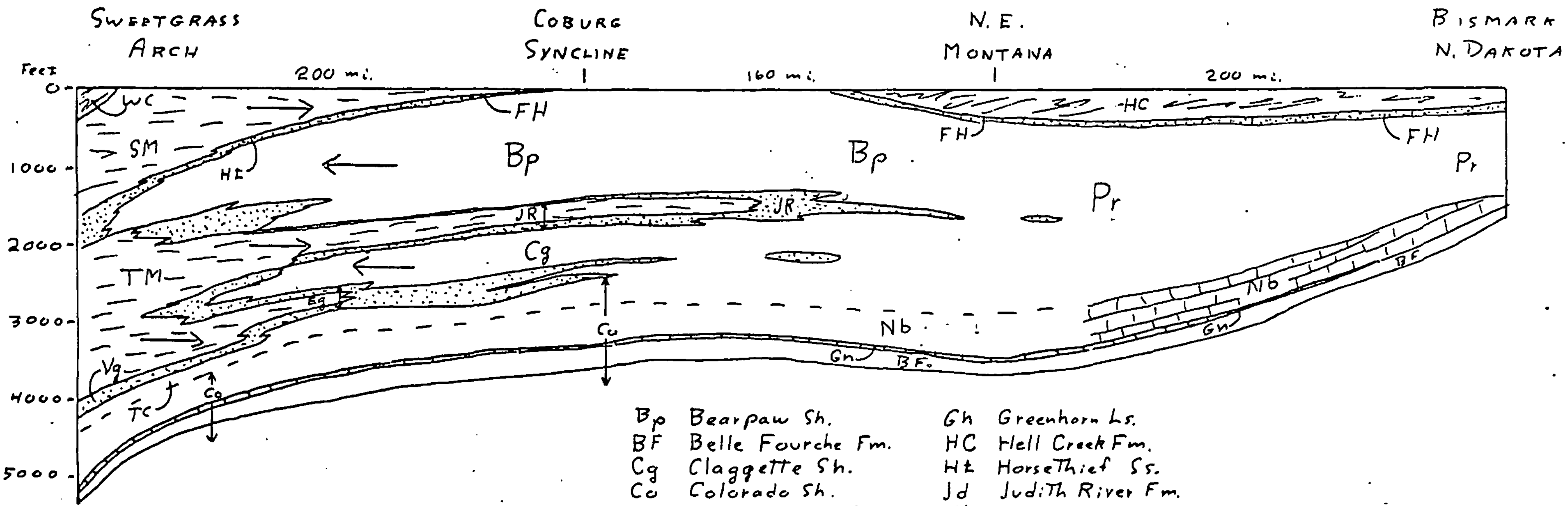
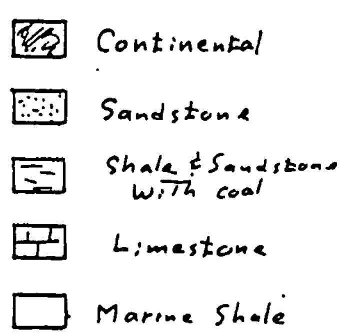


Bp	Bearpaw Sh.	Gh	Greenhorn Ls.
BF	Belle Fourche Fm.	HC	Hell Creek Fm.
Cg	Claggette Sh.	HT	HorseThief Ss.
Co	Colorado Sh.	Jd	Judith River Fm.
Eg	Eagle Fm.	Nb	Niobara Fm.
FH	Fox Hills Ss.	Pr	Pierre Sh.

SM St. Mary River Fm.
TC Telegraph Creek Fm.
TM Two Medicine Fm.
Vg Virgelle Ss.
WC Willow Creek Fm.



RESTORED CROSS SECTION No. 1
NORTHERN MONTANA-CENTRAL NORTH DAKOTA

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2/14/80

Mesozoic - Cenozoic Stratigraphy

Cretaceous Assgnment: Follow previous instructions. Do number 1, number 2 and one of the following four.

1. As you study the Cretaceous system you will find that the constituent formations in the Western Interior are being correlated in a number of ways: a) by absolute dating (Kauffman), b) by fossil zones (Kauffman, Cobban and Reeside, Red Book), c) by relation to the world standard (Cobban and Reeside) d) by relation to the reference section of the Western Interior (Cobban and Reeside, Red Book), e) by reference to designated cyclothemic sequences (Kauffman). Evidently the magnetostratigraphic method is not yet applicable.

Consider two well-known western formations, the Aspen Shale and Mesaverde Group, devise a chart showing how these two formations are classified according to the five methods mentioned above. I am hoping for something that will give a comparison. Which method or combination of methods is best for constructing a paleogeographic map? Think of what is practical more than what is potential.

2. Copy or trace the map in the Red Book showing coal fields of the Rocky Mountains (look under Coal). Give the explanation and names of the coal fields. Now, from the references add to your illustration the reserves estimated for the various fields.

3. Comment on the correlations of Cretaceous rocks of the Western Interior and those of a) Texas, and b) California. Use charts or diagrams if appropriate. Think of the actual means by which correlations have been and are being made.

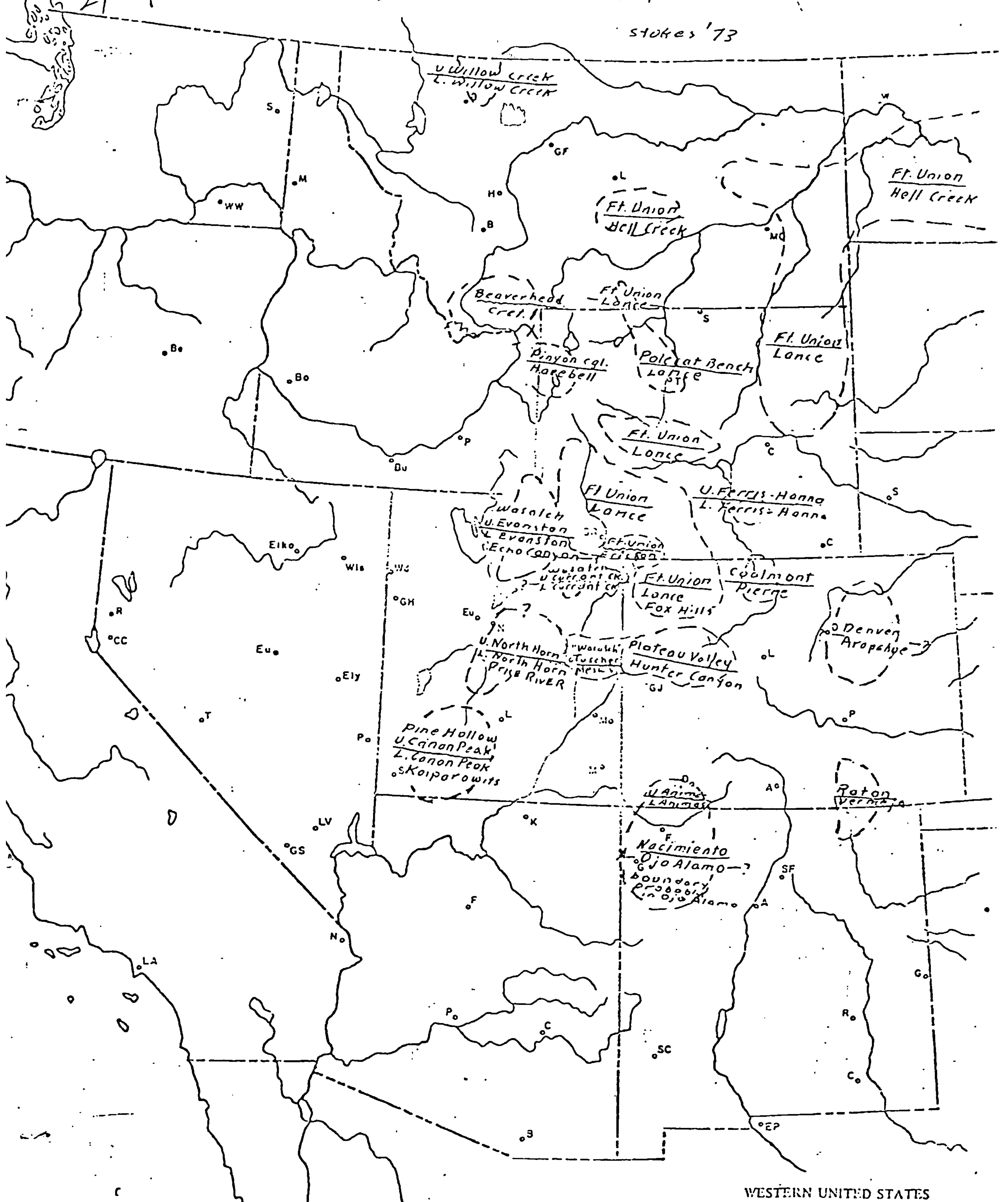
4. Follow instructions of # 3 with regard to correlations with northern Mexico and southern Arizona. Means correlate between the Interior and Mexico-Arizona.

5. Comment on the influence of the Idaho Batholith on western Cretaceous stratigraphy. Think of space relations, source material of various kinds, time relations and anything else that occurs to you. You will have to do some library work here.

6. Comment on the "space problem" of deriving the known volume of Cretaceous sediment from the indicated available land sreas.

Wm. L. Stokes, 1980

2/28/80 CRETACEOUS-TERTIARY BOUNDARY



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U of U Geology Dept.
W. L. STOKES
1971

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Summary of Cenozoic - Western Interior U.S.

W.L. Stokes, 1973

Paleocene: (5-7 million years)

Time of basin formation and concomitant deposition in the Rocky Mountains. The Rocky Mountain (or Laramide) Orogeny which commenced in the ~~Late Cretaceous~~ Paleocene continued very actively into the Paleocene. Most basin margins show angular unconformity between Cretaceous and Eocene rocks.

Great volumes of clastic sediment produced from uplifts were deposited chiefly in nearby basins. Except in the northern Great Plains little sediment left the mountains. The Cannonball Formation records last of the interior seas.

Lake beds locally, Flagstaff Lake, Central Utah, was largest.

Much lignite in Montana-Wyoming-North Dakota, carbonaceous material almost everywhere in Paleocene indicates heavy organic production. Climate generally humid, subtropical.

Not much igneous material except in Montana.

Great Basin was high and undergoing strong erosion. Very little Paleocene has been identified here.

Limited record in Pacific Northwest. Pacific Ocean extended eastward into area of present Cascades, not much igneous material.

Climate tropical to sub-tropical with appropriate animal and plant life.

Eocene: (22 million years)

Decline in active diastrophism but rapid erosion of previously created uplifts continued. This was the declining stage of the Rocky Mountain Orogeny. Little material left the area to build deposits outside the intermountain basins.

Sediments include: great conglomerate deposits (Wasatch): "red beds" and gray-banded, fine fluvial clastics in many basins; lacustrine deposits of mixed lithology with marlstone and oil shale as notable components. Green River Lake System is important in Wyoming-Colorado-Utah. Routes of external drainage unknown, water generally fresh but occasionally saline.

Great Basin remained higher than surroundings but begins to collapse late in epoch.

Sierra Nevada was relatively low with through-flowing drainage.

Pacific Northwest begins to show volcanic action but is generally low with coal forests along coast.

Increasing volcanic action in Yellowstone area. (Absaroka field, N.W. Wyo.)

Interior generally at low elevation, Green River lakes thought to be 1000'± of sea level. Vegetation generally sub-tropical, crocodiles, turtles and many primitive mammals.

Oligocene: (11 million years)

Deformation declines in importance but volcanic action reaches a maximum. Basins are mostly filled and the sediment cover rises to cover low uplifts and all but the higher parts of the major ranges. Higher erosional surfaces, now exhumed, record the height of the general fill.

Sediments generally contain much volcanic debris and are of light color because of the ash and tuff and also from considerable calcareous material.

Notable volcanic centers were in the Yellowstone area of Wyoming and the San Juan Mountains of Colorado. Greatest eruptions and volcanic effects were in the Great Basin where immense outpourings accompanied a general subsidence. In southern Utah-Nevada ignimbrites were produced on a large scale. Colorado Plateau "laccoliths" emplaced. There was also an increase of volcanism in the Pacific Northwest.

Stream systems began to be integrated with production of the present river systems. Superposed drainage patterns which characterize the Rockies have been let down from chiefly Oligocene surfaces.

Sierra Nevada low enough to be crossed by streams flowing westward out of the Great Basin.

Building of extensive light-colored deposits on the plains area of South Dakota-Nebraska generally known as the White River Formation. Extensive fauna of mammals that may be the most complete of any in existence comes from the badlands of this formation in South Dakota and Wyoming.

John Day beds of Oregon have lithology and fauna like the White River.

Florrisant Beds, isolated in basin in Precambrian, has important fauna and flora.

Miocene: (12 million years)

General uplift of western U.S. and possibly of the entire continent. (Shown by climatic change, by more hardy plant and animal life, by incision of streams, and rejuvination of mountains and by great volume of sediments deposited in Gulf Coast Geosyncline). Uplift may have been 2 to 5 thousand feet. Epeirogeny, not orogeny.

Erosion moved much of the Oligocene material and cut deeply into the older rocks. Miocene deposits are of limited extent in the Rocky Mountains proper but there are remnants of fair size in the Wyoming-Utah-Colorado corner (Browns Park Formation) and in southwestern Montana (Bozeman). Pedimentary deposits such as Bishop conglomerate may be this age.

Great Basin was subsiding and breaking up by normal faults which outline the present horsts and grabens. There may be great volumes of Miocene in the depressions but outcrops are rather limited. Light color of Great Basin Miocene is characteristic. Drainage was westward across the Sierra until late in Miocene; climate was relatively humid with extensive hardwood forests. Volcanics + Diatomaceous (Salt Lake Group)

Pacific Northwest affected by immense volcanic outpouring (the Columbia River basalts up to 12,000' thick and covering 200,000 square miles). Cascade range begins uplift late in epoch; this may be the locality of the last batholith to be emplaced in the U.S.

Extensive erosion in Colorado Plateau but routes of rivers is a major problem -- did the Colorado River exist?

Deposition on the interior plains east of the Rocky Mountains was chiefly in the central region -- Nebraska has best section. Much of the filling of the Rio Grande Trench is of this age.

Climates cooling and drying, disappearance of warm-loving fauna, savanna vegetation, spread of grass.

Pliocene: (13-15 million years)

A relatively quiet period tectonically and with only moderate volcanic activity.

Sedimentation at a low ebb with relatively little material produced or retained in the Rocky Mountains. Gravel surfaces are notable.

Climate both dry and cool The low precipitation may explain the retarded erosion, lack of lakes and small volume of sediment.

Deposits on the plains are usually grouped under one designation, the Ogalalla Formation. Its distribution is chiefly in the High Plains -- Nebraska, Kansas and northwest Texas. This is a mixed fluvial-lacustrine-aeolian deposit, average 200 feet thick. Rio Grande Trench continues to be filled.

Erosion slows in the Colorado Plateau -- problem of age of the Colorado River System. In general the river systems of the Rockies are becoming more integrated but those of the Great Basin are breaking up and no longer cross the Sierra. Ponding in the Great Basin, Columbia Plateau and Snake River Plains aided by lava dams.

In the Pacific Northwest extensive lava flows continue to build; older surfaces and topography submerged, Cascade Range grows both by uplift and formation of large cones.

Decline of warm-loving fauna and flora continues. This may be the driest of the Tertiary epochs. "Autumn of the Cenozoic".

A difficult epoch to interpret because of scarcity of deposits and fewness of good fossils.

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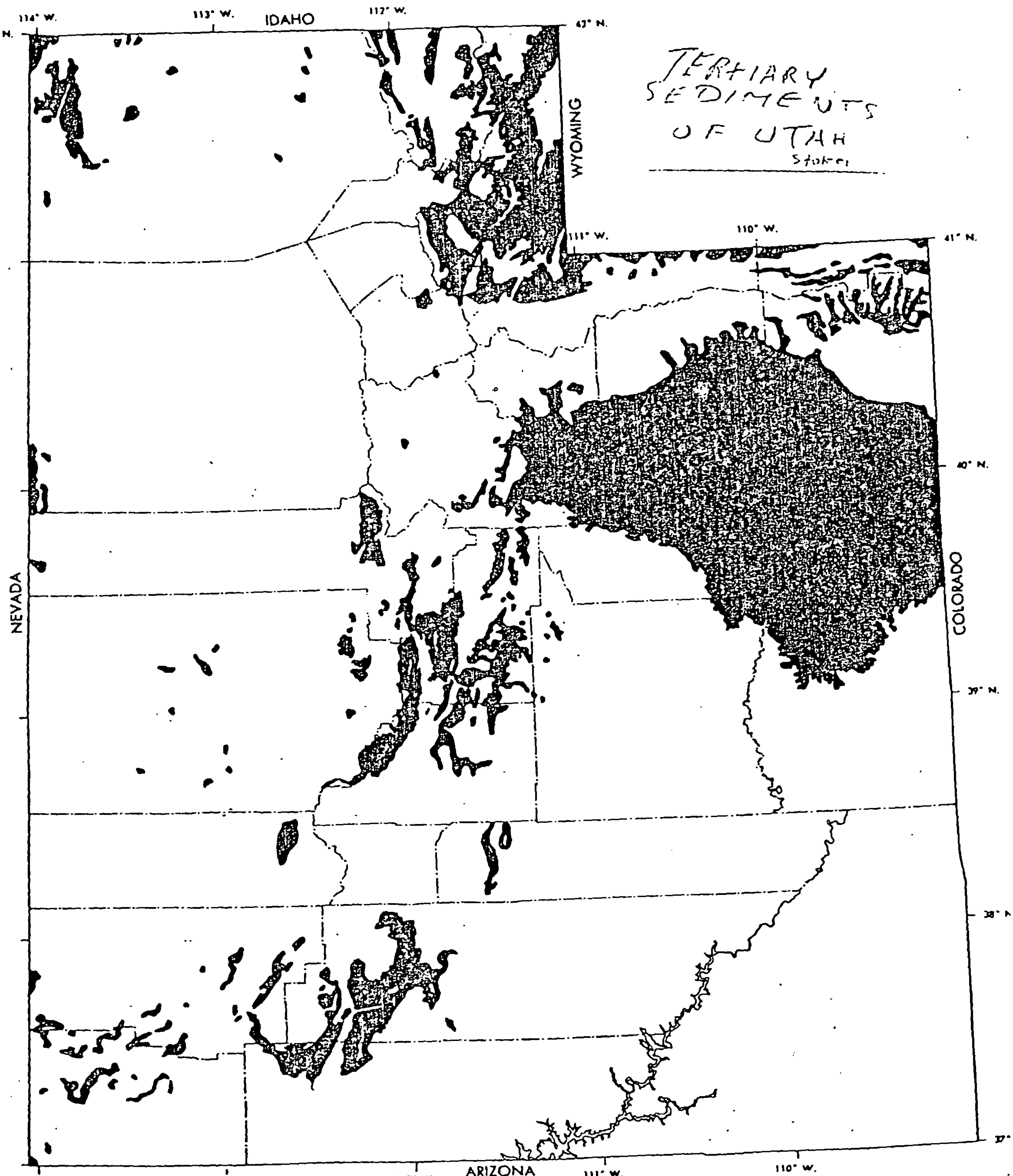
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TERTIARY SEDIMENTS OF UTAH
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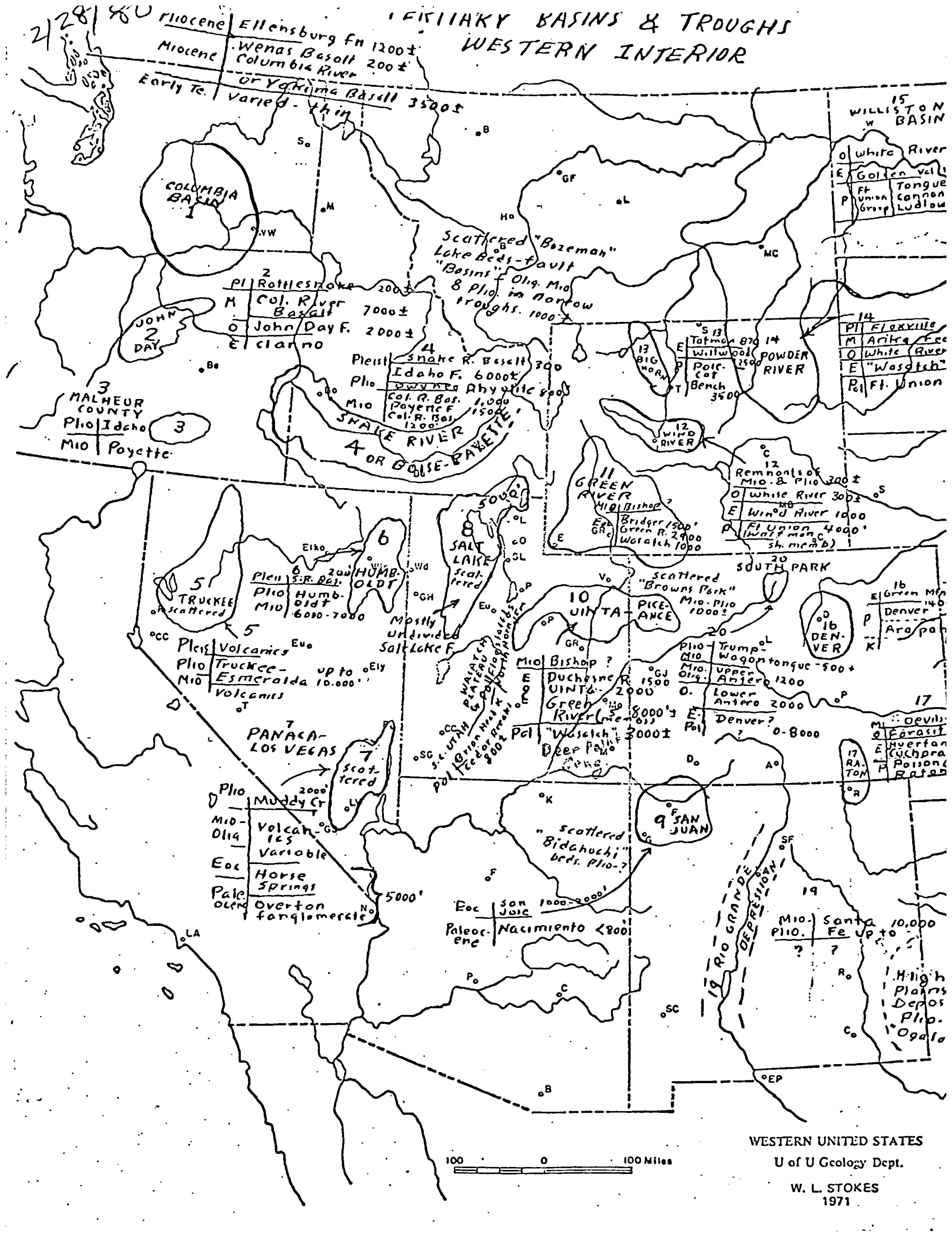

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TERTIARY BASINS & TROUGHS
WESTERN INTERIOR
Pliocene Ellensburg Fn 1200±
Miocene Wenas Basalt 200±
Columbia River or Yakima Basalt 3500±
Early Te. varied - thin
COLUMBIA BASIN
1
2
Pl Rattlesnake 200±
M Col. River Basalt 7000±
O John Day F. 2000±
E Clarno
JOHN 2 DAY
3
MALHEUR COUNTY
Plio Idaho
Mio Poyette
4
Pleist Snake R. Basalt 300
Plio Idaho F. 6000±
Owyhee Rhyolite 600
Mio Col. R. Bas. Poyette F Col. R. Bas. 1,000 1500 1200
SNAKE RIVER
4 OR BOISE-PAYETTE
Scattered "Bozeman" Lake Beds - fault "Basins" Olig. Mio & Plio. in narrow troughs. 1000
15
WILLISTON BASIN
O White River
E Golden Vall
P Ft Union Group Tongue Cannon Ludlow
14
Pl Flaxville
M Arikaree
O White River
E "Wasatch"
Pal Ft. Union
13
BIG HORN
E Tatman 870 Willwood 2500
P Pole-cat Bench 3500
14
POWDER RIVER
12
WIND RIVER
12
Remnants of Mio. & Plio 300±
O White River 300±
E Wind River 1000
P Ft Union (Waltman sh. memb) 4000'
11
GREEN RIVER
Mio Bishop
Eoc Bridger 1500' Green R. 2900 Wasatch 1000
20
SOUTH PARK
16
DENVER
E Green Mtn
P Denver
K Arapahoe
5
TRUCKEE scattered
5
Pleis Volcanics
Plio Truckee-
Mio Esmeralda up to 10,000'
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6
HUMBOLDT
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Pleis S.R. Bas. 200
Plio Humboldt
Mio 6000-7000
8
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"Scattered "Browns Park" Mio-Plio 1000±
Mio Bishop ?
E Duchesne R. 1500
UINTA 2000
Green River 8000's
Pal "Wasatch" 3000±
WASATCH PLATEAU
S.C. UTAH
Pal (Brian Head) Claron 800±
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Plio-Mio Trump-Wagontongue 500+
Mio-Olig. Upper Antero 1200
O. Lower Antero 2000
E. Pal Denver? 0-8000
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M Devils
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E Huerfano
P Cuchara Poison C. Raton
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PANACA-LOS VEGAS
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Paleocene Nacimiento <800'
19
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WESTERN UNITED STATES
U of U Geology Dept.
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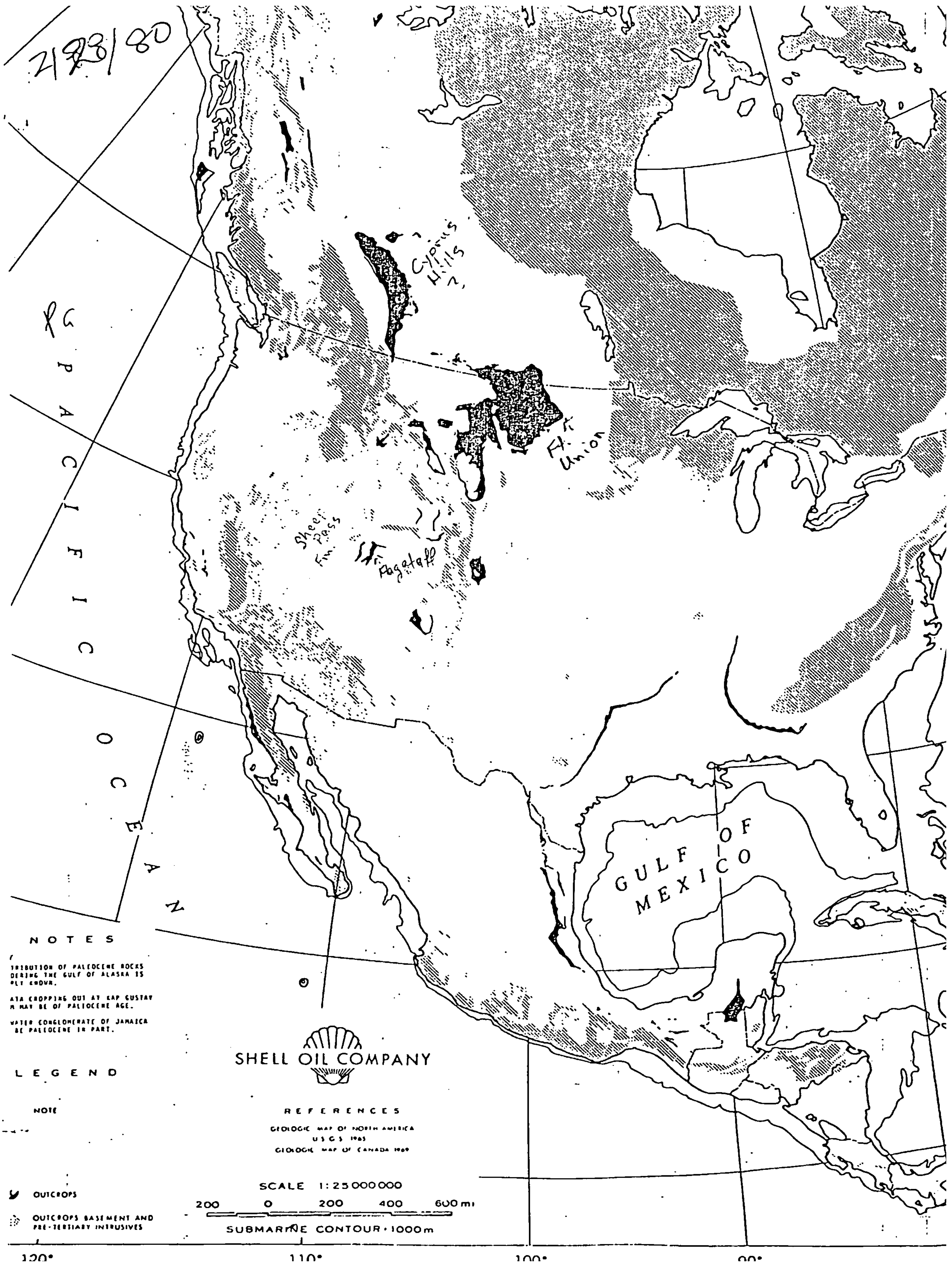

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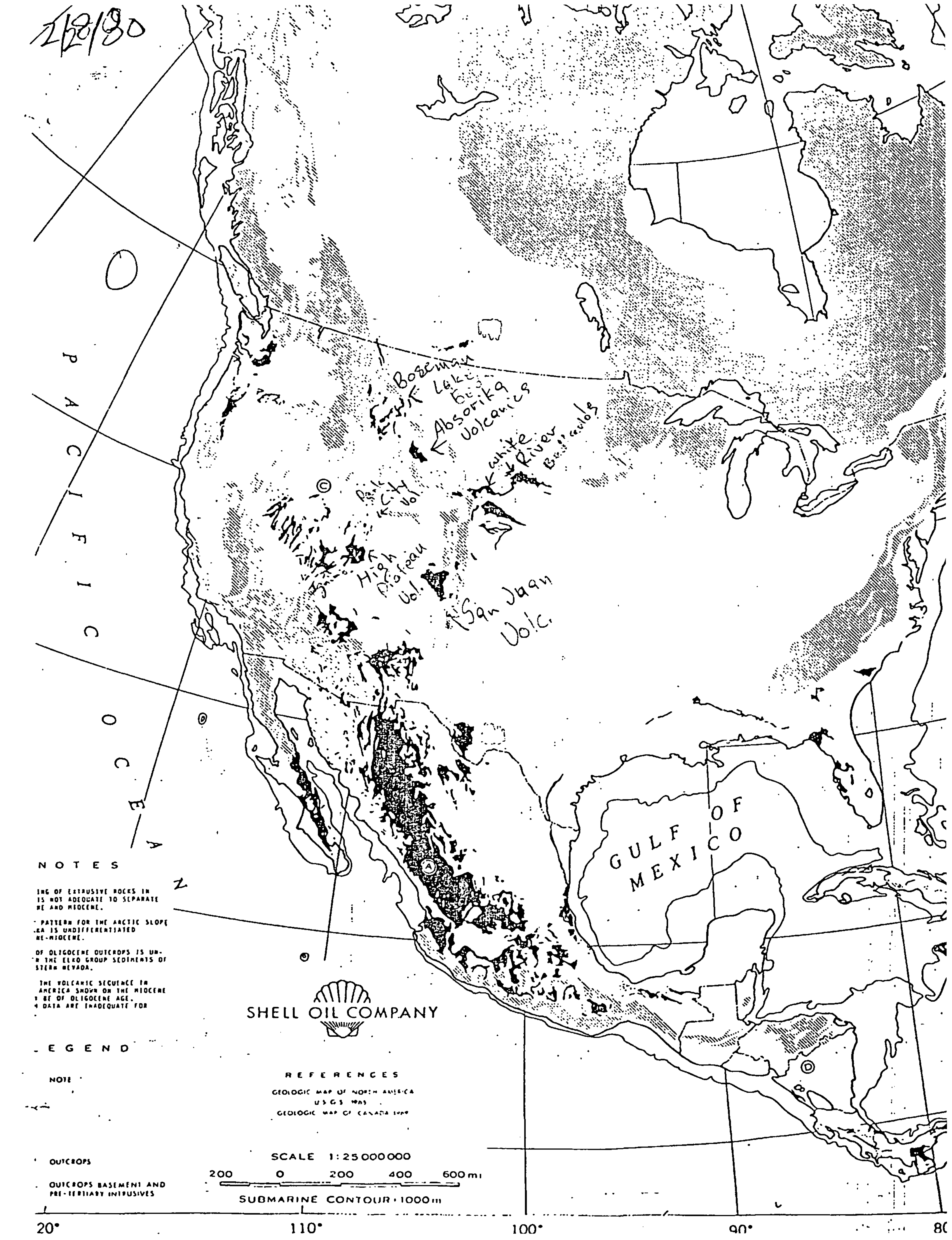
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High Plateau Vol.
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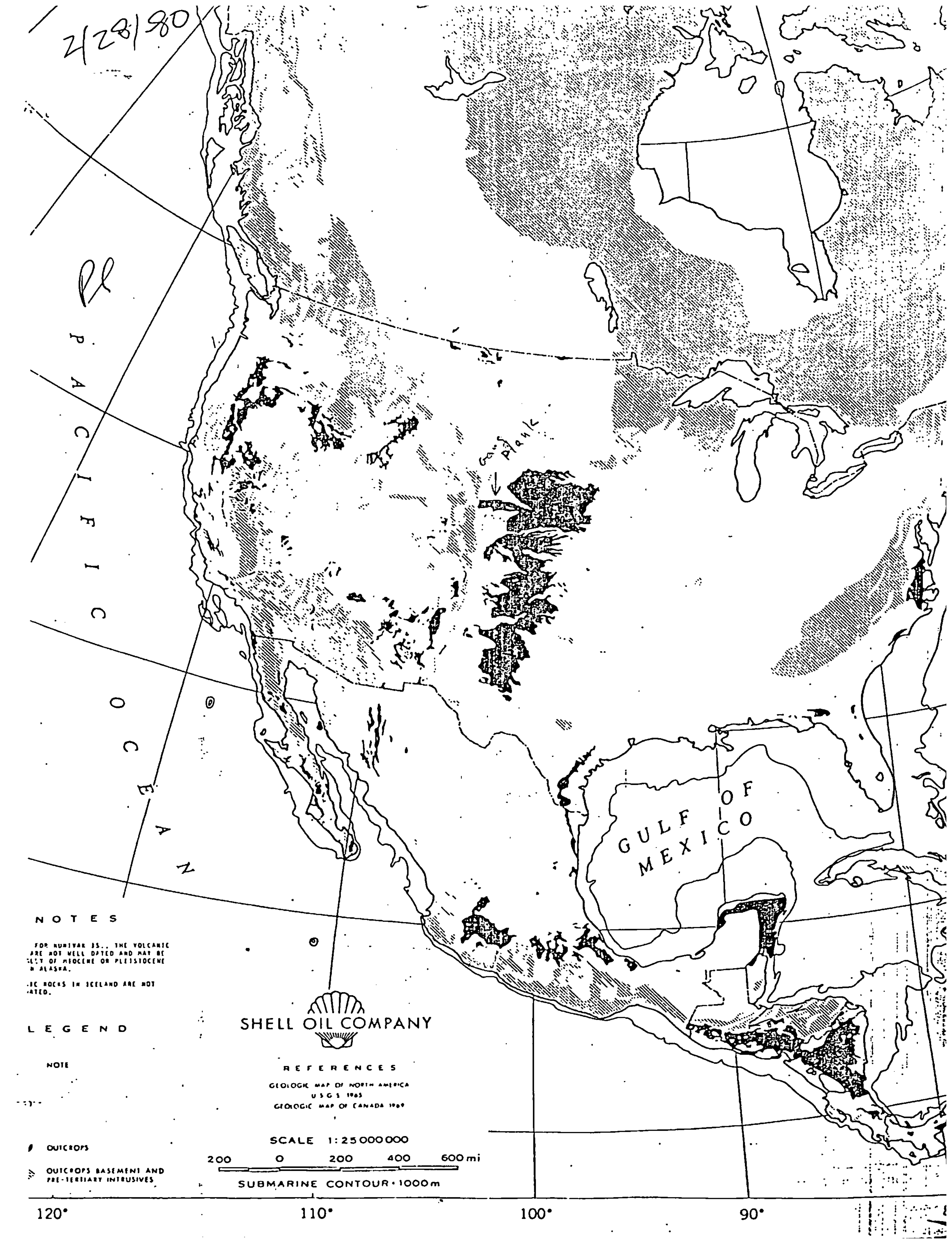
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Geological Society of America Bulletin, v. 87, p. 496–512, 16 figs., April 1976, Doc. no. 60402.

Early Tertiary sedimentation in the western Uinta Basin, Utah

ROBERT T. RYDER*
THOMAS D. FOUCH*
JAMES H. ELISON
Shell Oil Company, Western Division Exploration, P.O. Box 831, Houston, Texas 77001
3112 18th Street, Suite 6, Bakersfield, California 93301

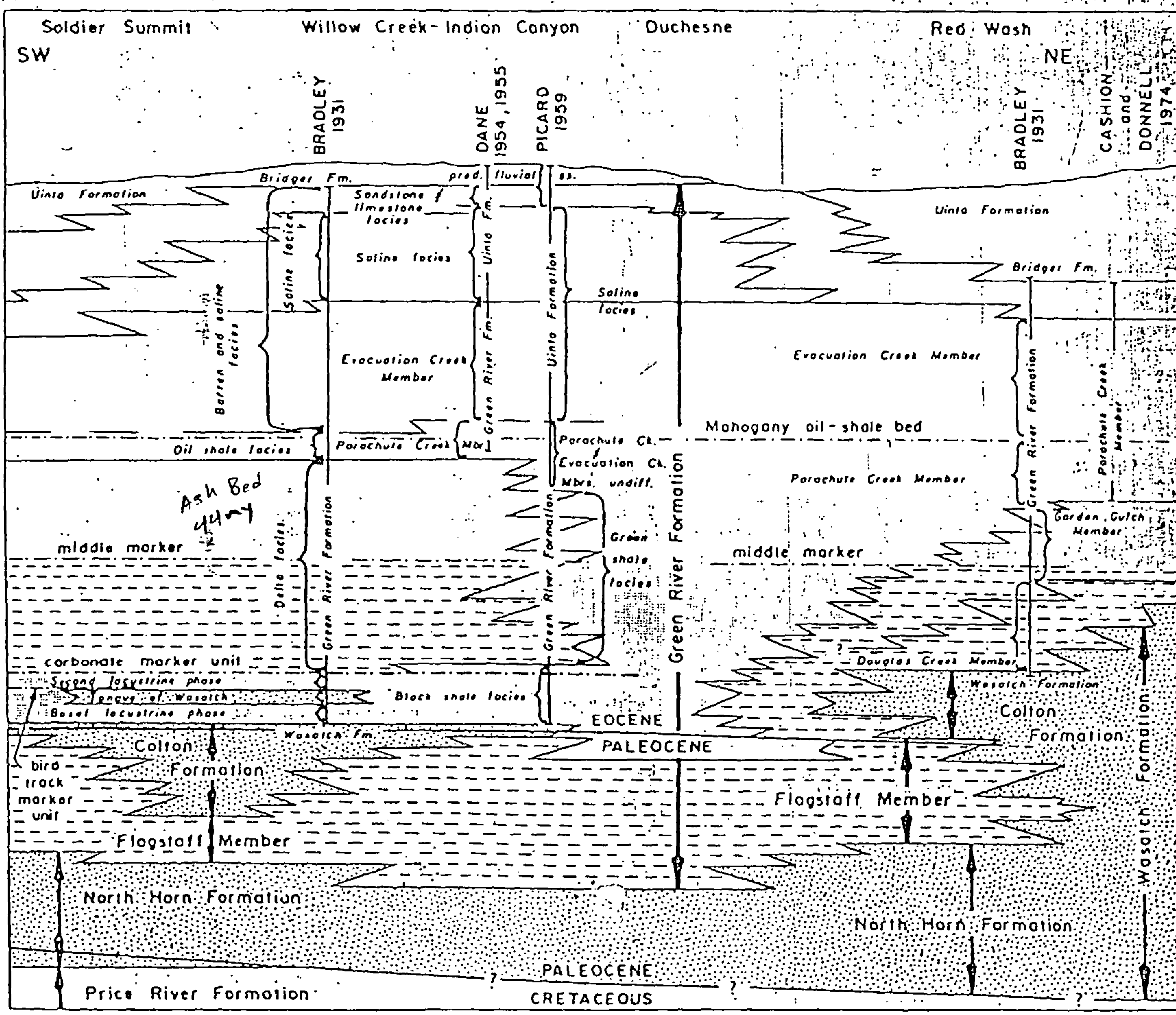


Figure 2. Generalized stratigraphic framework and nomenclature of lower Tertiary rocks in the Uinta Basin, Utah. Diagram is not to scale. See Figure 1 for orientation of the stratigraphic section. Dot-dash lines indicate major mechanical log markers and lithologic marker units. Italicized names indicate nomenclature introduced by previous authors. Formation and member names used in this paper are enclosed by vertical arrows. Major rock units are indicated by the following patterns: large-dot pattern, unit dominated by channel-form sandstone, red claystone, and siltstone; dash pattern, unit dominated by gray to green claystone, sandstone, and mud- to grain-supported carbonate; fine-dot pattern, unit composed of dark gray to brown claystone and mud-supported carbonate.

3/6/80

GEOLOGIC HISTORY OF SITE OF UINTA BASIN, UTAH

John C. Osmond, Consulting Geologist, Salt Lake City, Utah

The Uinta basin in northeastern Utah includes an area of 9,300 square miles. The differential vertical movements which created the basin and its rim began in Paleocene or Eocene time, during the deposition of the Wasatch Formation. The synclinal axis of the basin trends easterly and is slightly convex northward. The basin is asymmetric with a broad gentle-dipping south flank and a north flank which is up to 10 miles wide with beds near vertical and locally overthrust.

The configuration of the Uinta basin is controlled in part by pre-existing structures and geologic trends, but much of the present rim is the result of Tertiary tectonics.

The dominant tectonic factor in the development of the basin is the rise of the Uinta Mountains block and the simultaneous subsidence of the synclinal axis of the basin. This major flexure in the crust probably conforms to the edge of the late Precambrian trough in which the rocks of the Uinta Mountains block were deposited. This belt of Precambrian rocks was essentially dormant until Eocene when it began to rise from depths of about 16,000 feet below sea level to elevations of 7,000 to 13,000 feet above sea level.

During lower Paleozoic the Uinta basin area was on a stable crust just east of the Cordilleran geosyncline. In the Pennsylvanian Period the Uncompahgre uplift formed a mountain range part of which now underlies the southeastern part of the present basin. This major tectonic feature had a northwest trend, and several small folds or faults with similar trends developed in the eastern part of the basin. These structures were gradually masked by Mesozoic sediments. Slight Tertiary upwarping and differential compaction allow some of these structures to be reflected as folds plunging into the eastern part of the Uinta basin.

In Cretaceous time the Rocky Mountain geosyncline occupied the region and received clastic sediments from the Cordilleran geanticline in western Utah. In Late Cretaceous time easterly directed compressive forces resulted in folding and over thrusting at the western edge of the Uinta basin area and possibly also caused slight arching of the eastern edge.

During Tertiary time western North America was elevated above sea level. Concurrent with this epeirogeny the mountain ranges and basins of the Rocky Mountain Region were developed by differential uplift. To the west of the Rockies the Basin and Range province was also elevated, and tilted fault-block mountains and valleys were created. This difference between the Rockies and the Basin and Range province reflects the difference in the Paleozoic between the stable block and the geosyncline.

The Uinta basin was outlined in Tertiary time by the central part not being elevated as high as its rim. The northern sector of the rim was the most active in pushing upward. The eastern and southern sectors of the rim were formed by the stable elements supported by the Douglas creek arch and part of the Uncompahgre block. The southwestern sector of the rim was formed by the San Rafael swell, a Tertiary anticline formed by subsidence of adjacent areas and probably localized by an upper Paleozoic positive trend.

The western sector of the rim of the Uinta basin was created by the interplay of tilting of fault blocks above the eastern margin of the Paleozoic geosyncline and north-trending faults that developed over Mesozoic troughs which, in turn were superimposed on the margin of the geosyncline.

The structure of the Uinta basin is the result of regional uplift of a heterogeneous area of the crust which incorporated both the sturdy and the weak, or weakened, products of prior deformations. The absence of intrusive igneous rocks is striking.

STRUCTURAL AND STRATIGRAPHIC DEVELOPMENT, WASHAKIE AND SAND WASH BASINS, WYOMING AND COLORADO

Howard R. Ritzma, Consulting Geologist, Denver, Colorado

The Washakie and Sand Wash basins, southeast segments of the over-all Green River basin, are located in central-southern Wyoming and northwestern Colorado. The two basins are bounded by and contain structural elements of varying age. These are: Rock Springs uplift and Wamsutter arch (Late Eocene-Oligocene) to west and north, Sierra Madre and Park Range uplifts (Late Cretaceous-Paleocene) to east, and the White River, Axial and Uinta Mountain uplift (Paleocene-Eocene) from southeast through southwest. The basins are separated by a low arch and fault zone of late Tertiary age, which parallels the Wyoming-Colorado boundary.

The present obvious structural trends are the composite of many less obvious structural episodes of Late Cretaceous-Paleocene, early Late Cretaceous, Early Cretaceous, Permian- Pennsylvanian and early Paleozoic time. Structural elements related to these older orogenic episodes are now mostly concealed beneath younger sediments and basin downwarping. Evidence for many structural episodes, particularly those of the early Paleozoic, have been all but effaced by erosion related to younger orogeny. Maximun structural activity offured from latest Cretaceous through Mid-Eocene time.

The full sedimentary column (Eocene and older) in the Sand Wash basin exceeds 32,000 feet apportioned approximately as follows:

pre-Pennsylvanian	3%
Pennsylvanian-Early Cretaceous	15%
Late Cretaceous	46%
Paleocene-Eocene	36%

The maximum sedimentary column in the Washakie basin may exceed 36,000 feet in the central and northern parts of the basin. Several thousands of feet of late Tertiary sediments and igneous extrusives occur in limited areas of both basins.

Precambrian structural trends are imperfectly known in the bounding Uinta, Park Range and Sierra Madre uplifts, but have had obvious, important influence on subsequent structural trends and movements to the present.

3/6/80

OQUIRRH AND PHOSPHORIA BASINS IN NORTHWESTERN UTAH, NORTHEASTERN NEVADA, AND SOUTHERN IDAHO

R. J. Roberts, E. W. Tooker, H.T. Morris, M.D. Crittenden, and R.K. Hose, U.S. Geological Survey, Menlo Park, California and T.M. Cheney, Palo Alto, California

The Oquirrh and Phosphoria basins in northwestern Utah, northeastern Nevada, and southern Idaho are downwarped segments of the Cordilleran geosyncline superposed on a complex structural pattern of Precambrian and early and middle Paleozoic age.

The Oquirrh basin contains Pennsylvanian and Early Permian sedimentary rocks as much as 26,000 feet thick; the area of maximum sedimentation was in westcentral Utah. Three principle units are recognized: a lower unit of cyclically bedded bioclastic limestone and sandy limestone, a middle unit of interbedded limestone and quartzite, and an upper unit of quartzite. The lower and middle units were mostly deposited in an offshore shallow water environment; the upper unit in offshore moderately deep water; these grade laterally both eastward and westward into shallow nearshore facies.

The Middle Permian Phosphoria basin was partly coextensive with the Oquirrh basin, but the area of maximum sedimentation was in northeastern Nevada and southern Idaho, where locally 3,500 feet of shale, cherty shale, chert, dolomite, and limestone accumulated. This facies was deposited in an offshore deepwater environment, favoring formation of thick sponge-spicule chert and cherty shale units; these grade southward and eastward into carbonates and shales that were deposited in shallow nearshore environments.

In Cretaceous time the Paleozoic and early Mesozoic rocks of northeastern Nevada and western Utah moved eastward on great thrust plates that extended from southern Utah into Idaho. Movement took place on the Willard-Charleston-Nebo thrust belt in the Wasatch Mountains. Westward continuations of these thrusts crop out in northwestern Utah and eastern Nevada. Imbricate thrusts and tear faults within the upper plate have resulted in complex distribution of late Paleozoic facies.

GEOLOGIC HISTORY OF WIND RIVER BASIN, CENTRAL WYOMING

William R. Keefer, U.S. Geological Survey, Denver, Colorado

The Wind River basin was part of the stable shelf region that lay east of the Cordilleran geosyncline during Paleozoic and much of Mesozoic time. Rocks representing all systems except possibly the Silurian were deposited across the area during repeated transgression and regression of the epicontinental seas. Most formations are thicker and more complete in the western part of the basin than in the eastern part, and some units disappear eastward owing to truncation or nondeposition. Depositional environments, generally marine, were often influenced locally by slight fluctuations in sea level or by tectonic movements. The latter were limited to broad upwarping and downwarping along trends which with few exceptions, show little direct relation to structural trends developed later during Laramide deformation.

Near the close of the Jurassic, highlands began to form in the geosynclinal area west of Wyoming, and the major sites of deposition shifted eastward. During Late Cretaceous time the seaways lay in eastern Wyoming, and a thick sequence of alternating transgressive, regressive, and nonmarine deposits accumulated across the Wind River basin area. The latest marine invasion (represented by the Lewis Shale) covered only the eastern part of the basin.

Laramide deformation began in latest Cretaceous time with downwarping of the basin trough and broad doming of parts of the peripheral areas. The intensity of movement increased through the Paleocene, and culminated in earliest Eocene time in high mountains that were uplifted along reverse faults. A complete record of orogenic events is preserved in the more than 20,000 feet of fluviatile, paludal, and lacustrine strata that accumulated in the areas of greatest subsidence during this period.

Basin subsidence and mountain uplift had virtually ceased by the end of Early Eocene time. Clastic debris eroded from the mountains, augmented by volcanic debris, continued to fill the basin during the later stages of Tertiary time. Near the end of the period the entire region was elevated several thousand feet above its previous level, and the present cycle of erosion was initiated. Normal faulting, perhaps concomitant with regional uplift, locally modified the older structural features.

3/6/80

STRATIGRAPHY AND TECTONIC FEATURES OF PARADOX BASIN, FOUR CORNERS AREA

H.R. Ohlen and L. B. Mc Intyre, Shell Oil Company, Farmington, New Mexico

The Paradox is a northwest-southeast elongate structural and sedimentary basin, bounded on the east and northeast by the San Juan Mountains and the Uncompahgre Plateau of Colorado, on the south by the Defiance Plateau of Arizona and New Mexico, and on the west by the San Rafael swell of Utah. Approximately 20,000 feet of sediments are preserved within the Permo-Pennsylvanian basin. Surface exposures are mostly sediments of the Mesozoic System, represented by a few thousand feet of clastics. The La Plata, Carrizo, Abajo, and La Sal Mountains are Tertiary intrusives within the Paradox basin.

The thin early Paleozoic sediments transgressed easterly onto the northeast-southwest trending transcontinental arch with the Cambrian sandstones separated from the Devonian-Mississippian shelf carbonates by an Ordo-Silurian hiatus. Exposure of the Mississippian carbonates resulted in a karst-like regolith, the Molas Formation of Atokan and/or earliest Pennsylvanian age. Tectonic influence then gave a northwest-southeast structural grain to the Paradox basin, in which Pennsylvanian cyclic or rhythmic shelf carbonates, sapropelic " black shale" dolomites, evaporites and arkose-redbeds were deposited. Approximately 2,000 feet of shelf carbonates were deposited on edges of the basin where highlands were not present. Evaporites were deposited in the center and distal northwest end of the basin and reach a present thickness of 10,000 + feet as the result of salt flowage in the cores of the intrusive salt anticline. The many thin sapropelic dolomites are widespread throughout the basin and are the "timemarkers" used for correlation. Clastics shed from the Uncompahgre and San Luis uplifts resulted in several thousand feet of arkose and redbeds ("Pennsylvanian" Cutler Formation) being deposited on the northeast and east flanks of the basin. These highlands persisted through Permian time and similar clastic deposition ("Permian Cutler Formation") continued, depositing a thick wedge of arkose near the Uncompahgre front, which thins to the southwest to 2,000 feet of continental redbeds and eolian and marginal marine sandstones of the Cedar Mesa, Organ Rock, and De Chelly Formations.

Approximately 3,000 Feet of Jurassic-Triassic continental shales and eolian sandstones were depsoited over Permo-Pennsylvanian basin and highlands. These sediments are expsoed over the western two-thirds of the Paradox basin. Upper Cretaceous marine clastics, exposed at the eastern edge of the basin, are equivalent to a continental facies in western Utah.

The primary structural grain of the Paradox basin is a northwest-southeast lineatio paralleling the Uncompahgre uplift of Permo-Pennsylvanian time. This alignment is illustrated best by the several salt anticlines. Laramide tectonic both rejuvenated the older trend and developed new structural lineations such as the north-south striking Monument upwarp. Regional uplift, coupled with the development of peripheral Tertiary basins has placed the Permo-Pennsylvanian Paradox basin in a high structural and topographic position incised by deep superimposed drainage.

TECTONIC DEVELOPMENT OF IDAHO-WYOMING THRUST

Frank C. Armstrong and Steven S. Oriel, U.S. Geological Survey, Federal Center, Denver, Colorado

Three stages are evident in the tectonic development of southeastern Idaho and western Wyoming. First, the changing patterns of tectonic elements during deposition; second, development of northward-trending folds and thrust faults; and third, development of block faults that produced horst ranges and graben valleys.

During Paleozoic time about 50,000 feet of marine sediments, mostly limestone and dolomite, were deposited in a miogeosyncline and about 6,000 feet of mixed marine sediments were deposited on the shelf to the east. Detritus came from both east and west recurrently from Cambrian time on. Starting in Mississippian time, the belt between shelf and miogeosyncline, where thickness increase markedly, shifted progressively eastward.

During Mesozoic time about 35,000 feet of marine and continental sediments were deposited in the western part of the region and about 15,000 in the eastern, terrestrial deposits becoming increasingly dominant. Western positive areas became the chief source of detritus. The belt of maximum thickening and the site of maximum deposition shifted progressively eastward; maximum thickness of succeeding geologic systems are not superposed. In Late Triassic a belt to the west rose and the miogeosyncline started to break up. As Mesozoic time progressed the western high spread eastward, until by the end of the Jurassic the miogeosyncline gave way to intracratonic geosynclinal basins that received thick deposits, particularly in Cretaceous time. Cenozoic sedimentary rocks are products of orogeny in the region.

The second stage which overlapped the first, produced folds overturned to the east and thrust faults dipping gently west in a zone, convex to the east, 200 miles long and 60 miles wide. Stratigraphic throw on many larger faults is about 20,000 feet; horizontal displacement is at least 10 to 15 miles. Lack of metamorphism and mylonite along the faults is striking. From west to east, the thrust faults cut progressively younger beds, have progressively younger rocks in their upper plates, and are estimated to be successively younger. Thrusting started in the west in latest Jurassic and ended in the east perhaps as late as Early Eocene time; ditritus shed from emergent upper plates is preserved in coarse terrestrial strata of corresponding ages.

West of the thrust belt is a northwestward-trending area underlain mostly by lower Paleozoic rocks and flanked on east and west by upper Paleozoic and Mesozoic rocks. Scattered pieces of eastward-dipping thrust faults have been reported west of the older rocks. This central area of old rocks has been interpreted as (1) part of a large continuous thrust sheet moved scores of miles from the west, or (2) an uplifted segment of the earth's crust from which thrust sheets to the east and west were derived. Both interpretations have defects; relative thrust ages are difficult to explain under the first; a large positive gravity anomaly, expectable under the second is apparently absent.

Block faulting, the third stage of tectonic development, started in Eocene time. Faulting has continued to the Recent, as indicated by broken alluvial fans, displaced basalt flows less than 27,000 years old, and earthquakes. North-trending and east-trending fault sets are recognized. Old east-trending steep faults in the Bear River Range may be tear faults genetically related to thrusting. Movement along many faults has been recurrent. Patches of coarse Tertiary gravel on the flanks and crests of ranges, for which there is no provenance with present topography, may record reversed vertical movement along some north-trending faults. Present topographic relief of basins and ranges is tectonic.

3/6/80

GEOLOGIC HISTORY OF ROCKY MOUNTAIN REGION

John D. Haun and Harry C. Kent, Department of Geology, Colorado School of Mines, Golden, Colorado

In late Precambrian time, sedimentary rocks were deposited in a developing geosyncline in the Cordilleran region. Eastward extensions of this geosynclinal sea occupied parts of the Rocky Mountain Region. Following gentle deformation and erosion the sea spread eastward during Cambrian and Ordovician Periods.

Discontinuous Ordivician, Silurian, and Early Devonian rocks indicate short intervals of marine invasion interrupted by periods of erosion. A major invasion of the sea over the craton is recorded by the onlap of Devonian and Mississippian carbonates and Devonian evaporites which rest on rocks ranging in age from Precambrian to Early Devonian.

The pattern of widespread shallow seas of the Mississippian Period was interrupted in the Pennsylvanian and Permian Periods by significant tectonic activity (Ancestral Rockies). Portions of the uplifts remained positive until Triassic or Jurassic Periods and supplied coarse clastics to adjacent areas within late Paleozoic basins. At greater distances from land areas, sandstones, redbeds, evaporites, and carbonates accumulated.

Marine Triassic sediments were deposited in southeastern Idaho and adjacent areas. Triassic and Early Jurassic continental deposits accumualted throughout much of the region.

A series of Jurassic marine invasions from the Arctic initiated another major sequence of events. The boreal sea moved southward into the northwestern and western parts of the region in Middle Jurassic and successive transgressions reached as far southeast as northern Colorado by Late Jurassic. After withdrawal of the Jurassic sea, the pattern of overlap was continued by deposition of nonmarine Jurassic and Cretaceous sediments.

In the Early Cretaceous a sea again inveded from the north and in late Early Cretaceous joined a southern sea forming a seaway which persisted during the remainder of the Period. During Early Cretaceous, clasic sediments were derived from the craton to the east and from the Cordilleran region to the west. In Late Cretaceous, the western source area predominated.

The present tectonic framework was initiated during the Late Cretaceous and early Tertiary with the development of uplifts and intermontane basins (Laramide Orogeny) accompanied by the emplacement of the Idaho batholith and associated intrusions. Extensive thrust faulting occurred in the western part of the region. Lacustrine and fluviatile sediments, derived from surrounding uplifts, were deposited within the intermontane basins.

Volcanic activity was moderately important to the west during the Cretaceous Period, but igneous intrusion and volcanic activity became widespread throughout the Rockies in the Tertiary.

The present drainage system was largely developed as the intermontane basins filled. Subsequent stream erosion, accompanied by Pleistocene glaciation and regional uplift, resulted in the shaping of the present topography.

GENERALIZED HISTORY OF SEDIMENTATION AND STRUCTURAL DEVELOPMENTS OF BIG HORN BASIN

Leonard E. Thomas, Marathon Oil Company, Casper, Wyoming

The Big Horn basin of northwest Wyoming is primarily a Laramide structural basin. The area has been a portion of larger sedimentary basins throughout most of geologic history.

The basin is located on the eastern shelf of the Cordilleran geosyncline, east of the hinge line separating the shelf from the former deep parts of the syncline.

Local structural deformation on the sites of several Laramide anticlines in the basin is suggested by slight thinning noticeable in strata of Ordivician age. Local structural influence upon the present-day basin, however, is not evident until at least as late as the beginning of Upper Cretaceous.

During the pre-Laramide eras, periods of regional movements indicate a "sea-saw" action with repeated northerly tilting, deposition, emergence and erosion which resulted in truncation of the Ordivician, Devonian and Mississippian sediments from north to south. There is a complete absence of Silurian sediments.

During Pennsylvanian, Permian and Triassic, the area of the present-day basin underwent southerly tilting, deposition, erosion and truncation that resulted in the formations thinning from south to north.

Jurassic and Lower Cretaceous formations show an increase in thickness from south to north. They also show the development of a low-relief structural arch that appears to be the buried, northwest-plunging nose of the Casper arch and Laramide Range in south and central Wyoming.

By the beginning of Upper Cretaceous time the embryo of a structural basin may have been present which affected some of the basal sands of the Frontier Formation.

Later transgressions and regressions of the Upper Cretaceous seas continued until the Laramide Orogeny came into strong evidence at the beginning of Fort Union time. The period of intense movement continued into Eocene with thrust faulting followed by deposition and partial erosion of volcanoes on the western margin of the basin.

This movement resulted in peripheral mountain building, pronounced unconformities at the margins of the basin, the development of conglomerates in the Tertiory beds, and the development of the intense anticlinal folds preserved today.

3/6/80

DENVER BASIN

C. A. Martin, Continental Oil Company, Durango, Colorado

The Denver Basin is one of the largest structural basins in the Rocky Mountain area and extends across portions of Colorado, Nebraska, and Wyoming. The basin is typically asymmetrical with its axis paralleling and close to the Front Range. The deepest portion of the basin lies near Denver, where more than 12,500 feet of sediments are present.

Recently discovered outcrops suggest that during pre-Pennsylvanian time, the Denver basin was a normal marine shelf receiving sediments from early Paleozoic seas. Post depositional uplift along Siouxia exposed this area to deep erosion removing nearly all the early sediments.

In Early Pennsylvanian time, transgressive seas entered the Denver basin area from the Anadarko basin, depositing a predominantly clastic terrane. Near the end of the Atoka, the first major pulses of the Ancestral Rockies occurred. This uplift reached maximum proportions during the Des Moines. Clastic material eroded from the uplifted mountains was deposited contemporaneously with marine sediments deposited in the expanding seaway. This expansion reached its maximum during the Missourian and was followed by a slight Virgil regression. Continued Permian regression left a full suite of environments and facies ranging from normal marine through an evaporitic sequence to terrestrial deposits. Late Permian and Triassic sediments indicate the Ancestral Rockies were weakly positive and supplied sediments to a shallow hypersaline sea.. Non-depositional conditions persisted from Upper Triassic through Lower Jurassic, and into Mid-Jurassic time. During Middle and Upper Jurassic time, transgressive seas fluctuated across the basin from the northwest. As these seas regressed at the close of the Jurassic, a broad inland flood plain developed.

Early Cretaceous seas inundated the area from the north and south reworking earlier sediments and obscuring the Jurassic-Cretaceous boundary. Initial basin-forming movements occured at this time. Fluviatile material from exposed areas to the east and northeast developed a complex deltaic pattern as it interfingered with marine sediments basinward. Deltaic deposits also extended into the area from the southwest merging with sediments from the east. Another cycle of transgression and regression developed similar depositional patterns. It is within these two Early Cretaceous sedimentary cycles that significant reserves of oil and gas have been discovered. In the Upper Cretaceous, a major transgression joined the northern and southern seas into a large seaway crossing the downwarping basin, and lapping against the uplifted Front Range. This Laramide tectonic activity reached its peak during the Eocene with the basin acquiring its present configuration.

GEOLOGICAL HISTORY OF CENTRAL AND SOUTH-CENTRAL MONTANA

E. Earl Norwood, Continental Oil Company, Billings, Montana

Central Montana has had a complex structural and sedimentary history, especially the area of today's Central Montana uplift.

Precambrian and Cambrian subsidence allowed deposition of 1300 + feet of clastics in an east-west trending trough roughly coincident with the present-day uplift.

Pre-Devonian uplift and erosion followed stable depositional conditions during Ordovician time. Ordovician has been eroded from the western one-half of the study area. Silurian is absent from the entire study area.

Central Montana uplift area remained high during Lower, Middle and part of Upper Devonian time. Upper Devonian rocks lap onto the uplift and uppermost Devonian finally covered the area. Pre-Mississippian uplift removed these carbonates and shales completely from a large area of the uplift.

Mississippian system is comprised of the carbonate evaporite Madison Group and clastic Big Snowy Group. Stable conditions prevailed through most of Madison deposition, but central Montana began to subside in Late Madison time. The Big Snowy Group was restricted by continued subsidence which downwarped central Montana into a synclinorium.

Early Pennsylvanian streams draining eastern Montana cut valleys in the Central Montana trough. These valleys were filled as the streams attained old age primarily with Big Snowy Group derived sands and shales. This stream-channel deposit, the Lower Tyler Formation, contains the major reservoirs of central Montana, Middle and Late Pennsylvanian sediments covered central Montana; but all of the Late Pennsylvanian was eroded pre-Jurassic time. Pre-Jurassic folding accentuated Mississippian structure.

Jurassic saw uplift in the Belt Mountain area to the west. Jurassic laps onto this high and thickens eastward as well as in the trough area.

Lower Cretaceous deposition was controlled by uplift to the south and thickens from south to north.

The Laramide revolution upwarped the old trough into the Central Montana uplift and also generally folded the old synclines into anticlines as at the Sumatra trend. Isostatic adjustment at basement fault blocks was the force behind the down up down up movements of central Montana.

3/71

CLIMATOLOGICAL-GEOLOGICAL CORRELATIONS BETWEEN COLORADO-PLATEAU AND GREAT BASIN

TIME SPANS	COLORADO PLATEAU	GREAT BASIN
HISTORICAL	Onset of extensive gullying about 1850. Steady removal of valley fill and some denudation in higher elevations.	Fluctuations in levels of existing lakes. Much destruction of vegetation and soil cover under intensive grazing by domestic animals.
RECENT & NEAR-RECENT	Previously deeply eroded tributary canyons alluviated, re-excavated and re-filled several times. Human occupancy begins about 15,000± BC climates generally oscillating but arid	Fluctuations of lake levels very surficial erosion and deposition Gradual cooling and drying as shown by sucession of human cultural remains found chiefly in caves.
MIDDLE AND LATE PLEISTOCENE	Sparce sedimentary record includes glacial deposits in the Lasal Mountains. Minor modification of land forms drainage probably integrated.	Extensive surficial deposits and land forms associated with lakes of varied size. Some over 1000' deep. Fair paleontological and geosol record, numerour wet-dry cycles.
LATE PLIOCENE AND EARLY PLEISTOCENE	Practically no dated deposits. Probably extensive bedrock erosion by secondary and tertiary streams.	Probably considerable volume of sediment laid down in deeper parts of trenches and valleys. Practically no datable surface deposits. No positive evidence of large lakes. Probably arid.
EARLY AND MIDDLE PLIOCENE	Scanty sedimentary record includes Bidahochi Formation (Ariz.) Drainage patterns mostly unknown. General disintegration of weathered regolith with increasing aridity.	Sparce paleontological evidence and few proven sedimentary deposits. General denudation of ranges and building of alluvial fans with increasing aridity.
OLIGOCENE-MIOCENE	Scanty record, no extensive deposits. Drainage pattern unknown. Laccoliths intruded. Green River captured.	Fair sedimentary record, also much igneous material, general subsidence. Deciduous hardwood forests, humid climate, deep weathering.
PALEOCENE-EOCENE	Great downsinking in Uinta-Piceance Basins produce extensive lake systems, elevation 1000-2000 feet. Humid climate; external drainage but routes unknown.	Relatively higher elevation than surrounding areas. Sparce Paleontological and sedimentary record, drainage both eastward and westward.

2/28/80

Mesozoic - Cenozoic Stratigraphy

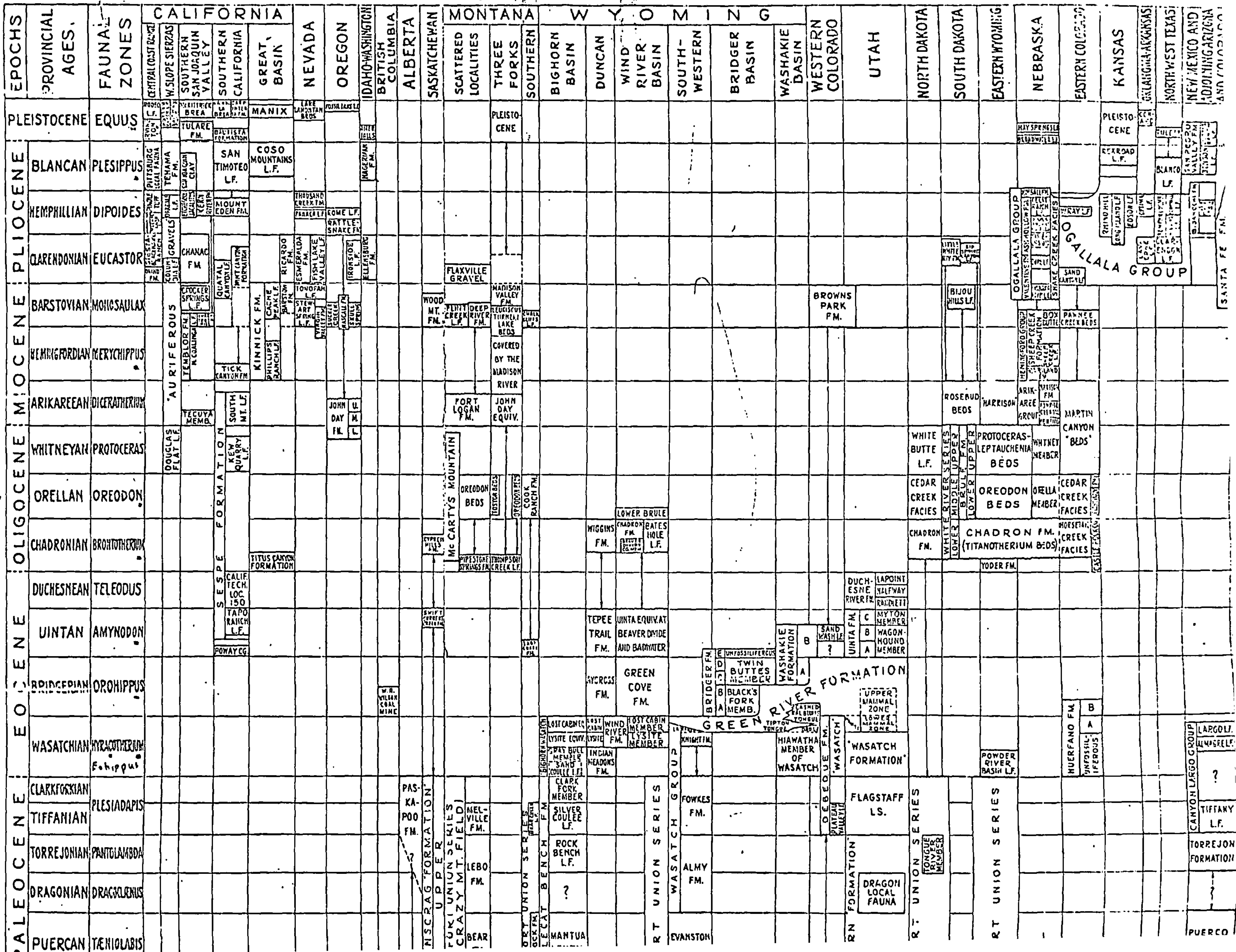
CENOZOIC ASSIGNMENT

1 Write a brief summary of the stratigraphy of the major Tertiary basins of the Rocky Mountains. No less than six and no more than ten please. Cover such topics as type, thickness and derivation of sediments. Show formal stratigraphy by simplified columnar sections grouped on one page. Try to condense your descriptions to about 1/2 page per basin.

W. Lee Stokes '80

N.A. Mammalian Zones + Faunas

2/28/80



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LAKE BONNEVILLE

The story of Lake Bonneville is pieced together by the geologists using the evidence at hand. There is still much to be learned about this subject. The most noticeable evidence of the lake is the terraces on the flanks of the mountains; however the sequence of sedimentary deposits formed while the lake was present tells us the order in which the terraces were formed, and consequently these tell us more of the actual history. Diagram I shows the relations of the terraces of the various levels of the lake and the types of sediments deposited. This is only a sketch, so the vertical and horizontal scales are not accurate.

Currents in Lake Bonneville created many shore-line features such as sand and gravel bars. Examples of these are Point of the Mountain, Stockton Bar (south of Tooele), and the spit at the south end of the Hogup Mountains.

The general landscape has been much as it is at present since before Bonneville time. However such things as the lake terraces and deposits, some faults, mudflows, and alluvial fans present in the current landscape have naturally modified the earlier landscape. During Bonneville time there were of course fish living in the Lake, and such mammals as mammoth, musk ox, camel, horse, deer, wild cat, and mountain sheep were living along its shores.

You are perhaps aware that Lake Bonneville was not a constant or permanent feature. Instead, the lake level varied greatly, as shown on the second diagram enclosed. Most of the fluctuations in the level of the lake were controlled by the climate of the time--that is by the glacial and interglacial intervals. However, it has been postulated that the sudden rise in the Lake to the Bonneville Level (see Diagram II) may have been due to a sudden change in the course of the Bear River. Thus, prior to Bonneville time the Bear River flowed into the Snake River

of the river and diverted it into the lake: as this water reached the lake there was a sudden 28% increase in inflow which resulted in the Bonneville level. Incidentally this abrupt change in the course of the Bear River is noticeable on a map and might be of interest to your fellow students.

A few random statistics, some of which might be of interest, follow:

Maximum elevation of the Lake was 5,200 + feet (Bonneville Level)

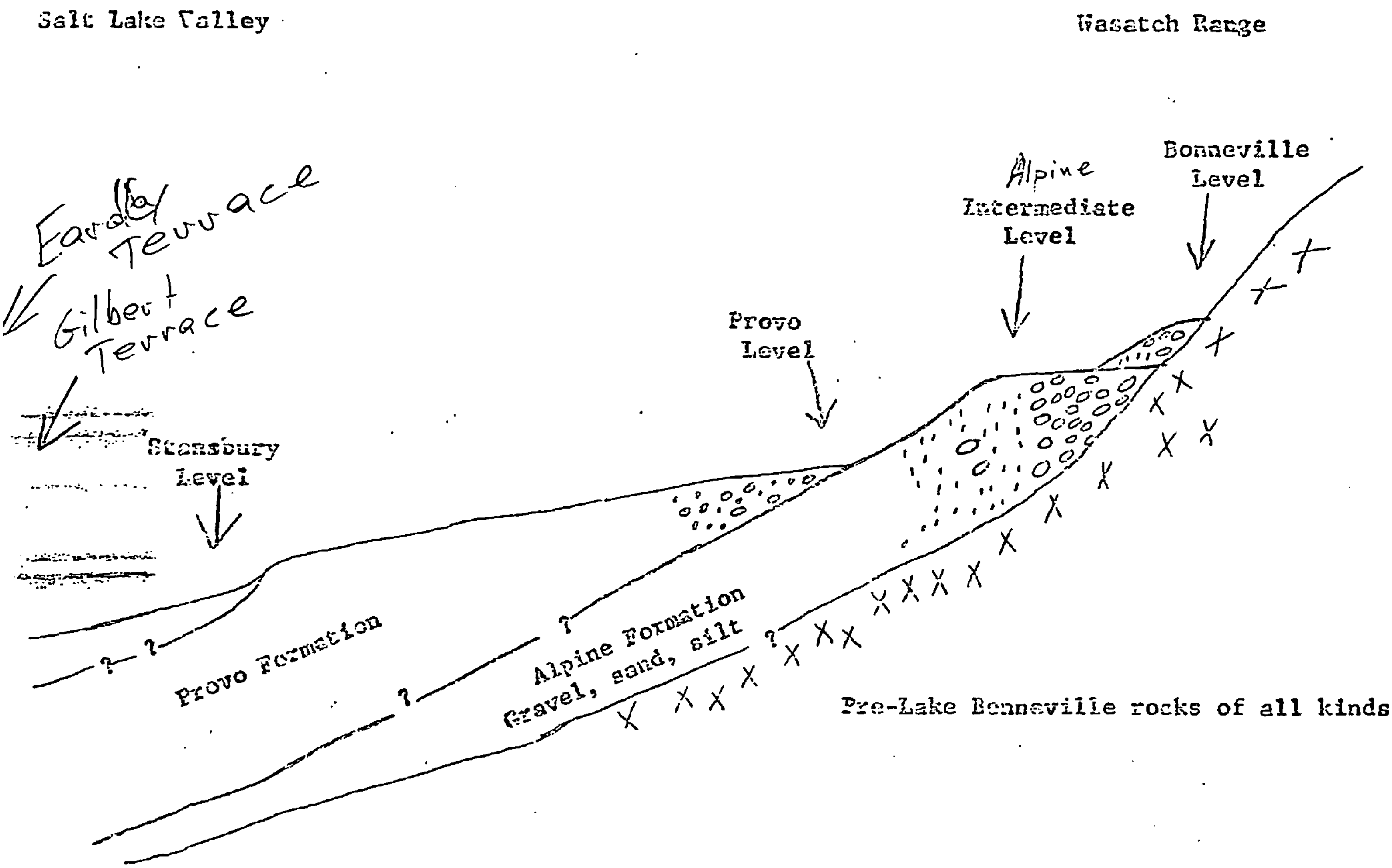
Greatest depth was probably about 1,000 feet.

When the lake level was above 4,800 ft, the lake spilled out through Red Rock Pass near Preston, Idaho; when the lake level was below 4,800 ft it probably did not have any outlet but was limited by evaporation.

Present Great Salt Lake is, of course, a small remnant of the former Lake Bonneville, as well as Sevier and Utah Lakes.

At its maximum level Lake Bonneville had an area of 19,940 square miles.

DIAGRAM I

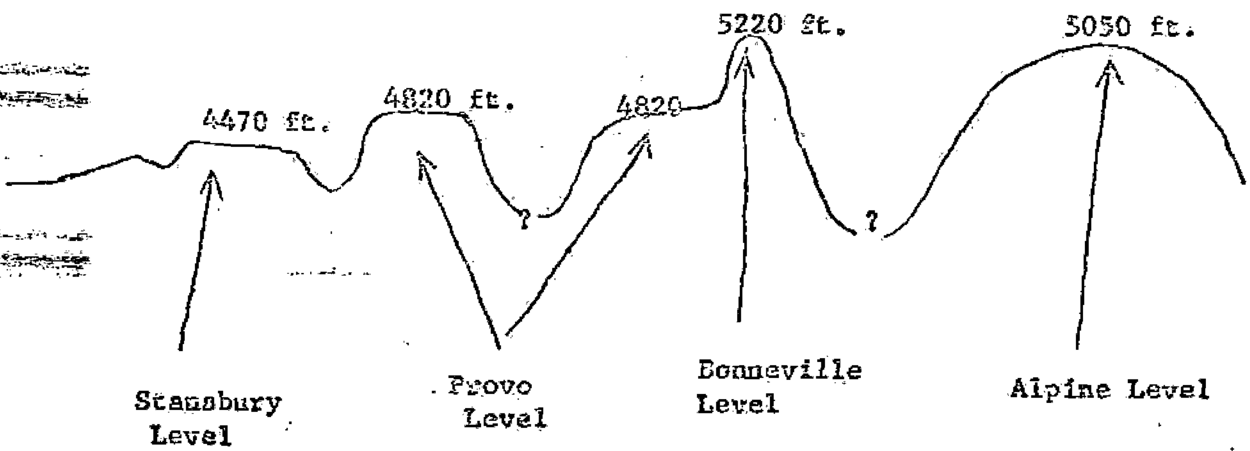


Stratigraphic sequence is:
Alpine Formation
Bonneville Formation
Provo Formation
Late Lake Sediments

Gravel, sand, and silt

THIS IS A SKETCH: HORIZONTAL & VERTICAL SCALES ARE NOT ACCURATE.

SKETCH SHOWING VARIOUS LEVELS OF LAKE BONNEVILLE



Note: Total time interval involved and absolute duration of each level are not known.

SUBJ
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Paleo

Paleozoic Stratigraphy

1/5/'78

John Atwood

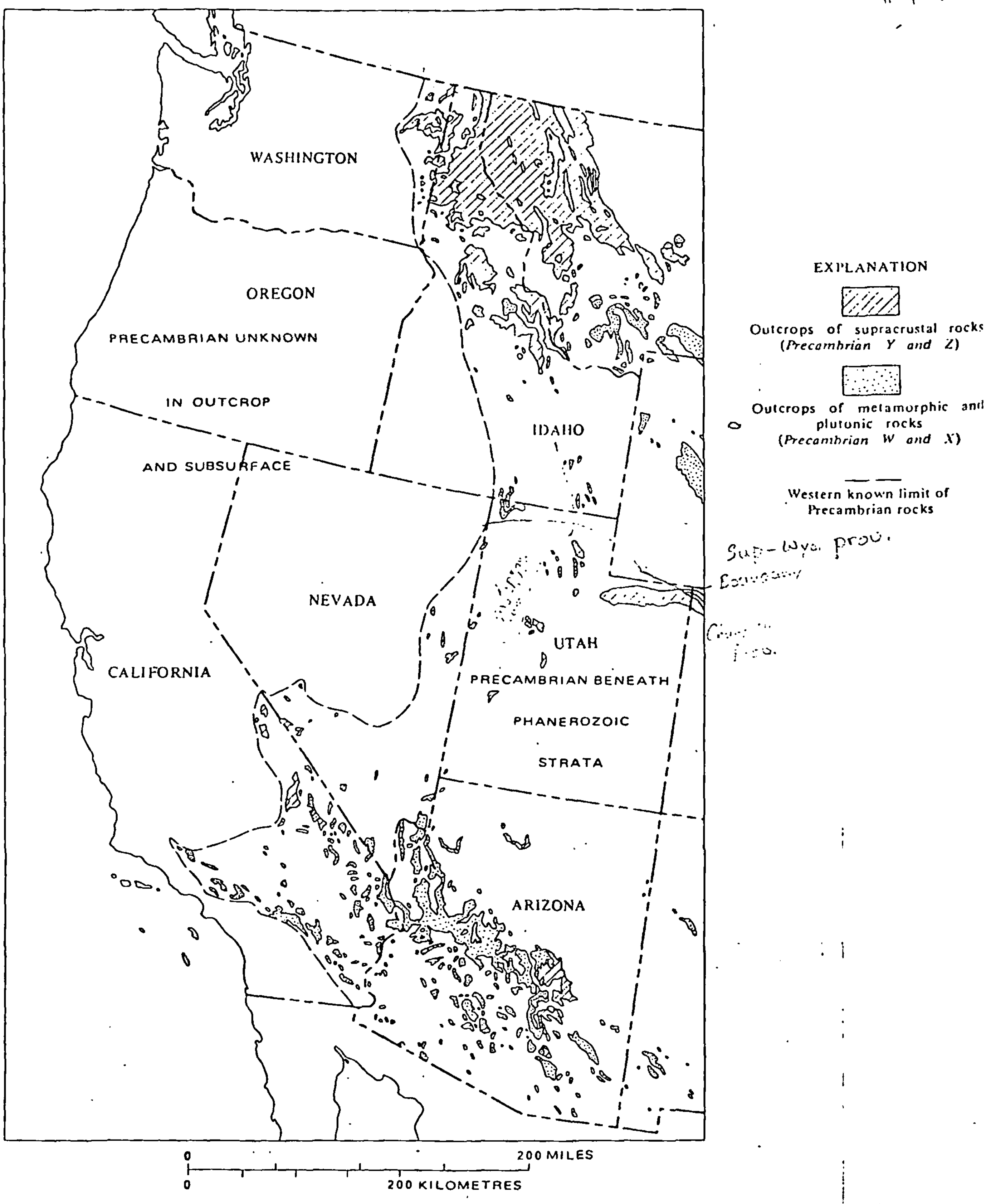

FIGURE 30.—Map of western United States, showing western known extent of Precambrian rocks.

1/5/79

Geology 676

Generalizations

The Precambrian Framework for Paleozoic Stratigraphy of the Western Interior

I. Two lithologic types dominate: a) predominantly gneiss of great thickness, base unknown, b) predominantly quartzite with minor clastic meta sediments.

1. The gneissic or crystalline "basement" appears to underlie the entire region except for that part lying north and west of a line running approximately northeasterly from Lake Tahoe (bend in Nevada-California line) to Yellowstone Park and north-northwest across Montana. This line appears to mark the edge of ancient continental crust, beyond it there has been extensive deposition on oceanic crust.

   a) Age of basement in most of Arizona, New Mexico, Utah, & Colorado is primarily 1.75 m.y. with overprint (metamorphic reset) of 1.45 m.y. Age of basement in most of Wyoming (primary) 2.7 m.y. Age of basement in most of Montana (except southwest) is 1.7 m.y.

   b) The gneissic terrane is apparently a deeply eroded orogenic belt that constitutes an integral part of the Transcontinental Arch, which trends from the Mojave Desert into Labrador, Canada. On the geosynclinal theory the rocks are of shallow water mainly sedimentary origin. The "grain" is distinctly northeasterly.

2. The quartzitic series is found chiefly in a northerly trending belt from northern Arizona into Canada. From the main belt which is on the order of 200 to 500 miles wide there are several narrow offshoots that trend east-west. These are the Belt Trough with at least 60,000 feet of sediment dated at 900-1400 m.y., the Uinta Trough with sediments dated at 600 to 1,700 m.y., and the Central Arizona Trough with sediments dated at 1,100-1,450 m.y.

   The meta sediments of the quartzose sections are fine to coarse clastics, there is very little carbonate and almost no volcanic material. A very large portion of the Belt Series is argillite indicating relatively low-energy environments of long duration. On the other hand, the Uinta Group is coarser and shows more dynamic primary structures.

3. The gneissic terrane is separated from the quartzose sediments by a profound unconformity having generally local erosional relief of less than 500 feet. It is generally a very clean, clear-cut surface. The cutting of this surface occupied on the order of .5 billion years or more. What happened to the great bulk of sediments eroded during this interval is problematical. Although not exactly appropriate the name Epi-Archean Unconformity has been used.

4. Although there was definitely a North American Craton or proto-continent at the time the gneissic material accumulated and was metamorphosed the configuration, orientation, and relation to other cratons is not fully understood. Majority opinion seems to be that there was one great land mass at this time. The dominant grain of Transcontinental Arch as well as of the exposed Precambrian of the Canadian Shield is northeasterly, a fact not clearly explained by global tectonics.

5. A great reorganization of tectonic patterns, probably reflecting fundamental changes in global stress relations took place in the interval from about 1.4 to .9 b.y. At this time the western margin of North America became oriented approximately north-south, or probably more precisely in the direction that it has maintained ever since. Strike of the quartzite series sedimentation is at a high angle to that of the "grain" of the gneissic terrain. Primary structures indicate westward flowing drainage and eastern cratonic source for the clastic constituents. This evidence strongly suggests a western marine repository at this time.

6. No clear-cut evidence for an ancestral Pacific Ocean is known for the earlier Precambrian and it is not even safe to say that the western seaway of the late Precambrian was the Pacific Ocean. It has been suggested that a grand continental separation (eastern Asia from western North America?) commenced during the late phases of deposition of the quartzite series. If this is so the migration of Asia has been extensive and investigations into similarities of the opposing coasts should be undertaken. In any event the existence of a wide Pacific Ocean by the Cambrian seems undeniable but its width is unknown.

7. Peculiar sediment-filled troughs such as the Uinta Mountains which penetrate at right angles into the margins of cratons are called aulacogens (Russian discovery and nomenclature). Research on such features is getting under way.

W. Lee Stokes, 1974

1/5/79

PRECAMBRIAN
(with emphasis on Utah)

Classification

Precambrian rocks of the Western Interior may be classified in a number of ways:

A. Metamorphic grade:

Archean or Archaeozoic - includes all crystalline, highly metamorphosed rocks. No base exposed. Tertiary intrusives were once called Archean. e.g. Cottonwood Stock.

Proterozoic - includes the less metamorphosed Precambrian sediments, separated from the Archean by a great erosional unconformity. Also referred to as Algonkian.

B. Age:

1. U. S. Geological Survey system, (see Subdivision of the Precambrian: an interim scheme to be used by the U.S. Geological Survey, 1971)

Designation	Age bracket (millions of years)
Precambrian Z	Base of Cambrian to 800
Precambrian Y	800 - 1,600
Precambrian X	1,600 - 2,500
Precambrian W	2,500 and older

2. General scheme used in many areas of the world - specific dates vary.

Late Precambrian	1800 m.y. to 570 m.y.
Middle Precambrian	2600 m.y. to 1800 m.y.
Early Precambrian	earliest rocks to 2600 m.y.

3. Canadian (they have much more to deal with)

Proterozoic	2390 to 570 m.y.
Archean	earliest rocks to 2390 m.y.

C. Provinces or regions:

1. Wyoming - Superior Province: A belt trending from northern Utah toward and probably originally connected with the Lake Superior area. Includes northern edge of Utah, most of Wyoming, eastern Idaho, eastern Montana and the northwestern corner of Colorado. Chief rock producing episode 2.5 - 2.7 b.y.

2. Churchill Province: Encloses and wraps around the Wyoming-Superior Province. Wide band on all sides except across southern Idaho. Includes western Montana, most of western Idaho, most of Utah except northern edge, all of Nevada except northwest quarter, practically all of Arizona, northwest half of New Mexico most of Colorado. Chief rock production 1.6 - 1.8 b.y.

3. Utah-Nevada Province: Condie[1] has proposed adding the Utah-Nevada Province to include late Precambrian sedimentary rocks deposited unconformably on the igneous-metamorphic complexes of the Churchill and Wyoming provinces. He recognizes five subprovinces: the Big Cottonwood, Uinta, Southern Nevada, Grand Canyon and Pahrump (Condie, p. 83). The age limits set by Condie for rocks of the Utah-Nevada Province are 600 to 1000 million years.

4. Other Post-Churchill Provinces: All other Precambrian rocks of the western interior were included by Condie in the Belt Province (1.0-1.3 b.y.), the Apache Province (1.2-1.4 b.y.) and the Sioux Province (1.2-1.7 b.y.).

5. Critique: Classification of the Precambrian rocks is progressing but much remains to be done. The time-honored division into Archeozoic (metamorphosed) and Proterozoic (non-metamorphosed) looses practically all validity when applied to such areas as the Pacific margin where the Mesozoic is the metamorphosed basement. However, when accompanied by age-dates the names may still be used as they are in Australia and Canada.

Although there may be much practical value in recognizing great orogenic provinces the outlining and definition of these is only beginning. Among the serious problems of this method are: a) difficulties of setting exact boundaries between provinces; b) upsetting of radiometric dates in any older province by subsequent events, particularily during the creation of cross-cutting mountain ranges such as the Rockies,; c) overlapping or penetration of basins of non-metamorphosed sediments onto and across older basement terranes. These do not represent full-scale orogenic events and the dates they provide may be from sedimentary rocks. The Belt, Purcell, Windemere, Uinta, Apache, Grand Canyon and otherlate Precambrian units represent an unusually quite time. Can these be lumped in provinces or subprovinces equivalent in any way with the older crystalline terranes?

Schemes based on authentic radiometric dates give historical truth and in the end have the greatest weight of any criteria. But such dates are few and far between and for many thick sequences of rock appear remote or improbable.

Consider what has happened recently in the study of the Precambrian of the Western Interior. Condie's scheme falls under suspicion with the recovery of a solitary radiometric date from the upper part of the Uinta Mountain Group.[2] This date of 952±5 m.h. seems to make all or most of the Big Cottonwood and Uinta subprovince rocks too old for the Utah-Nevada Province. Instead they may now be in the age range

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[1] Condie, Kent C., 1969, Geologic Evolution of the Precambrian Rocks in Northern Utah and Adjacent Areas, in Guidebook of Northern Utah, prepared for the Rocky Mountain Section, Geological Society of America. p. 75.

[2] Crittenden, M.D., Jr. and Peterman, Z.E., 1975, Provisional Rb/Sr Age of the Precambrian Uinta Mountain Group, Northeastern Utah: Utah Geology, vol. 2, no. 1, p. 75-77.

of the Churchill Province. Here they are out of place because they are mostly unmetamorphosed. Can the remaining Post-Churchill rocks of other subprovinces be lumped in any meaningful way into Condie's provinces or subprovinces? What is to be done with these rocks?

It is proposed for purposes of brief disscussion to subdivide the Precambrian rocks of the Western Interior into three units based chiefly on age.

Oldest: All crystalline, basement rocks the metamorphism of which took place before 1.6 b.y.. This would include practically all of the Wyoming and Churchill Provinces.
Intermediate: All sedimentary and metasedimentary rocks between the top of the crystalline basement and the horizon of the latest Precambrian glacial event usually marked by diamictites. May include metamorphic rocks younger than 1.6 b.y. Age range approximately 1.0 to 0.7 b.y..
Youngest: All sedimentary and metasedimentary rocks between the glacial deposits and the beginning of the Cambrian. May include metamorphic rocks younger than 1.6 b.y. Age range approximately 700 to 575 m.y.

The older crystalline rocks of the Western Interior (Wyoming Province) appear at the surface in the great mountain uplifts of Wyoming including the Wind River, Big Horn, Beartooth, Owl Creek, Teton and Sweetwater Ranges. They also appear in the Black Hills. Elsewhere in the basins the corresponding rocks are at great depths, over 2500 feet below sea level in the northeast Green River Basin.

About 80 percent of the Wyoming Province consists of granitic rocks and assoicated gneiss, migmatite and agmatite. "Of the remainder, amphibolite and (Chiefly orthoamphibolite) and matic igneous rocks compose about 10 percent and metagraywackes, metaquartzites, metavolcanics, iron-formation, and alpine ultramatics collectively compose about 10 percent. All the metasedimentary-metavolcanic terranes are engulfed, metamorphosed and in some cases partially granitized by surrounding granitic rocks. The metamorphic grade of the engulfed terranes varies from the greenschist to the amphibolite, or, rarely to the granulite facies."(Condie, p. 75).

Younger crystalline rocks of the Western Interior (Churchill Province mainly) appear at the surface in the Farmington Mountains and Antelope Island (northern Utah); Uncompahgre Uplift (west-central Colorado); many uplifts in central Colorado including the Front, San Juan, Sawatch Park and Medicine Bow Ranges; in the Sangre de Cristo, Sanhia, and San Andreas Ranges of New Mexico and dozens of scattered outcrops in Arizona.

Rock types are chiefly gneiss, schist, migmatite, and amphibolite, with granitic intrusions and pegmatite veins. The Farmington Canyon Complex is described by Condie (1969, p. 80) as being composed chiefly of granitic gneiss and migmatites (~90 percent) with lesser amounts of amphibolite and pegmatite (~5 percent). The gneiss and migmatities, in turn are composed of quartz, microcline, sodic plagioclase, and biotite.

Large granitic plutons are found intruding the Churchill rocks. The Vernal Mesa Quartz Monzonite is an igneous body of batholithic proportions bisected by the Utah-Colorado line at the west end of the exposed Uncompahgre Uplift. It is dated at 1.443±22 m.y.. A preliminary guess based on inspection of mapped exposures of Churchill rocks south of the Wyoming Province is that granitic rocks increase southward from northern Utah to central Arizona.

The late Precambrian history of the Cordilleran region was notably uneventful but there is fortunately one episode of regional, even worldwide importance to break the monotany. This was an ice age which left scattered effects from Alaska to California. Chief glacial action is estimated at about 750 to 850 m.y.. Evidence of this climatic event is tillite (called diamictite by most workers). Where this distinctive material is found it serves to divide the Upper Precambrian of western North America into two series:

"The older series is younger than 1450 to 1250 m.y. in most areas. It includes the following partial temporal equivalents: lower Tindar Group, Alaska; Purcell Series, British Columbia; Belt Supergroup, Montana, Idaho, and Washington: Big Cottonwood Formation, Utah; lower Pahrump Group, southern California; Grand Canyon Series, northern Arizona; Apache Group and Troy Quartzite, southern Arizona.

"The younger series commonly begins at the base with a thick diamictite, and may be separated from the underlying series by a marked unconformity. Diamictites in the Great Basin include massive continental tillites, plus proximal and distal turbadite facies with or without ice-rafted clasts. Equivalent diamictites may be exposed in: the Rapitan Formation, Yukon and Northwest Territories; Toby Conglomerate of the Windemere Group, British Columbia; Pocatello Formation, Idaho; the Mineral Fork Tillite and Dutch Peak Tillite of Cohenour (1959), Utah; and the Kingston Peak Formation of southeastern California.

"Deposition of the younger series probably began 750 to 850 m.y. ago, and continued into the Cambrian in basinal reactions. Probable temporal equivalents within the postdiamictite sequence include the remainder of the Windermere Group of British Columbia and equivalents in northeastern Washington, and the Brigham Group and equivalents in the Great Basin and southeastern California." ("Regional Correlation of Upper Precambrian Strata in Western North America", M.D. Crittenden, John H. Stewart and C.A. Wallace. 24th Int. Geol. Congress, Montreal, 1972, sec. 1, p. 334.)

Later Precambrian rocks of the Cordilleran Province are only weakly metamorphosed and may be described as metaquartzites, siltites, argillites, dolomites and limestone. Where metamorphosed by Mesozoic events as in eastern Nevada and western Utah they are described as schists and phyllites. The Windemere and Belt series are will known for their silty fine-grained composition. Rocks of the Utah-Nevada subprovince are about half fine clastics and half coarse clastics. The well-known Grand Canyon Series lying between the Vishnu Schist and the Cambriean is described in terms of such ordinary unmetamorphosed rocks as shale, sandstone and dolmite. The late Precambrian of the San Juan

Mountains is described as somewhat altered but much less so than the older basement complex. The Apache Group of southeastern Arizona consists of the Pioneer Shale, the Dripping Springs Quartzite, and the Mescal linestone, no particular metamorphic effects are implied.

Regional gradients are becoming evident in the later Precambrian rocks. Considering certain areas as reference points for contrasts and comparisons:

(a) Thickness increases northerly and northwesterly from southeastern Arizona.

(b) Carbonate content increases southwesterly, northerly and southerly from northern Utah-southern Idaho.

(c) Coarseness decreases southwesterly, northerly and southerly from central Utah.

(d) Diabase intrusions decrease inall directions from southeastern Arizona, none in Utah-Idaho.

Tentative Tectonic implications:

(a) There was an ancestral Wasatch Line and Las Vegas Line by about .7 b.y. ago.

(b) There was an ancient highland of the Churchill Province undergoing erosion in what is now southwest U.S. Perhaps two high areas - one in central Colorado, the other in Mojave Desert. Could be considered part of the Transcontinental Arch. Needs a name?

(c) Geographic and tectonic edge of the continent was not far from Wasatch Line, prograding westward by deposition.

(d) Continental divide not greatly different from present location, drainage was westerly for most of the Western Interior.

Wm. Lee Stokes, 1976

1/5/79

Second Precambrian Assignment

On a single page show in diagrammatic form the World-wide correlation scheme proposed by Salop[1] for the Precambrian. Place next to it in proper position the current scheme of the U.S. Geological Survey[2]. In a third column, using lithologic symbols and names place the Precambrian section of the western United States.

1 Salop, L.J. 1977, "Precambrian of the Northern Hemisphere," Elsevier Pub. Co.

2. King, P.B. 1976, "Precambrian Geology of the United States" U.S.G.S. Professional Paper 902.

W. L. Stokes, 1979

1/12/79

Cambrian-Precambrian Boundary in the

Western Interior

Every student of geology knows that there was a Cambrian Period and that its beginning was one of the greatest landmarks of all time. As described in most textbooks the beginning of the Cambrian was marked by the sudden appearance of abundant fossils. In their innocent enthusiasm earlier stratigraphers even up to the time of C. D. Walcott found no problems with determining a base for the Cambrian. This was, as we now realize, traceable to peculiar accidents of earth and human history. Where geology grew up and the geologic column was constructed there were very few places were geologists really had to draw an actual accurate base to the Cambrian. Over most of North America, for example there is a great unconformity beneath the Paleozoic rocks. Rocks below the unconformity are unmistakably ancient,mostly metamorphosed and generally greatly deformed. Rocks above are mostly of Cambrian age, unmetamorphosed and little deformed. Anyone can see and map this contact. It is one of the tourist attractions in the Grand Canyon.

What is missing at the horizon of the unconformity is much of the Precambrian and also part of the Cambrian. For most of North America this eliminates the knotty problem of defining the true beginning of the Cambrian. The great unconformity was so profound, so unmistakable, and so reliable that geologists had to invent a plausable explanation for it. They called the unrepresented, missing part of the record the Lipalian Interval. The belief was that during this time the seas were far away in the ocean basins and that the continents were all fairly well out of water. Meanwhile, life appeared and evolved to fairly high levels in the remote depths of the ocean. Naturally, when the seas flooded inward at the end of the Lipalian interval they brought an abundance of fossils giving the appearance, in the rocks at least, of a sudden almost miraculous appearance of life.

But tougher ground lay ahead. As work progressed westward into the Basin and Range the great unconformity seemed to sink lower and lower until it could no longer be found. Whereas the oldest fossils upon the basement over most of the continent are Late Cambrian it was soon apparent that going westward the unconformity is overlain by Middle Cambrian, then by Early Cambrian and then by beds with a few fossils or none at all. One thing seems to hold true, however the crystalline basement is overlain by quartzite or sandstone representing residual reworked material of near-shore origin.

Faced with the problem of what to do with the thickening body of sediment between the crystalline basement and the earlies trilobites the following solutions to the boundary problem have been proposed.

(a) Include all sediments down to the top of the crystalline basement in the Cambrian.

(b) Include all sediment to the base of the first thick quartzite below the lowest trilobites in the Cambrian.

(c) Include only trilobite-bearing rocks in the Cambrian.

(d) Include beds with not only trilobites but also traces of trilobites in the Cambrian.

(e) Include all beds with evidence of metazoans in the Cambrian.

(f) Include only beds with trilobites or certain other specific organisms such as archaeocyathids and/or brachiopods.

(h) For sections with no fossils below a certain zone and no reliable radiometric dates include a thickness of rock in the Cambrian equal to that in a nearby section which does bear diagnostic fossils older than those of the section in question. In other words if the earliest fossils of a locality are middle Middle Cambrian and there is a thick undated, unfossiliferous section below reference would be made to the nearest section where the base of the Cambrian is known, the thickness of sediments below the middle Middle Cambrian is noted in this reference section and an equivalent thickness is placed in the Cambrian of the problematical section.

Wm. Lee Stokes, 1979

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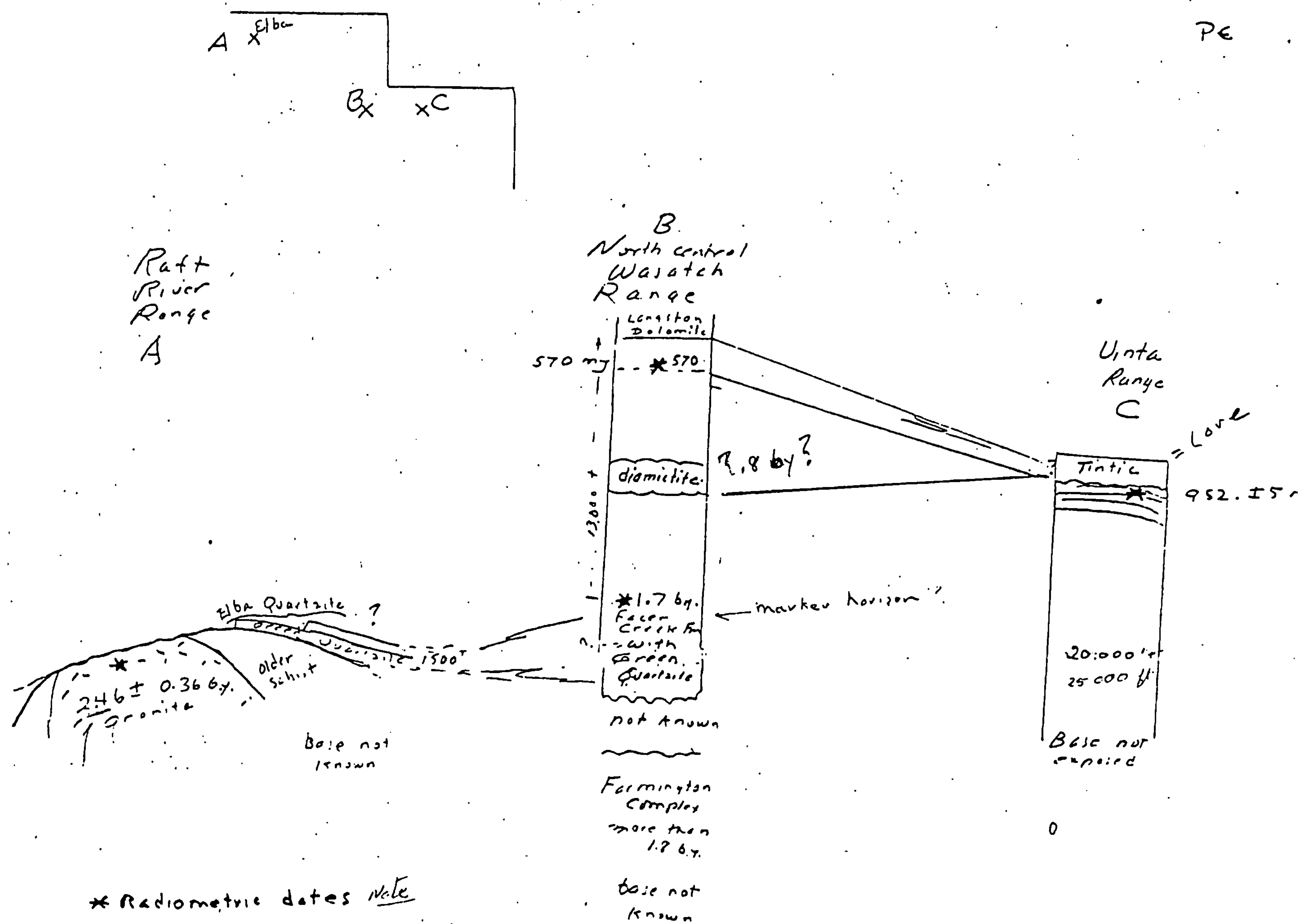


* Radiometric dates Note

W. Stokes 1976

5/24/79

PALEOZOIC SYSTEM BOUNDARIES IN THE WESTERN INTERIOR

Historical

The geologic column and time scale is memorized by beginning students of geology and is a constant tool of the professional worker during his active career. The geologic column is the framework for organizing most geologic reports and furnishes a unifing scheme for geologic maps everywhere. When supplied with relative or absolute dates the geologic column becomes a time scale. Fossils, rocks, minerals and events are cataloged in terms of this scale and the entire story of the prehistoric past is organized chronologically by reference to its successive divisions.

Anything so deeply ingrained and widely used should be well understood by those who work with it. The time scale grew up in a haphazard way as geology developed and most scientists would admit that this is not the best way for a basic classification to come into being. The historical fact is that geologists could not put off their need for classification until the entire earth had been mapped. In other words, they could not know all that would ultimately come under their classification when they began to classify. In hindsight it seems fortunate indeed that western Europe and Great Britain in particular should have been the birthplace of geologic science. Here, in a relatively small region all major segments of fossil-bearing periods of earth's history are represented to some degree and the geologic column as pieced together there has proved of use throughout the earth.

This is not to say that the geologic column and time scale are entirely satisfactory - far from it. To many workers it represents an archaic system that we are stuck with and have little hope of ever getting anything better. I like to discuss the topic of the extent of our slavery to the classical geologic time scale and have gone so far as to recommend a substitute for it.

What are the chief problems associated with use of the geologic time scale? Judged by the standards applied to any classification scheme: namely, how well and completely it covers that which it attempts to classify, the geologic time scale must be judged as somewhat inadequate. To be very specific geologists have encountered great volumes of sediments and even greater amounts of metamorphic and igneous rock that do not fit into the framework of the orinally defined type sections of the various systems. For the time being the knotty problem of dealing with the Precambrian must be ignored. What follws deals chiefly with the Phanerozic - the interval in which fossils are available for study and correlation.

Building upon their knowledge of the originally defined periods and their known fossil contents geologists went forth to map the entire earth. For want of anything better their correlations had to be based on fossils, the evolutionary changes, migrations, and extinction of which added problems and challenges every step of the way. But always there seems to have been enough fossils to keep correlations reasonable and the system going. Always there were methods other than paleontology to fill the gaps and geologists able to make educated guesses on scarcely any evidence at all.

But what to do with deposits formed during times of non-deposition or lacking adequate fossils in the classical type locations? The decision to be made in all cases was to which of two sequential systems any newly discovered sections should be assigned. Here and there a rebel has suggested that new systems ought to be created between the old ones but such attempts have almost entirely failed. Throughout the earth the gaps between the classical system have been filled in until at last it can safely be said that every instant of Phanerozoic time is represented by known deposits somewhere on earth. But this optomistic statement must be tempered by noting that every segment of the record is not equally well represented by accessible deposits and clearly not equally fossiliferous.

During the gradual accumulation of knowledge about the fossil-bearing rocks and their apportionment among the systems of the geologic time scale geologists have undergone a complete reversal in thinking about where the boundaries ought to be placed. At first it was believed there were universal breaks between the bodies of sediment representing the systems. These breaks were sought after and became major dividing lines on geologic maps. Now, that universal pauses, breaks, or unconformities have proven to be an illusion geologists are looking for those places where there are no breaks so that the systemic boundaries may be permanently fixed by objective fossil evidence.

An ideal reference section or stratotype should possess the following characteristics: 1) it should record continuous uninterrupted sedimentation, 2) it should contain a relative abundance of fossils of many types, including, 3) adequate numbers of the group or groups of fossils that have the greatest actual or potential use for fine-scale correlations in the particular interval of time under study. If and when the single section which possesses these qualifications to the highest possible degree is discovered, described and decided upon by experts, that section, regardless of its location, should become the world reference or stratotype. Stratigraphers, half in jest, have recommended that the exact plane of transition or true systemic boundary should be marked by a golden spike and henceforth held inviolate. From such a golden spike the designated time plane will have to be traced elsewhere thoughout the earth by any or all possible means.

The present situation is one of having to chose between a number of possible geographic localities with continuous sedimentation and evolution between each of the various consecutive systems. Geologists being human, have been less than objective in selecting the most suitable sites for the golden spikes of stratigraphy. Each geological expert or organization with a spot to suggest or defend has championed it with some degree of bias with the result that there are no clear-cut final choices at the present time.

The Practical Side of the Problem

The evolution of thinking about the placement of the systemic boundaries is reflected in practical field work and cartography. Any map showing chronostratigraphic units ideally should show lines between rocks of different periods regardless of their lithology if the map is to fill its intended purpose. Although maps of lithostratigraphic units are basically intended to show rock units without regard to time of formation there is still the necessity of appending a geologic age to each formation shown. Most of the earlier generations of map makers probably felt they should make special efforts to draw correct lines between the systems on any sort of map. It was somewhat of an adverse reflection on the skill of a fieldman not to be able to find and map the systemic boundaries. If William Strata Smith could do it so could they. After all that is the way formations are memorized - by associating them with specific periods. And it seems safe to say that the impression given in beginning textbooks is that geologists have no trouble in putting each and every formation in its proper period.

But these attitudes are changing with the discovery that there are no universal unconformities and the time honored breaks between periods are of no more significance than other breaks that actually exist. Geologists are no longer embarassed by not being able to draw every systemic boundary they know to exist in their map areas They are not too concerned about mapping formations that include two or three geologic systems. Nevertheless they must constantly bear in mind that there are two types of units - the lithostratigraphic and the chronostratigraphic and they may be called upon to produce maps showing both types of units.

The following 8 diagrammatic sections of the Western Interior represent the stratigraphic settings where the systemic boundaries come closest to being accurately fixed by the following criteria:

a. Continuous sedimentation

b. Fossil-bearing

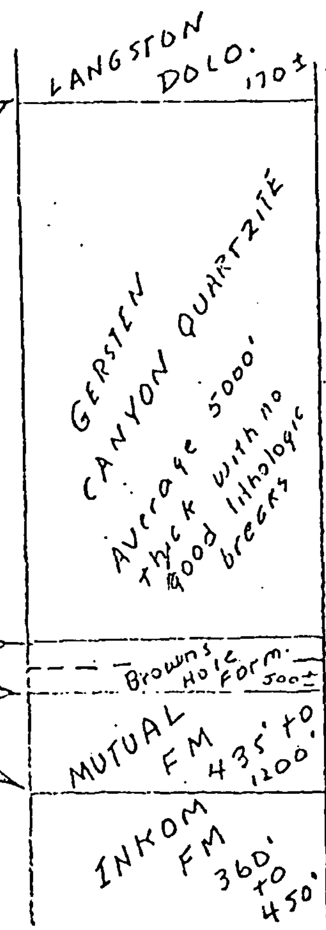
c. Contain representatives of the fossil group(s) most useful in drawing time planes in the interval concerned.

Think about the problems as you study subsequent sections or formations containing important time planes.

PRECAMBRIAN - CAMBRIAN

HUNTSVILLE AREA

Excellent cartographic bountary



The Langston Formation contains the Albertella trilobite zone recognized throughout the North American Province. No earlier zones have been discovered in the Northern Wasatch. Three previous zones each of which is certainly longer than the Albertella in duration are represented in the Gersten Canyon Quartzite or by the break at the base of the Langston.

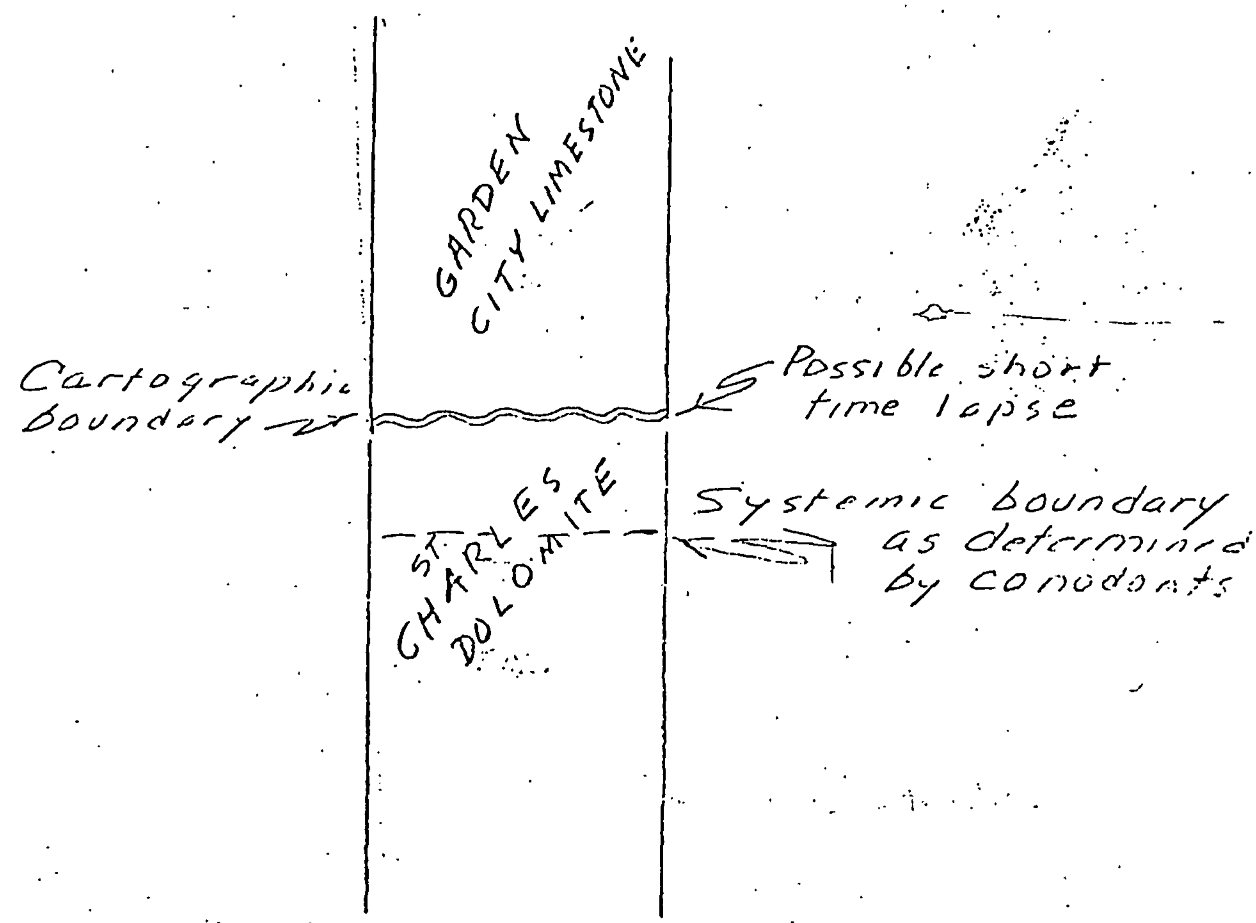
These are all excellent cartographic boundaries. There is no reason to suspect great lapses of time at any one of them.

Assuming beginning of Cambrian at 575 million years ago; the boundary is not far from a position in the Browns Hole Formation which has basalt flows dated at 570 m.y., an unusually fortunate date.

Reference:
U.S.G.S. Geologic Quadrangle Map; Browns Hole Quadrangle GQ-968, Max D. Crittenden, 1972

CAMBRIAN-ORDOVICIAN

BLACKSMITH FORK-CACHE CO.

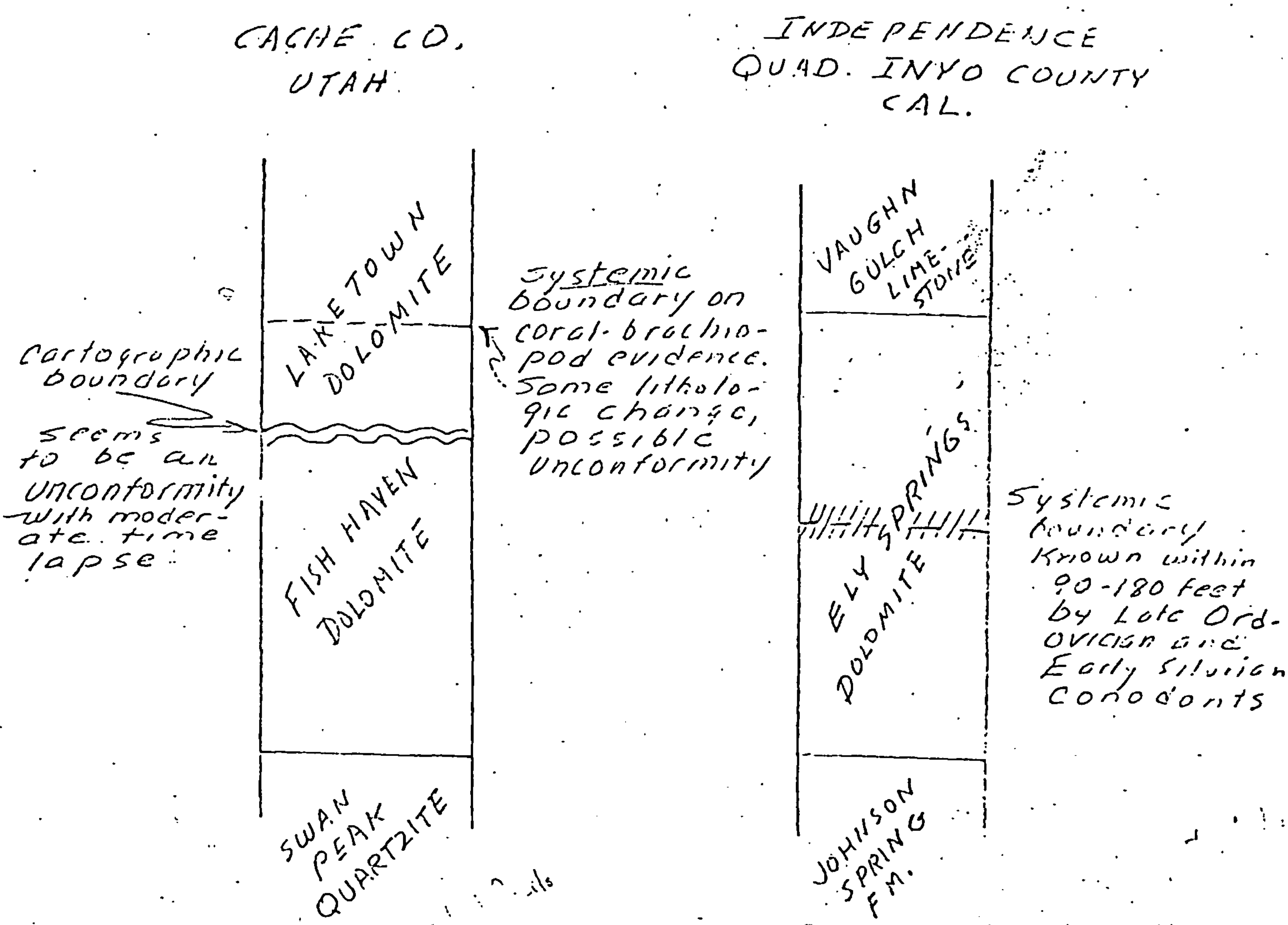


The cartographic break is between the Garden City Limestone and the St. Charles Dolomite, easily recognized by the change in lithology. This break is followed on all older maps.

The systemic break, determined by conodonts which are the best current world-wide index fossils, is below the lithologic break in the upper part of the St. Charles. Just how far down is not known but it is within the St. Charles Dolomite

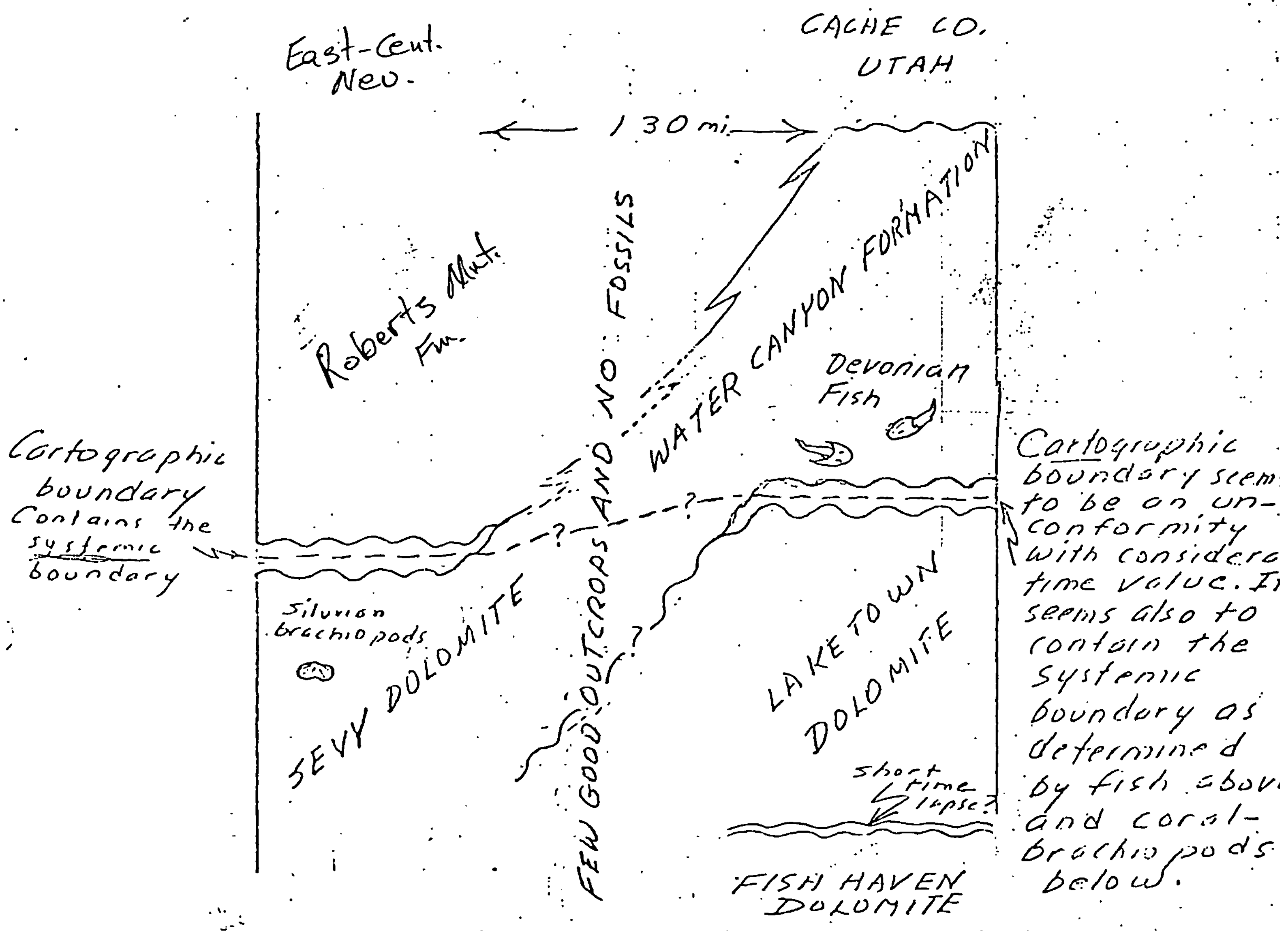
Reference: James L. Mason, 1976, Conodont biostratigraphy of the lower Garden City Formation (Lower Ordovician) northern Utah, Unpublished M.S. thesis, University of Utah.

ORDOVICIAN - SILURIAN

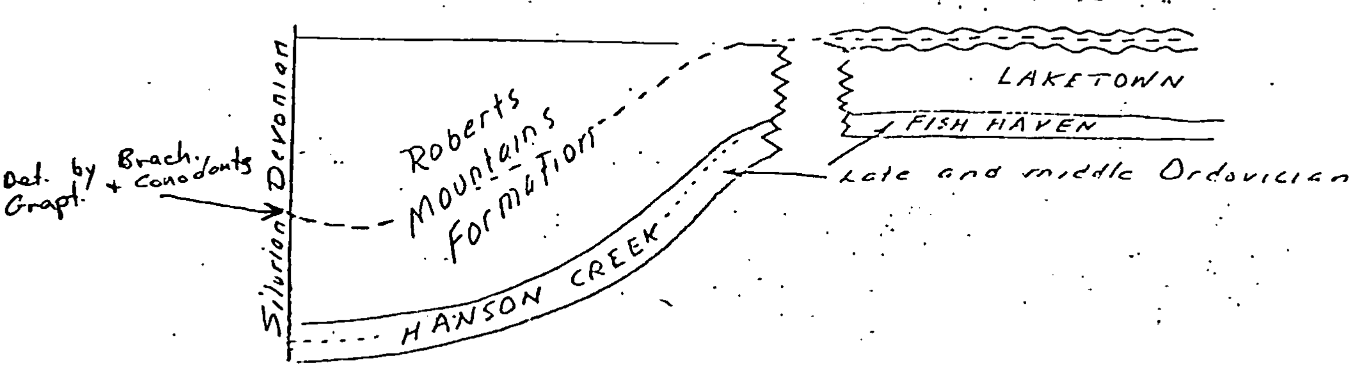


References. Richard H. Miller, 1975, Late Ordovician—Early Silurian Conodont Biostratigraphy, Inyo Mountains, California, Geol. Soc. Amer. Bull. Vol. 86, No. 2, p157-162

SILURIAN - DEVONIAN



EUREKA CO. NEVADA



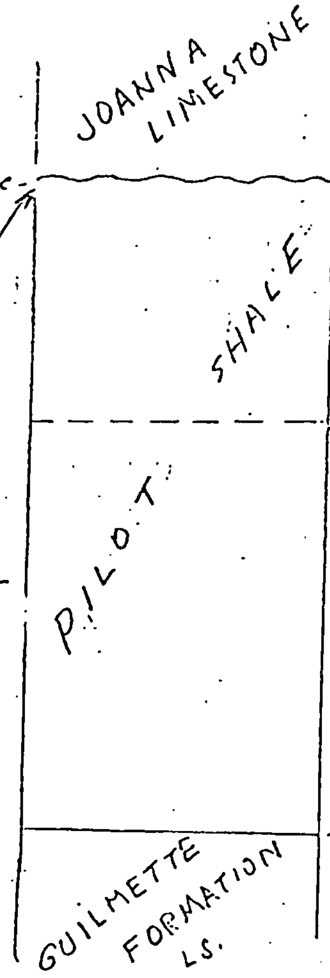
References: John C. Osmond, 1962, Stratigraphy of Devonian Sevy Dolomite in Utah and Nevada, Bull. AAPG, Vol. 46, p. 203
J. G. Johnson, 1974, Great Basin Silurian to lower Lower Devonian - a biostratigraphic case history, Bull. AAPG, Vol. 58/1, p. 139 f.

DEVONIAN - MISSISSIPPIAN

CONFUSION BASIN

The contact between the ledge-making Joanna Limestone and slope-forming Pilot Shale is one of the easiest levels to recognize and map in the entire geologic section of the Western Interior

A minor unconformity at base but no great time lapse.



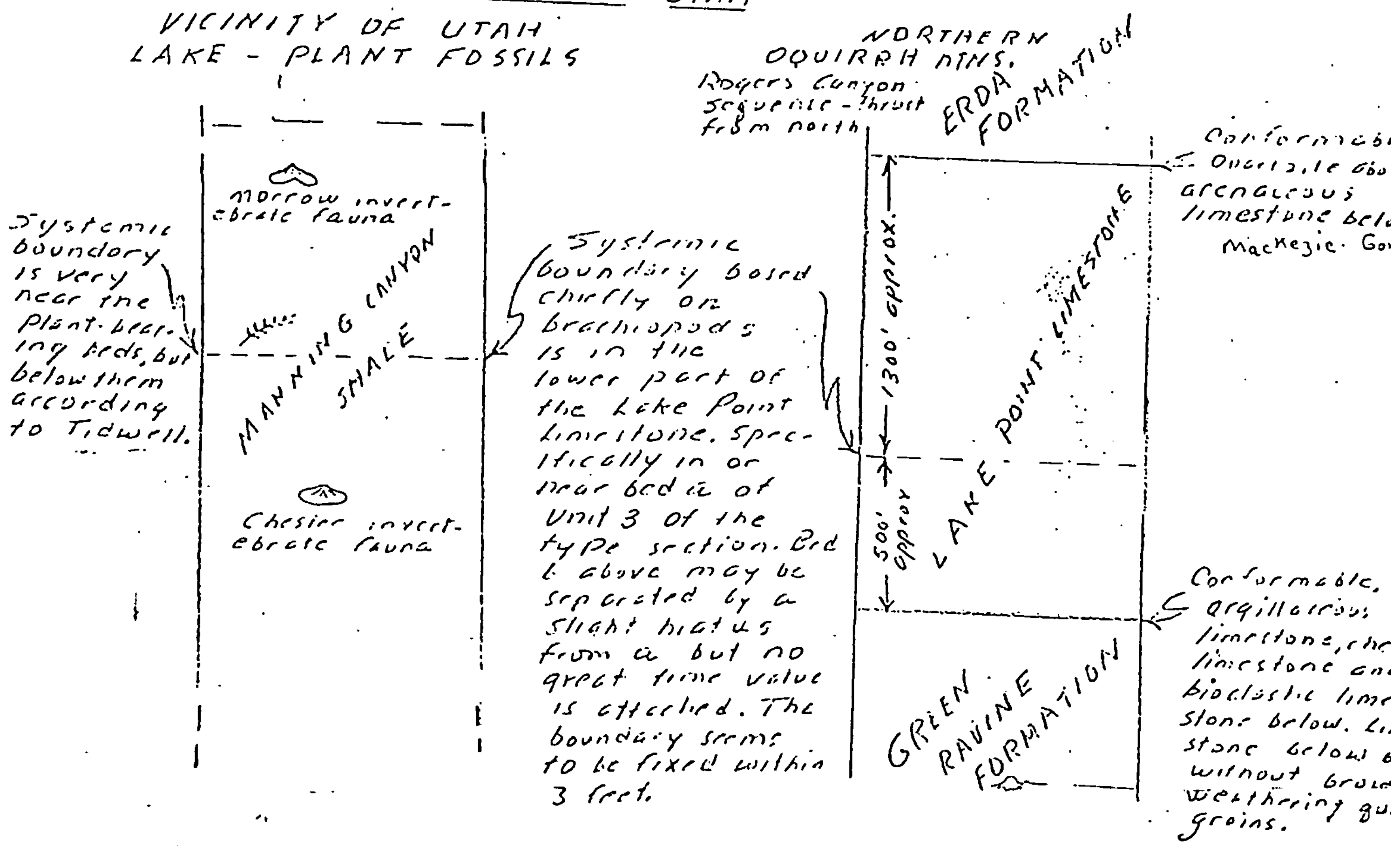
Systemic boundary located to within about 150 feet by brachiopods. (cyrtopsis) Cyrtopsis? occurs below and Spirifer Centronatus above

Good cartographic contact between shale and limestone. No great time lapse represented.

References: R. K. Hose, 1966, Devonian stratigraphy of the Confusion Range, west central Utah: U.S.G.S. Prof Paper 550-B

MISSISSIPPIAN - PENNSYLVANIAN

CENTRAL UTAH

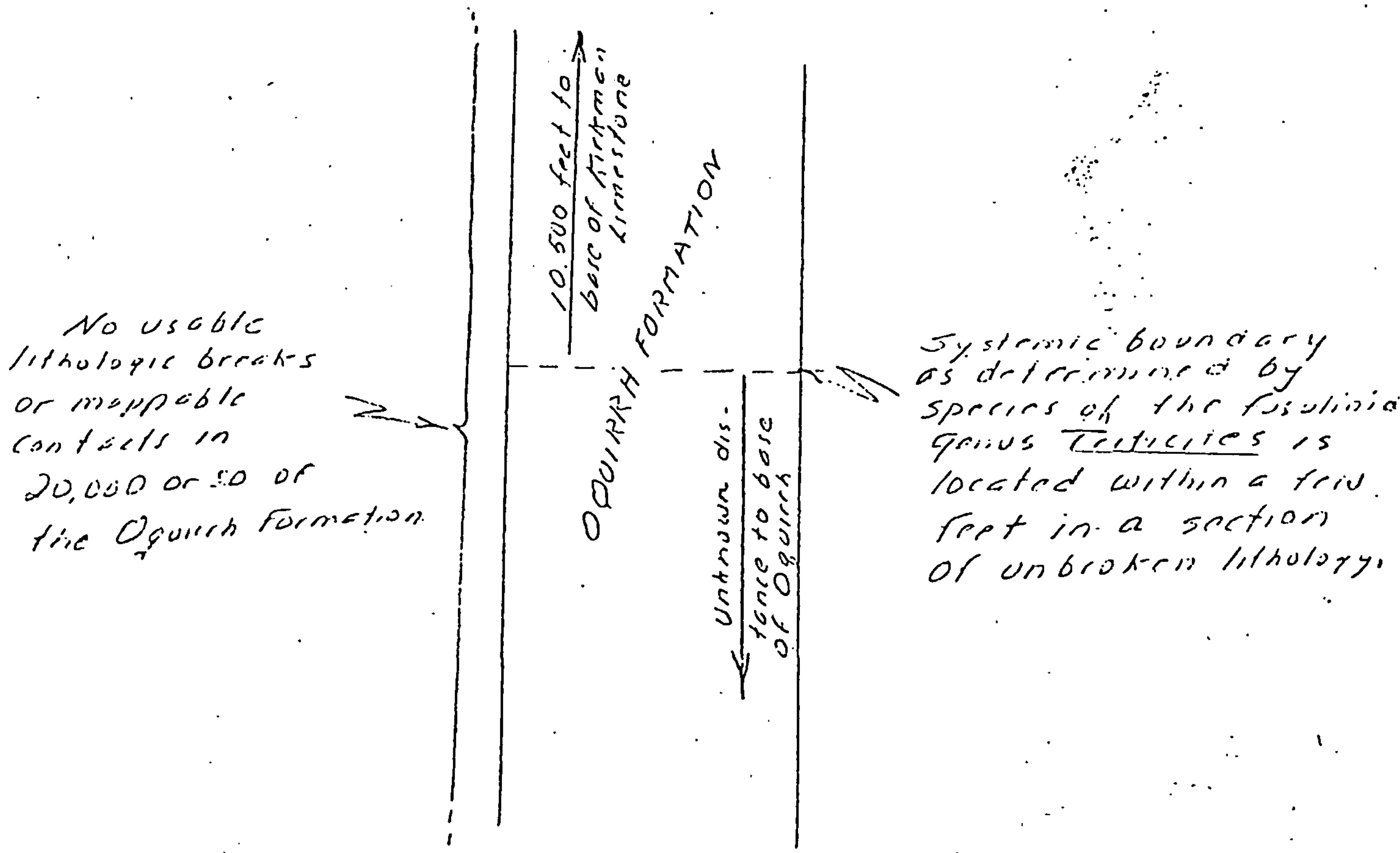


References:

Tidwell, W.D., 1967, Flora of Manning Canyon Shale Part I: A lowermost Pennsylvanian flora of the Manning Canyon shale, Utah, and its stratigraphic significance, Brigham Young University Geology Studies, Vol. 14. p. 35.

Gordon, Mackenzie, Jr. & Duncan, Helen M., Biostratigraphy and Correlation, in Upper Paleozoic Rocks in the Oquirrh Mountains and Bingham Mining District, Utah., U.S.G.S. P. Paper 629-A

PENNSYLVANIAN - PERMIAN

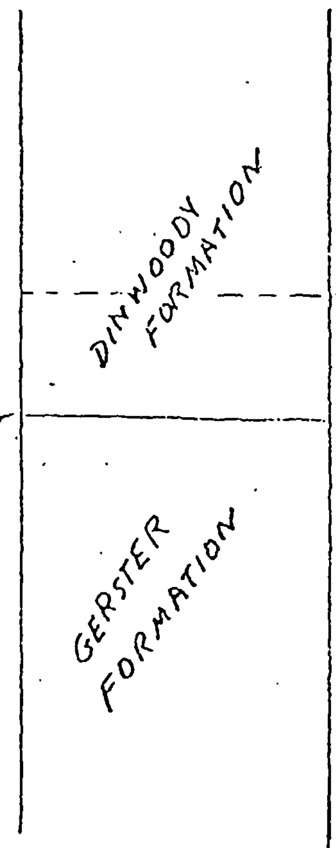


Reference: Thompson, M.L., 1954, American Wolfcampian Fusulinids, (section by M.L. Thompson & H. J. Bissell on Central Utah, p27 f.) in University of Kansas Paleontological Contributions, Protozoa, Article 5. May 5, 1954.

PERMIAN - TRIASSIC

TERRACE MOUNTAINS
UTAH

Systemic boundary is within olive-colored shale sequence usually mapped as Dinwoody. Position will probably be fixed within a few inches or feet by conodont index species. Magnetic stratigraphy will come into use here.



Cartographic boundary is at the top of gray limestone with large productid brachiopods estimated 80' below Permian-Triassic boundary

Reference: Stifel, Peter B., 1964. Geology of the Terrace and Hogup Mountains, Box Elder County, Utah, Unpublished Ph.D. thesis, University of Utah.

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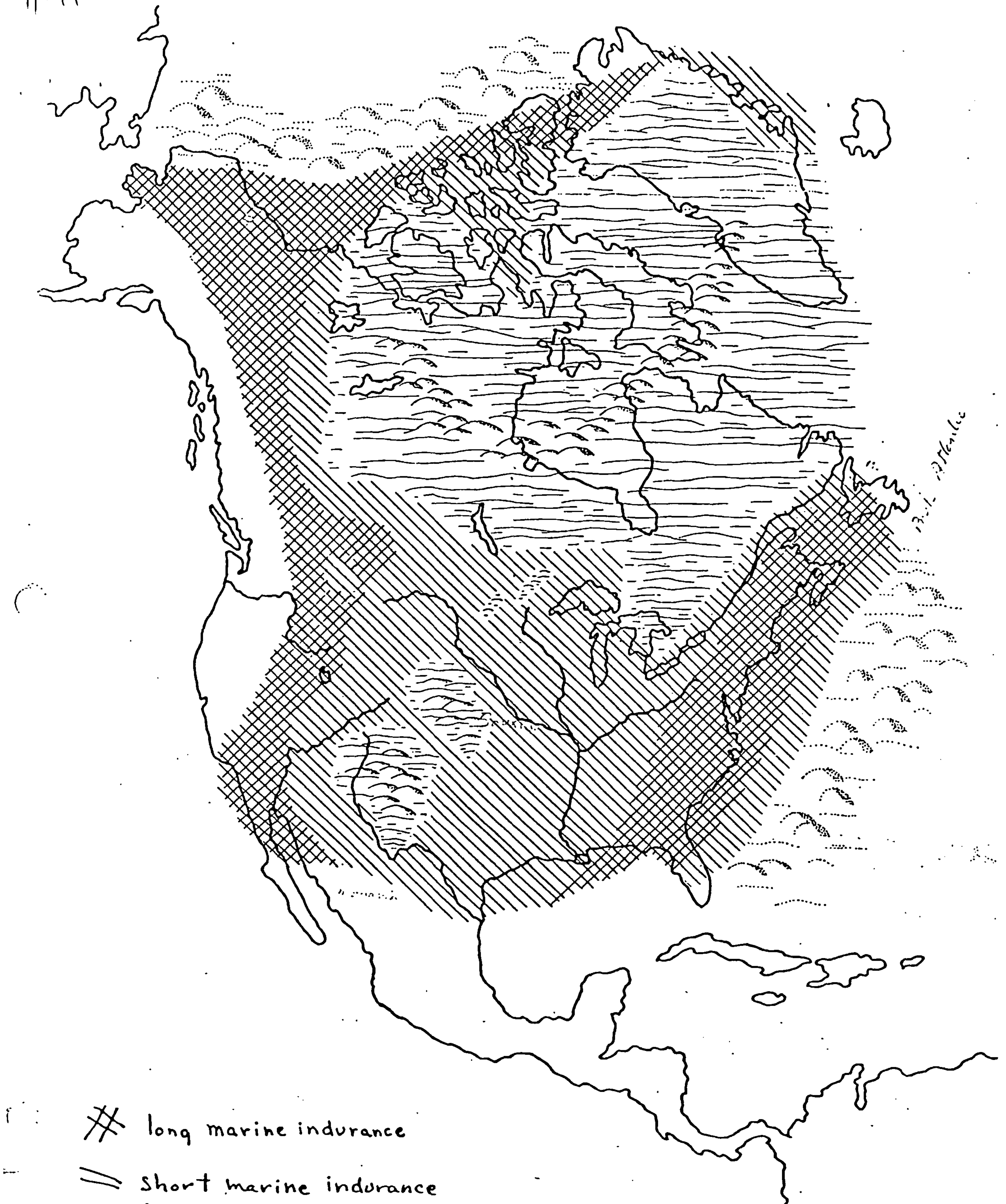
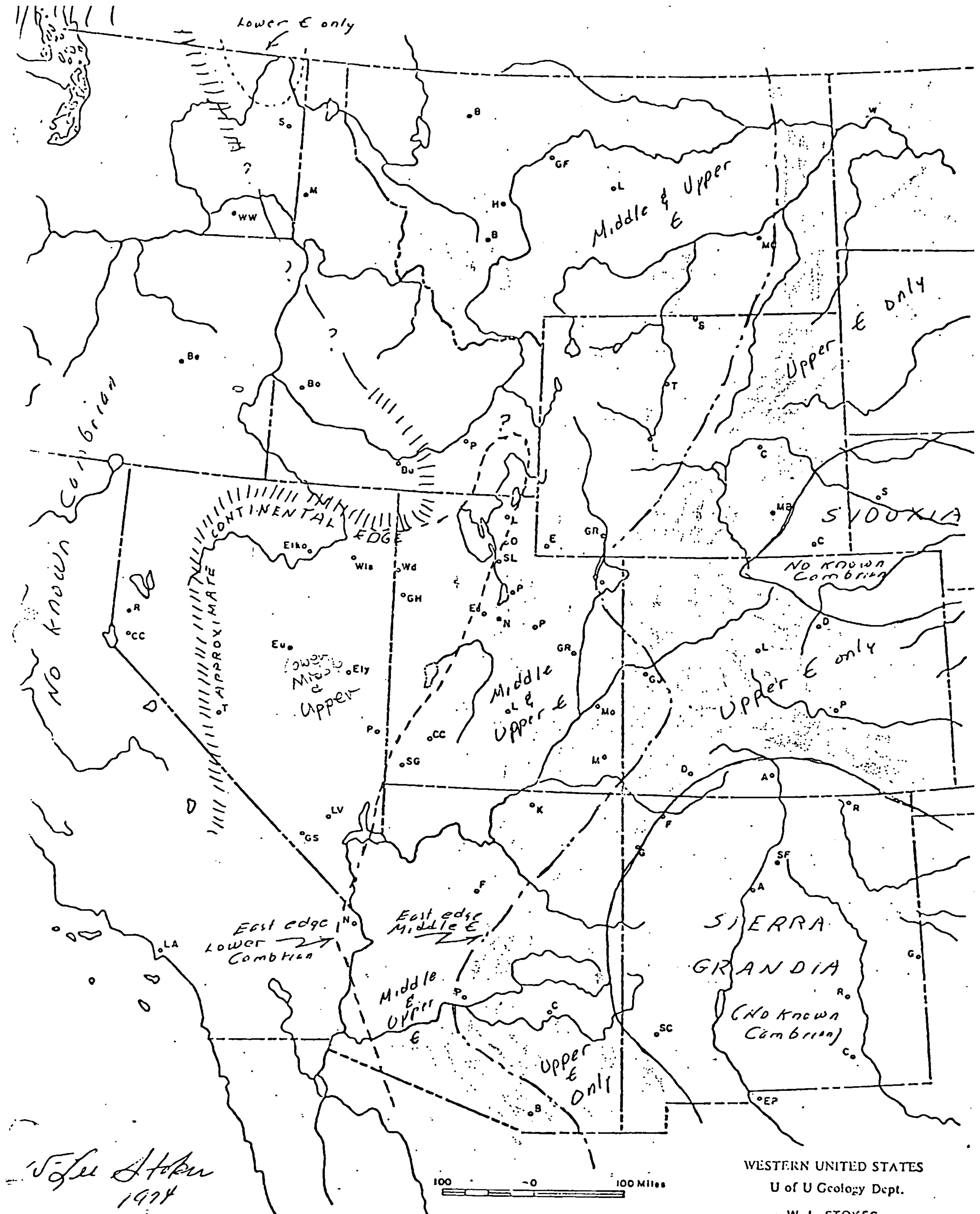




WESTERN UNITED STATES
U of U Geology Dept.
W. L. STOKES
1971

Wm Lee Stokes
1974

1/12/79

Generalizations

Paleozoic stratigraphy of the Western Interior

CAMBRIAN

I. Historical:

A. Named by Adam Sedqwick in 1835 after the Roman name for Wales. Was not well defined and accepted until 1879 when Charles Lapworth separated and distinguished the Ordovician. Position of the base is still debated.

B. The American geologist C. D. Walcott suggested a division of the Cambrian into Lower, Middle and Upper time and rock units. Corresponding to these divisions are the Waucobian, Albertian and Croxian for much of western North America and the Georgian, Acadian and Saratogan of eastern North America. These place names appear to be unnecessary.

II. Tectonics and Paleogeography

A. A cratonic nucleus, corresponding to the present exposed and covered shield was the major controlling feature. Its relation to other continental or oceanic areas is unknown.

B. The Transcontinental Arch - a southwesterly oriented extension of the craton was the chief positive element in the United States. It controlled sedimentation on the south and east.

C. The Cordilleran Orthogeosyncline had already been established along the western border of North America in Late Precambrian time. The edge of the Cambrian continent was roughly parallel to but farther inland than the present one. If there were notable coastal irregularities they were probably not entirely tectonic.

D. During the Cambiran the Wasatch Line was evident as a sort of arc-shaped hinge or transition zone between the shelf area and the orthogeosyncline.

E. Tectonics of the Pacific border beyond presently preserved Cambrian rocks are poorly understood. There may have been a bordering land mass (Cascadia of older writers) but this is doubtful. The ancestral Pacific Ocean may have been narrower, partly occupied by fragments that would become part of Asia. Certainly not much clastic material can have had a western source.

F. The Cambrian, together with the Late Precambrian was a time of unusual tectonic quiet. No evidence of mountain building or strong movements of lithospheric plates. Few evidences of igneous activity.

G. Among the important positive features of the earlier Cambrian stages are the lowland mass, Montania, the Uinta peninsula, and the seemingly rugged and irregular Defiance-Mongollon Highlands. With the passage of time only a few low island-like lands remained-Siouxia, Sierra Grandia, Defiance-Mogollon, and Mojavia. All these are aligned along the Transcontinental Arch.

III. Lithology and Sedimentation

A. Cambrian was a lengthy period - 70-100 m.y. Volume of original sediment was great and much remains deeply buried or is being eroded.

B. The Cambrian of the Western Interior is the great classic example of a time-transgressive system. The ocean flooded onto the craton depositing successively coarse clastic formations (18 named), followed by mixed clastic-carbonate and fine clastics and finally thick carbonate formations. Each suite rises and becomes younger eastward. Walther's Law applies; that which occurs vertically is repeated laterally.

Although the general trend was transgressive there were many temporary regressions. Cycles, perhaps of eustatic origin are also evident. Lithofacies trend generally parallel with Wasatch Line.

C. Rock types are dominantly shallow marine or littoral. A very rough quantative estimate is: sandstone-quartzite-grit-conglomerate, 40%; fine-clastic; 20% limestone and dolomite, 40%.

1. Practically no igneous material - a few thin flows in Early Cambrian of Utah-Nevada. Rare pillow lava.

2. Arkose rare, the Worm Creek quartzite may be a longshore deposit from the north (Montania).

3. Chert is remarkable rare until late Middle Cambrian.

D. The most fruitful concept in Cambrian stratigraphy is that of persistent lithofacies belts: a) inner detrital, b) carbonate bank, and c) outer detrital. These represent the parallel and contemporaneous deposition of near shore, craton-derived clastic-rich material, varied carbonate rock types much of which are of algal shallow-water origin and deeper-water deposits rich in fine terriginous clastics. Inter-relations are complex. Christina Lochman-Balk recognizes the following lithotopes: Littoral-transgressive sand, littoral regressive sand, tidal (intertidal to intratidal) sand, tidal (intertidal to intratidal) mud, tidal sand-mud, tidal carbonate, tidal, (intertidal to intratidal) carbonate ooze, sublittoral shelf-argillaceous sand, sublittoral shelf-sandy mud, sublittoral shelf-carbonate, miscellaneous sublittoral deeper water.

E. Cambrian carbonates and fine-clastics show a remarkable variety of primary features: laminations, mottling, dessication cracks, intraformational conglomerates and breccias, stromatolites, oolites, oncholites, etc.

IV. Biostratigraphy

(chiefly a topic for Geology 675, Spring Quarter)

Cambrian life was abundant but relatively simple. Fossils indicate approximately the following order of decreasing abundance of animal forms: trilobites, (including agnostids), brachipods, echinoderms, coelenterates (no corals), conodonts (Late Cambrian only), gastropods (incl. hyolithids), possible cephalopods, pelecypods and bryozoans. Base of most Cambrian food pyramids is shallow-marine algae.

Trilobites are abundant and extremely diverse and are the basis for practically all Cambrian biostratigraphic units. Currently recognized zones (Correlation chart, 1944) are temporary and unsatisfactory. Studies by numerous competent paleontologists will result in great refinement and percisions but there are basic philosophic differences among current workers. Various types of zones - range, assemblage, and concurrent - range - are to be found in the literature. A new concept that may be useful is that of the biomere defined as "a regional biostratigraphic unit bounded by abrupt non-evolutionary changes in the dominant elements in a single phylum. These changes are not necessarily related to physical discontinuities in the sedimentary record and they are, or may be, diacronous.

The degree to which trilobites are environmentally controlled gives rise to other differences of classification. Are separate schemes of zonation needed for the cratonic and extra-cratonic environments? Were there subtle paleoclimatic or sedimentological zones that cannot safely be lumped in the same systems?

Finally, much study will be required to establish the relative accuracy and utility of biostratigraphic versus lithostratigraphic divisions. The Cambrian of the Western Interior will be a fruitful ground for much future study where important principles applicable to other systems will be developed.

V. Boundaries

A. Criteria for fixing the base of the Cambrian are not agreed on, it is a problem of world-wide significance. As a practical matter it exists only where there was continuous sedimentation from positively identified Precambrian into positively identified Cambrian which was the situation in much of western North America. Solutions include: (a) commence Cambrian at the major unconformity above the crystalline basement, b) commence Cambrian at base of a strong transgressive clastic unit, c) commence at first certain signs of metazoans, d) commence at first certain evidences of trilobites-including trace fossils, e) commence with first body fossils of trilobites.

B. The Cambrian-Ordovician boundary is not agreed on. It will eventually be established on faunal frounds but only after study of the most complete and continuous sections available. A new standard for Late Cambrian of North America (away from Wisconsin Dome) is needed. Chief guides will be trilobites and conodonts. Conodonts migrate rapidly, trilobites slowly, this results in a diacronous boundary problem. Perhaps the Great Basin will become the worlds stratotype for the Cambrian-Ordovician Boundary.

VI. Economics

1. Cambrian formations are host rocks for hydrothermal deposits in a number of mining districts chief of which are the Cottonw-od, Tintic, and Ophire districts in Utah and the Eureka, Pioche, and Groom districts in Nevada. In the carbonates the chief metals are silver-lead, zinc, gold and copper. Vein deposits in the quartzitic sections yield copper, gold and silver.

2. Limestone and dolomite used for flux and cement making are quarried from Cambrian rocks in a few places. Largest production comes from the Cricket Mountains, near Black Rock, Millard County, Utah.

VIII. Scenic

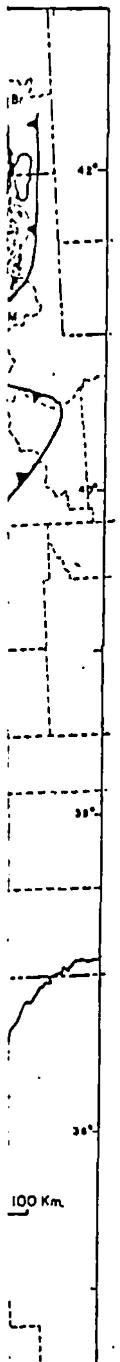
Cambrian rocks are mostly drab and uninteresting. Among the few notable scenic outcrops are the spectacular cliffs forming the western scarp of the House Range and several stretches in Logan Canyon. Cambrian rocks form the intermediate slopes in the Grand Canyon. There are many sinks and solutions features in Late Cambrian rocks in the Bear River Range west of Bear Lake.

Bibliography

1. 1972, "Geologic Atlas of the Rocky Mountain Region" Rocky Mountain Association of Geologists, Denver. The Cambrian is discussed by Christiana Lochman-Balk, p. 60-76.

2. 1971, "Cambrian of the New World" ed. C. H. Holland, Wiley Interscience, New York. Chapters on western Cambrian by Allison Palmer and Christiana Lochman-Balk, Good bibliography.

W. L. Stokes



? A—Albion Range ? Probably Carboniferous
An—Antelope Range
Ba—Bannock Range
BM—Bare Mountain
BR—Bear River Range
BeM—Beaver Mountains
BD—Beaver Dam Mountains
BeR—Belted Range
Ca—Canyon Range
Ch—Cherry Creek Range
Co—Cortez Range
Cr—Cricket Range
DC—Deep Creek Range
De—Delamar Range
DR—Desert Range
DM—Drum Mountains
DuR—Dugway Range
DuM—Durst Mountain
E—Eureka area
ET—East Tintic Mountains
G(N), Eg(S)—northern and southern Egan Range
El—Ely Range
F—Frenchman Mountain
Fu—Funeral Mountains
G—Galena Range
GC—Grand Canyon
GR—Grant Range
Gr—Groom Range
Ha—Halfpint Range
Hi—Highland Range
Ho—Hot Creek Range
HR—House Range
IR—Independence Range
In—Inyo Mountains
L—Lakeside Mountains
LC—Last Chance Range
Lo—Long Ridge
Ma—Malad Range
MM—Marble Mountains
Mo—Morman Mountains
MV—Mt. Velma area
Mu—Muddy Mountains
NY—New York Mountains
NR—Nopah Range
Oq—Oquirrh Mountains
Os—Osgood Mountains
Pa—Pahranagat Range
PaR—Panamint Range
PR—Pavant Range
Po—Portneuf Range
Pr—Promontory Range
PM—Providence Mountains
RS—Resting Springs Range
RR—Ruby Range
SF—San Francisco Mountains
S—Santaquin area
SC(N)(S)—northern and southern Schell Cr
Range
SM—Sheep Mountain
Sh—Sheeprock Range
ShR—Shoshone Range
SI—Silver Island Range
SR—Snake Range
Sp—Specter Range
SpM—Spring Mountains
St—Stansbury Range
To(N)(C)—northern and southern Toiyabe Ran
V—Virgin Mountains
WW—Wah Wah Mountains
Wa—Wasatch Mountains
We—Wellsville Mountains
Wh—White Mountains
WP(N)(S)—northern and southern White I
Range
Wi—Willard Peak

FIG. 2. Index map of the Great Basin region showing distribution of known areas of outcrop

1/12/79

CAMBRIAN ASSIGNMENT
Jan 12 '79

Do one of the following after researching the Cambrian literature.

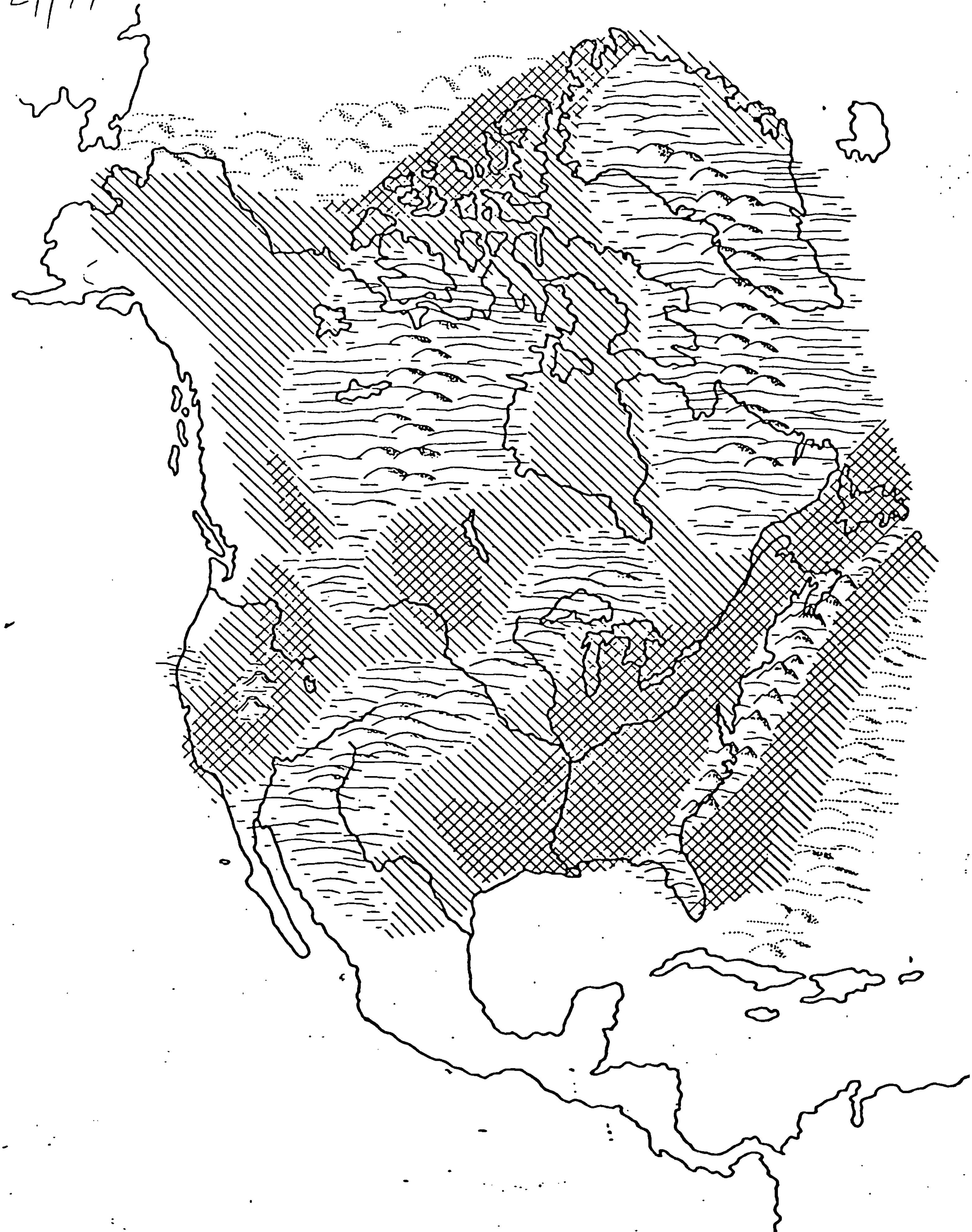
1 Discuss the evidence for a Uinta Peninsula, explain it in terms of tectonics if possible and give an opinion as to whether it is a good term for general usage. About 2 pages plus references and one diagram will do.

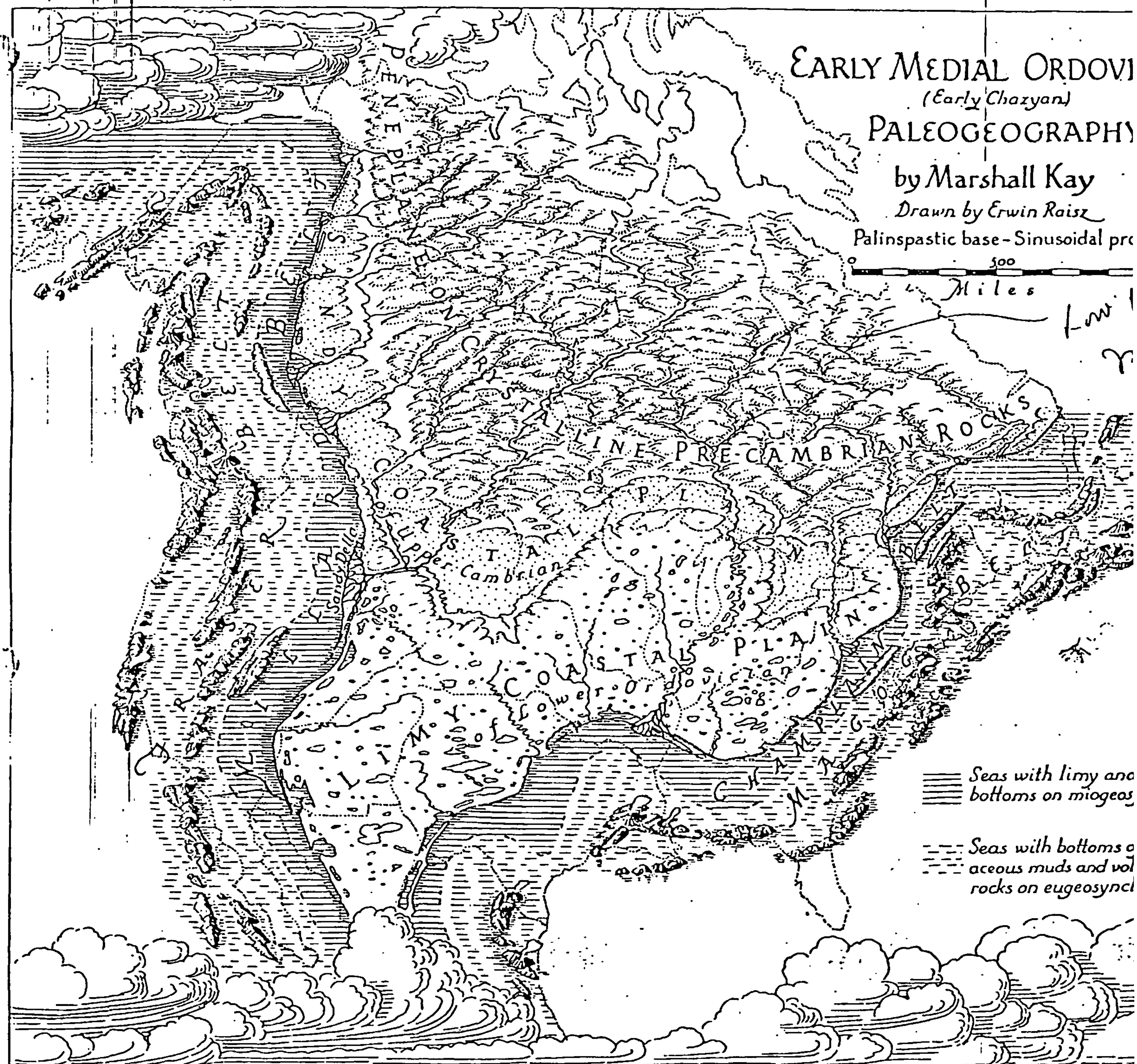
OR

2. Discuss the various terms and terminology in print describing the carbonate deposits of the Cambrian. Do you think a unified scheme is possible or desirable? I mean environmental, tectonic, and geographic terms. If you favor a certain terminology please give it in a diagram.

Several pages plus biblio. and diagram will do.

1/24/74

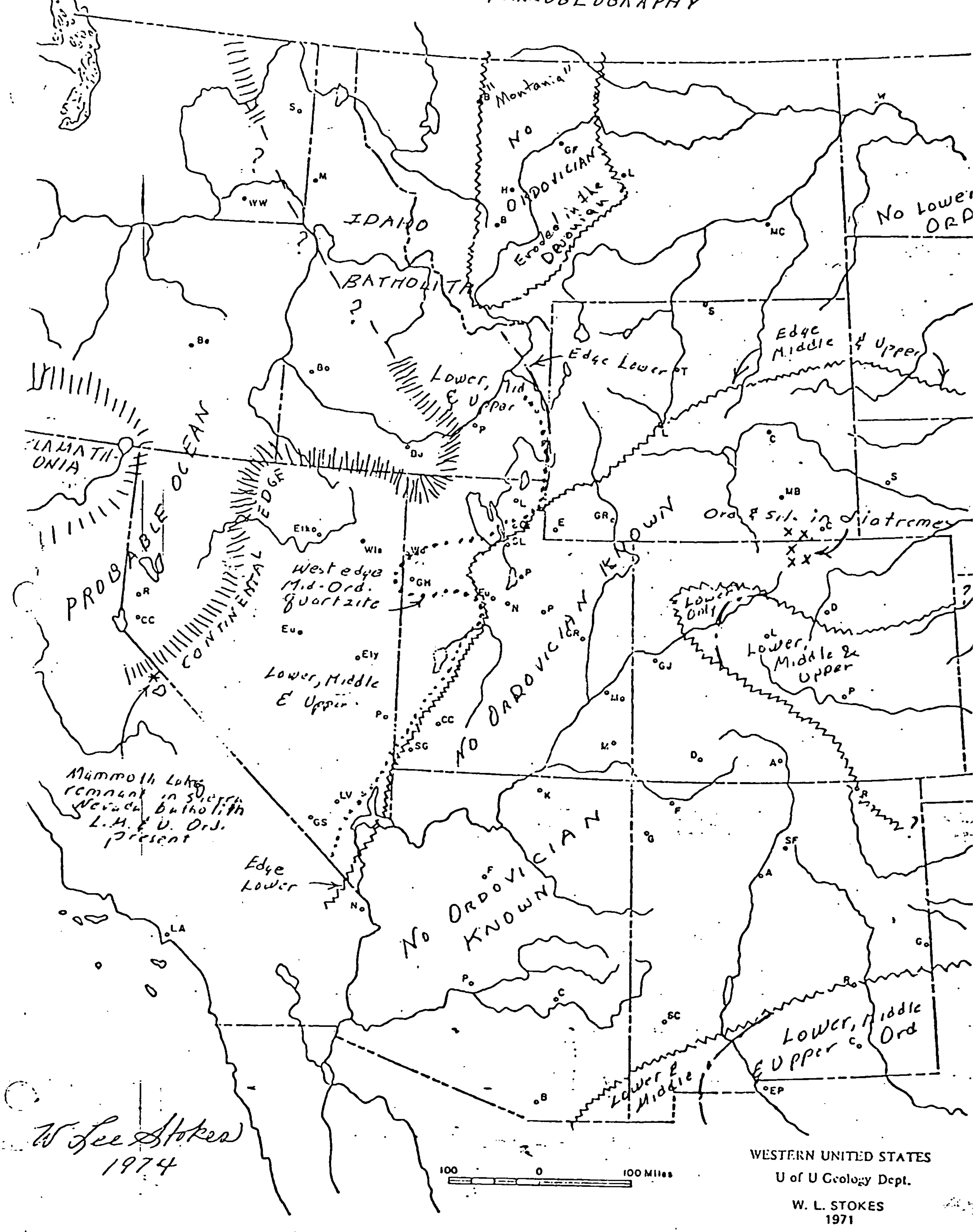


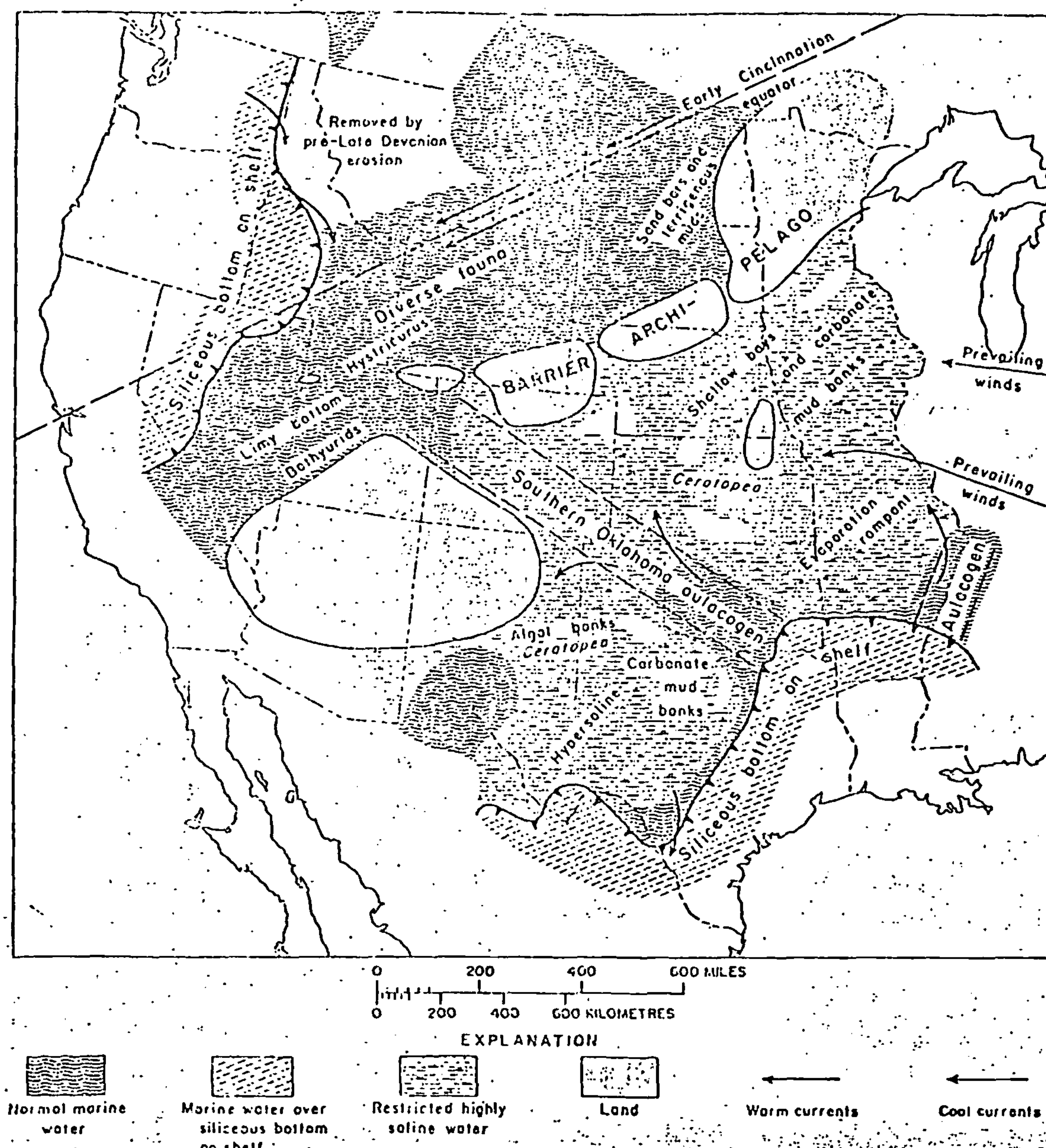


PALEOGEOGRAPHY OF EARLY MEDIAL ORDOVICIAN OF NORTH AMERICA

Kays - Geosynclines of N America

ORDOVICIAN PALEOGEOGRAPHY





TEXT-FIG. 5. Interpretation of early Ordovician palaeoenvironments, western United States.

organisms made up less than 15% of the Oneota fauna and none of the Shakopee fauna; gastropods and cephalopods comprise most of the fossils.

The dolomitic strata of the Jefferson City, Cotter and Powell formations around the Ozark uplift (text-fig. 2) (Cullison 1944) have yielded a composite fauna consisting at the generic level of gastropods (including *Ceratopea*) 27%, cephalopods 30%, trilobites 24%, brachiopods 12% and sponges 6%. That is 57% of this fauna is molluscan and only 18% is composed of sedentary or attached genera. Three of the

1/24/79

Generalizations

Paleozoic Stratigraphy of the Western Interior

ORDOVICIAN

I. Historical

A. Named by Charles Lapworth in 1879 after the Ordovices, former inhabitants of the type area in Wales. The Ordovician had been classed by Murchison as Lower Silurian and by Sedgwick as Upper Cambrian.

B. In America the custom has been to recognize Lower, Middle and Upper divisions. The corresponding place names, Canadian, Mohawkian and Cincinnatian are widely used but the areas designated by these terms may not be the best available as standards for the United States. Great Basin Early Ordovician is probably better than that of the east.

II. Tectonics and Paleogeography

A. The major tectonic features of the Cambrian - Craton, Transcontinental Arch, Cordilleran geosyncline - continued to be the controlling elements of the Ordovician.

B. A distinct subdivision of the orthogeosyncline into an outer (eugeosyncline) and inner (miogeosyncline) is evident throughout the period. (see appended chart.)

C. The nature of the western margin of the continent remains unknown but the existence of volcanic archaeopelagos with rather extensive islands is certain. Ordovician deposits, representing the oldest proven Paleozoic of the Pacific border are found in the Klamath Mountains. This seems to represent evidence of an incipient land mass, Klamathonia.

D. Negative (downsinking) reactions mark the beginning of the Williston Basin as a persistent major tectonic feature. Every Period Represented in Williston Basin

E. The Transcontinental Arch shows clearly positive influences as there is little Ordovician preserved along it in the Western Interior. An exception is in the narrow Colorado Sag which trends northwesterly across Central Colorado and contains rocks of Early, Middle and Late Ordovician age. Of significance are isolated down-dropped remnants of Ordovician-Silurian in the Precambrian, Colo-Wyo, near Laramie.

F. The Wasatch Line is clearly in evidence, the erosional and depositional edges of many Ordovician formations follow this feature.

G. The Tooele Arch, apparently an ancient westward expression of the Uinta Mountain trend seems to have influenced sedimentation in west Utah. (Uinta-Cortez Axi

H. A rather ill-defined island-like area with no Ordovician may mark the persistence of Montania in western Montana.

III. Lithology and Sedimentation

A. Ordovician has an estimated duration of 70 m.y. Volume of sediment is less than Cambrian, more than Silurian. Exposures are numerous in the Basin and Range and there are extensive subsurface remnants in Wyoming shelf and Williston Basin.

B. The rock record reflects the following history: continuation of Late Cambrian patterns of sedimentation and marine environments through most of the Early Ordovician; withdrawal of the seas and deposition of thin, but extensive clastic formations, perhaps partly subaerial, during most of the Middle Ordovician; return of the seas in an extensive flooding producing relatively pure carbonate formations in the Late Ordovician.

C. Rock types are mostly marine of shallow to deep water origin. Rough quantatative estimate: limestone 30%; fine clastic (including deep-water black shale) 20%; quartzite, and sandstone, 10-15%; dolomite, 25%; chert, 5-10%; volcanic derivatives, 5-10%.

D. Rock types are those usually associated with geosynclinal belts: miogeosyncline has typical carbonates (including exceptionally pure dolomite), clean washed sand, and minor fine clastics; eugeosyncline has abundant black shale, chert, graywacke, argillite, and lava. Transition suite is very complex.

Vinini Fm. Chert, Graptolitic Shale

C. abt. 80% of continent submerged

E. The lithologic succession in the miogeosyncline is distinctive and persists over great distances. Lower formations are limestone (Garden City, Pogonip), the middle formations are quartzite (Swan Peak, Eureka) and the upper formations are dolomitic (Bighorn, Fish Haven, Hansen Creek).

1. The middle clastic units are typical of an almost continent-wide Mid-Ordovician sand sheet that includes the well-known St. Peter Sandstone. Origin not well understood.

2. The upper dolomite is part of another very extensive carbonate sheet which extends far into Canada.

F. The passage from the typical miogeosynclinal section into the eugeosyncline is abrupt - 20 miles or less. West of this zone of rapid facies change silica-rich rocks of volcanic and deep water origin are dominant. The relationships are difficult to map because of extensive involvement in the later Antler and Sonoma Orogenies.

IV. Biostratigraphy

Ordovician faunas are an orderly outgrowth of those of the Cambrian with important additions so that all known preservable phyla were in existence by end of the period. The approximate order of decreasing abundance is: trilobites (Early Ord. only), graptolites, brachiopods, coelenterates (corals), bryozoans, echinoderms (crinoids, blastoids), cephalopods, conodonts, gastropods, pelecypods, eurypterids, chordates. Weakly constructed reefs begin to appear. Predaceous forms such as orthocone cephalopods, eurypterids and ostracoderms are gaining.

Trilobites continue to be abundant and useful guide fossils during the Early Ordovician but appear to have suffered a great decline with the Mid-Ordovician regression. The best world-wide zone indicators are graptolites, the first abundant floaters. Other useful index species are found among the corals, conodonts, and cephalopods. No positive land life yet.

V. Boundaries

See Cambrian generalizations for lower boundary problem.
The Silurian-Ordovician boundary in the Western Interior is within an unconformity almost everywhere. A world standard for the boundary will have to be fixed in some yet undefined locality with abundant fossils and contin-

uous sedimentation. Because of the scarcity of fossils and widespread dolomitization both above and below the contact in the Western Interior many mistakes seem to have been made in placing the systemic boundary. Use will have to be made of what fossil evidence is available, corals and brachiopods which can be recovered in silicified form from dolomite are proving useful in the miogeosyncline; graptolites are best in the black shale.

VI. Economics

Ordovician rocks are host rocks in a number of Great Basin mining districts. Dolomite is mined for flux and cement in the Fish Haven in Tooele County, Utah. Oil is produced from deep pools in the Williston Basin.

VII. Scenic

Nothing spectacular. Many stretches of Logan Canyon show fine cliffs of Ordovician formations. Logan Cave in this canyon is in the Garden City Limestone.

Bibliography

1972 "Geologic Atlas of the Rocky Mountain Region" Rocky Mountain Association of Geologists" Denver. The Ordovician is discussed by Norman H. Foster, p. 76-85.

W. L. Stokes,

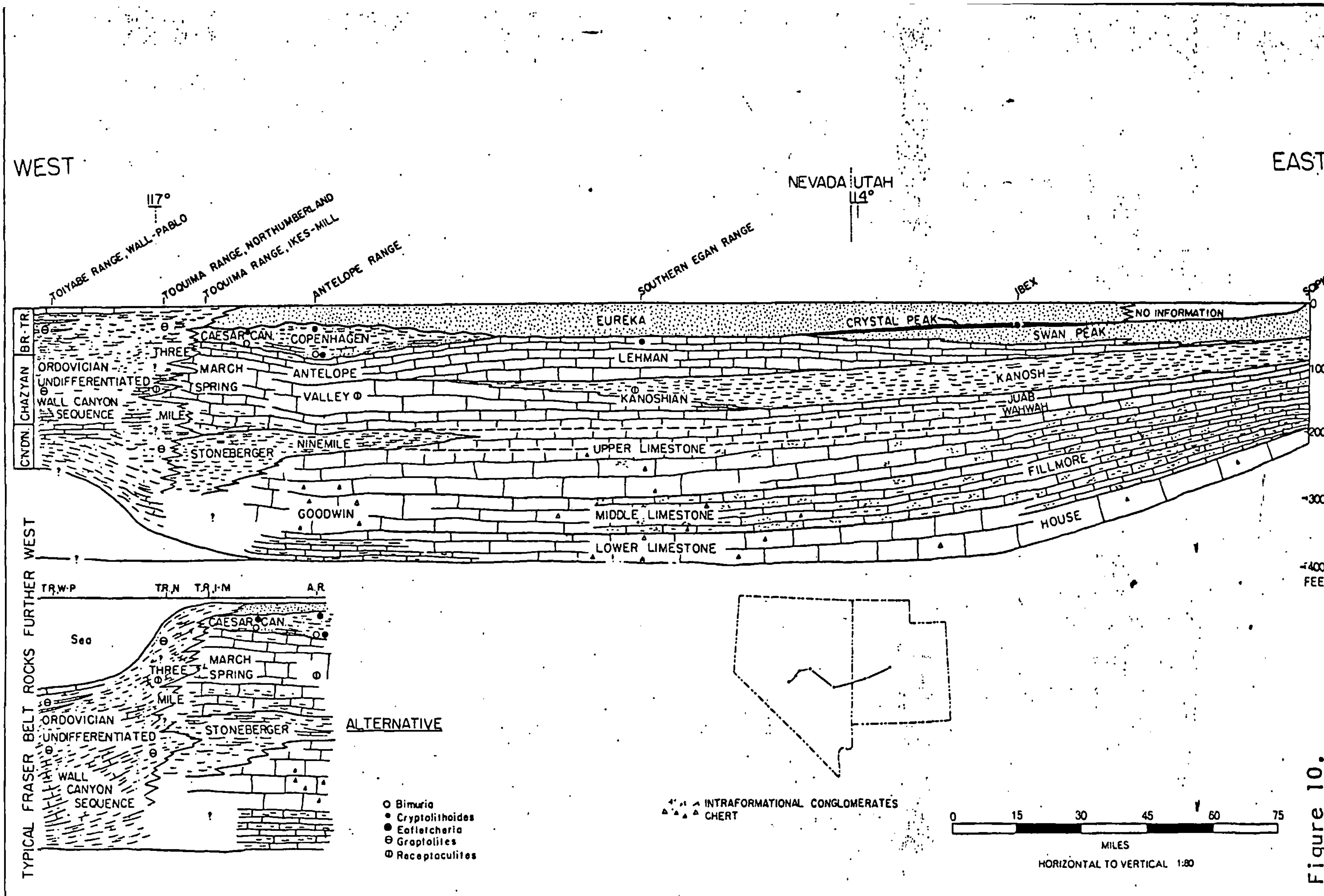


Figure 10.

Restored section of Lower & Middle Ordovician rocks from central Uta

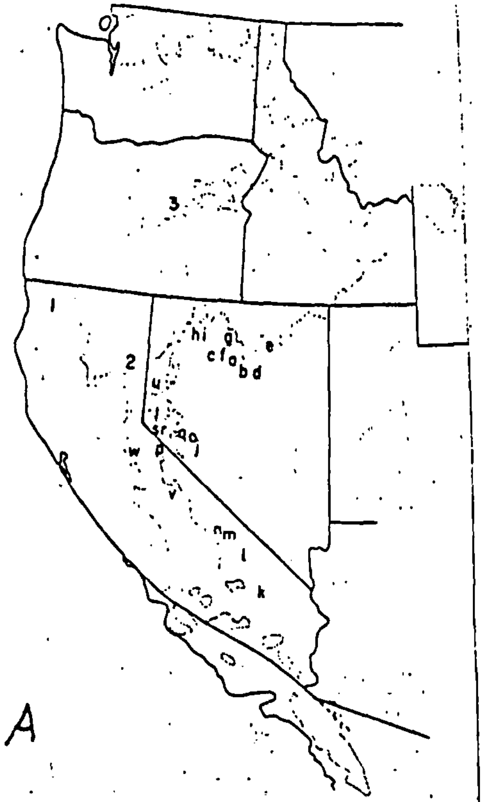


Figure 3. Locations of studied suites. Map constructed as described in caption for Figure 1. a. Scott Canyon Formation, Antler Peak quadrangle. b. Shwin Formation, Shosone Range. c. Valmy Formation, Water Canyon, Sonoma Range. d. Valmy Formation, Shoshone Range. e. Schoonover Formation, Jacks Creek, Independence Range. f. Pumpernickel Formation, Sonoma Range. g. Farrel Canyon Formation, Osgood Mountains, h., i. Happy Creek Formation and overlying Triassic(?) rocks, Jackson Mountains. j. Excelsior Formation, Hawthorne area. k. Soda Mountains Formation, Soda Mountains. l. Warm Springs Formation, Butte Valley. m. Unnamed unit, northern Inyo Mountains. n. Unnamed unit, Alabama Hills. o. Gillis Formation, Hawthorne area. p. Unnamed unit (Excelsior?), Bodie Hills. q. Wassuk Range, no samples. r. Unnamed unit southwest of Yerington. s. Triassic-Jurassic, Pine Nut Range. t. Unnamed units of Triassic-Jurassic(?) age near Genoa and Carson City. u. Peavine sequence, Reno. v. Volcanic rocks of Shadow Creek and Mammoth Crest, Devil's Postpile area. w. Unnamed Triassic-Jurassic(?) unit in San Andreas quadrangle. 1. Klamath Mountains. 2. Devonian island-arc assemblage of northern Sierra Nevada (from Hietanen, 1973). 3. Triassic keratophyres of central Oregon inlier (from Dickinson and Vigrass, 1964).

P. 1914

Figs A & B "Paleozoic and Lower Mesozoic Volcanism and Continental Growth in the Western United States" Rogers J.W. et al, GSA Bull. Vol. 85 # 12, 1974.

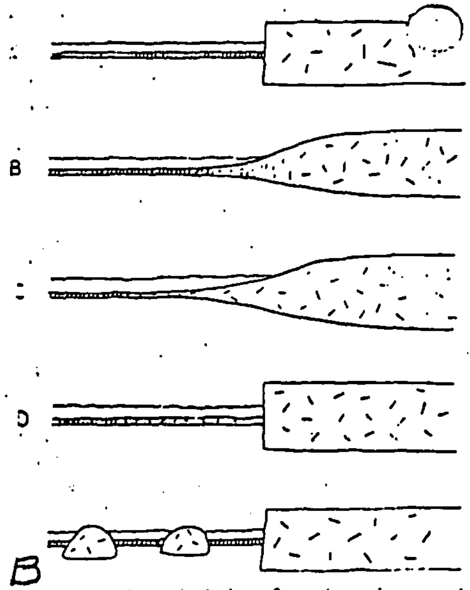


Figure 2. Hypothetical configurations along continental margins. Granite and basalt patterns represent continental and oceanic crust, respectively, in relation to sea-level lines. Gradational transition between continental and oceanic material is shown in section B.

P. 1916

Figs C & D "Silurian (Llandovery) Downdropping of the western Margin of North America" J.G. Johnson & E.C. Potter, Geology, Vol. 3, No. 6, 1975

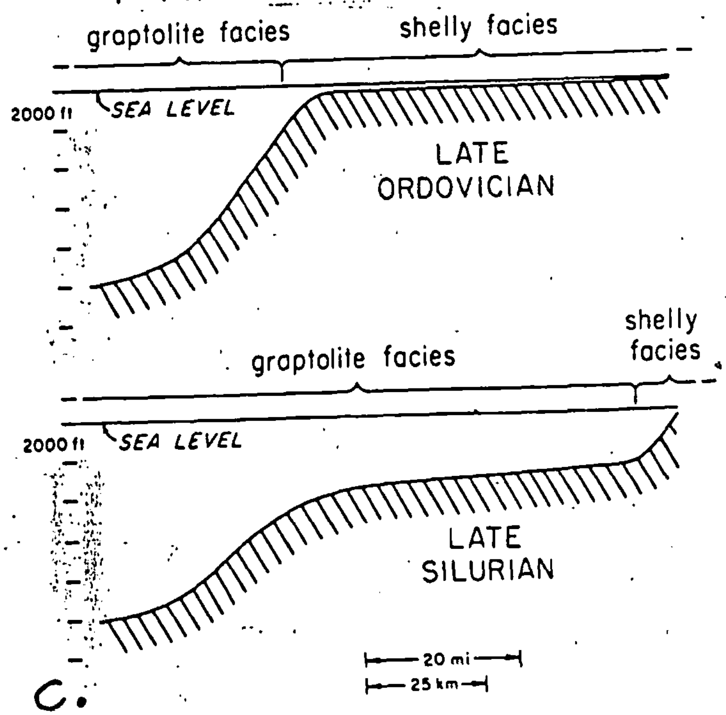
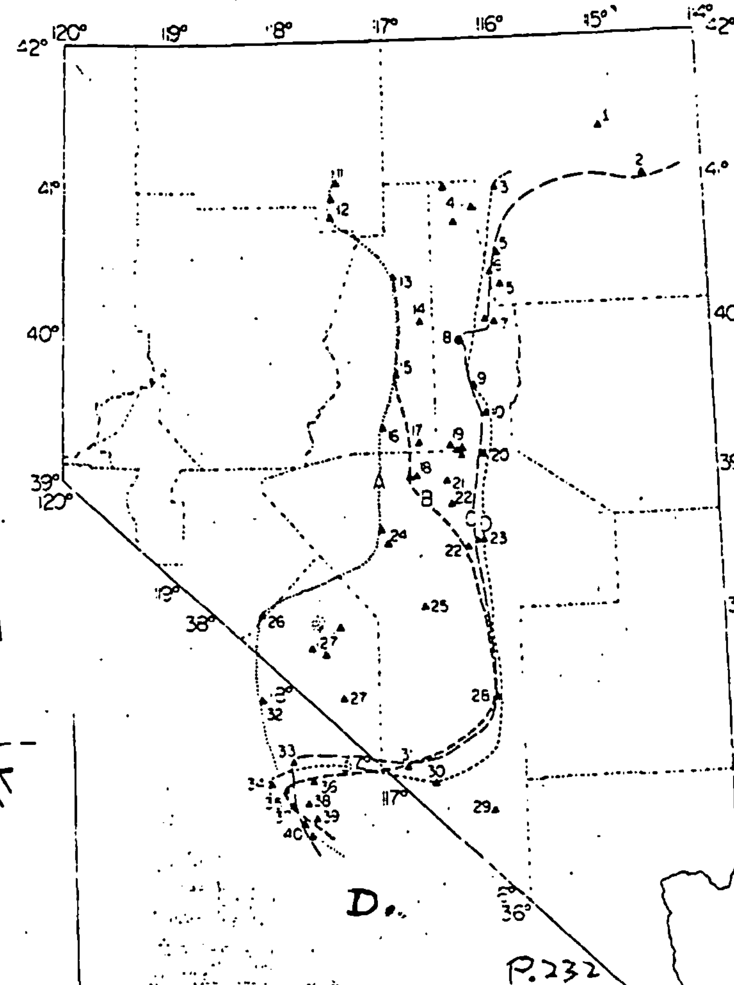


Figure 2. Two east-west profiles at 40th parallel, to show altered bathymetry due to downdropping of 64-km-wide strip of continental margin. B and C are plotted in Figure 1. We recognize that graptolite and shelly facies ordinarily overlap. Shelly facies is plotted here to include its supposed maximum extent and depth.

P. 233



P. 232

Figure 1. Index map of Nevada, showing county boundaries. A = minimum westward extent of Lower Cambrian cross-bedded quartz sandstone, B = minimum westward extent of Upper Ordovician-Lower Silurian shelf carbonate, C = boundary between graptolite facies and shelf carbonate facies of Upper Silurian rocks, D = boundary between open-marine limestone facies and shelf dolomite facies of lower Middle Devonian rocks. References for numbered localities: 1–Johnson, 1970; 2–Thorman, 1970; 3–Evans and Ketner, 1971; 4–Roberts and others, 1967; 5, 6–Smith and Ketner, 1972; 7–Roberts and others, 1967; 8–Winterer and Murphy, 1960; 9–Merriam, 1940; 10–Merriam, 1963; 11–Ferguson and others, 1952; 12–Ferguson and others, 1951; 13–Gilluly and Gates, 1965; 14–Gilluly and Masursky, 1965; 15–Stewart and McKee, 1968; 16–Stewart and Palmer, 1967, and Washburn, 1970; 17–McKee and Ross, 1969; 18–Kay and Crawford, 1964; 19–Merriam, 1973; 20-23–Kleinhampl and Ziony, 1967; 24–Ferguson and Cathcart, 1954, and Kleinhampl and Ziony, 1967; 25–Kleinhampl and Ziony, 1967; 26–Ferguson and others, 1954; 27–Albers and Stewart, 1973; 28–Cornwall, 1972; 29–Burchfiel, 1964; 30–Cornwall, 1972; 31–Cornwall, 1972, and Cornwall and Kleinhampl, 1964; 32–Nelson, 1962; 33–Ross, 1967a; 34, 35–Stevens and Ridley, 1974; 36–Ross, 1967b; 37, 38–Ross, 1967a; 39–McAllister, 1956, and Ross, 1967a; 40–Ross, 1967a, and Stevens and Ridley, 1974.

11/5/79

1/24/79

Ordovician

Suggested readings for those on limited time.

1 Best all-around paper: Reuben J. Ross Jr., "Ordovician Sedimentation in the western United States", in The Ordovician System, University of Wales and National Museum of Wales, 1976. In readings drawer.

2. Best for Rocky Mountains: Norman H. Foster. "Ordovician System" in Geologic Atlas of the Rocky Mountain Region, Rocky Mountain Association of Geologists, Denver. Colo. 1972, p. 76-86

3 Best on the mid-Ordovician quartzites: Robert Q. Oaks Jr. and others, "Summary of Middle Ordovician Stratigraphy and Tectonics, northern Utah, Southern and Central Idaho": in Twenty-Ninth Annual Field Conference -1977 Wyoming Geological Association Guidebook. Copy in readings drawer

4. Best on the Cambrian-Ordovician boundary: James F. Miller, "Definition of the Cambrian-Ordovician Boundary" in Upper Cambrian to Middle Ordovician Conodont Faunas of Western Utah, Dept of Geography and Geology. Southwest Missouri State University, Geoscience Series, No. 5, 1978, p. 14 f.. Copy in readings drawer.

5. Best (or at least good) on eugeosynclinal facies: Ralph J Roberts, Preston E. Hotz, James Gilluly and H.G. Ferguson, "Paleozoic Rocks of North-Central Nevada": A.A.P.G Bull. vol. 42, p 2813 f. Copy in readings drawer.

W. L. Stokes, 1979

1/26/79

Ordovician Assignment

From now on there will be one problem per period but an additional quarter-long assignment to be explained this week.

Do ONE of the following

1 Discuss the evidence for the Tooele Arch and its influence on Great Basin Stratigraphy.

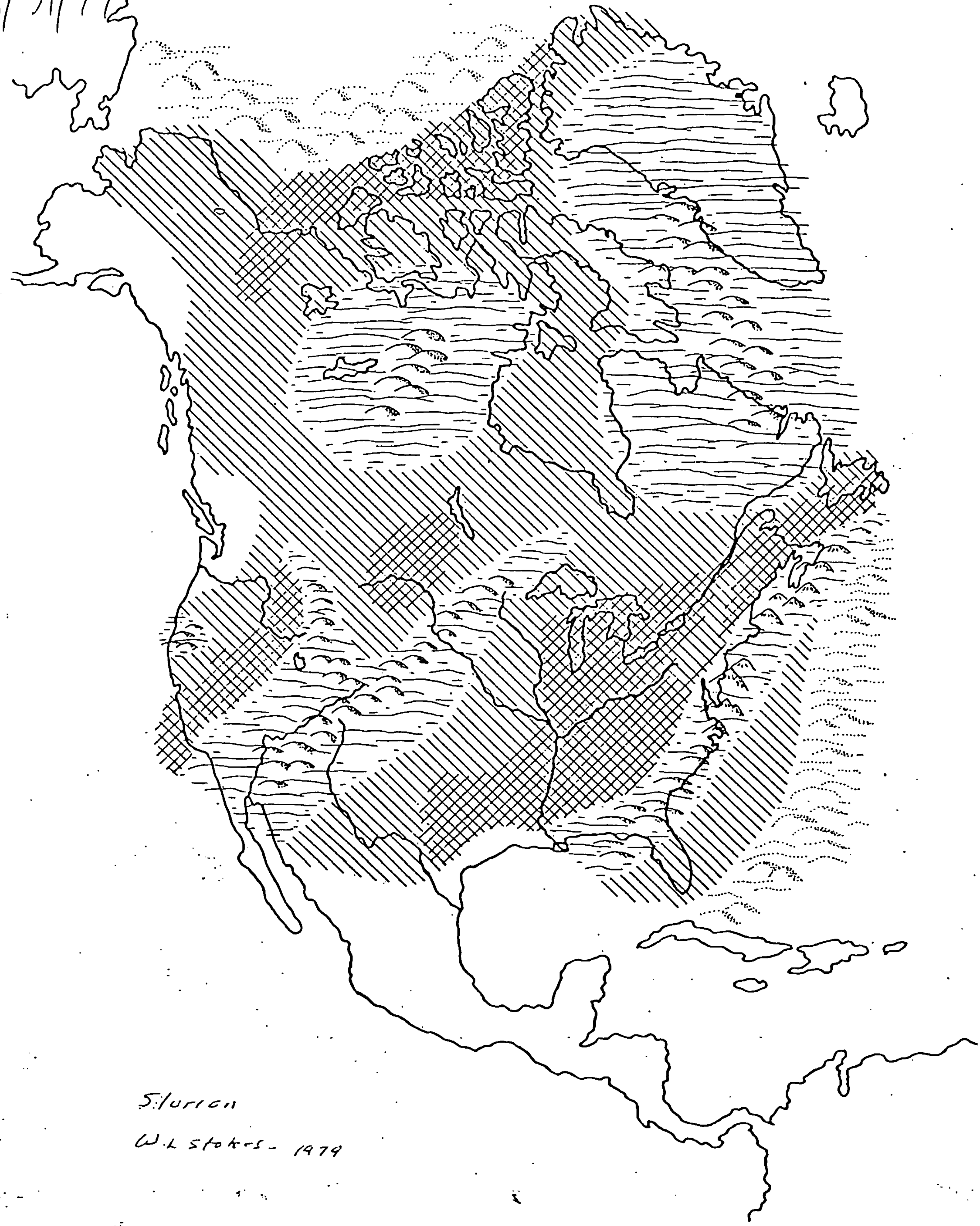
See Bick, 1966, Geology of the Deep Creek Mtns: Utah G&M Survey Bull. 77. In reference set and library. (p. 58)

2. Discuss the evidence for the Southern Oklahoma aulacogen and its effect on Ordovician rocks of Colorado.

See paper by Ross in The Ordovician System, 1976. 2 reprints in references, p. 83

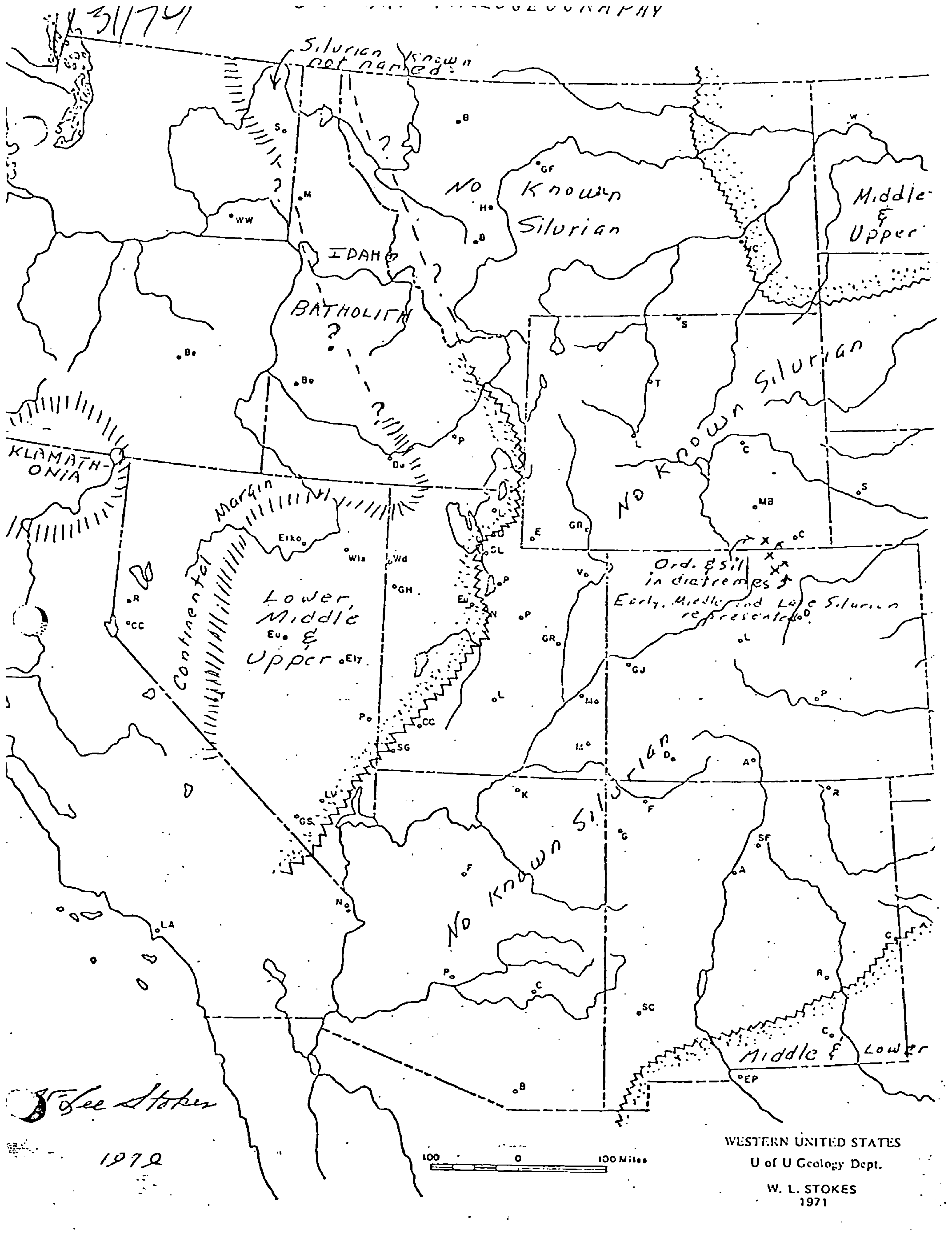
W. Stokes
Jan 1979

1/31/79



Silurian

W. L. Stokes - 1979



PALEOGEOGRAPHY
31/74
Silurian known not named
No Known Silurian
Middle & Upper
IDAHO
BATHOLITH
No Known Silurian
KLAMATH-ONIA
Continental Margin
Lower, Middle & Upper
Ord. & Sil in diatremes
Early, Middle and Late Silurian represented
No Known Silurian
Middle & Lower
Lee Stokes
1972
100
0
100 Miles
WESTERN UNITED STATES
U of U Geology Dept.
W. L. STOKES
1971

1/31/79

Generalizations

Paleozoic Stratigraphy of the Western Interior

SILURIAN

I. Historical

A. The Silurian was named by Roderick Murchison in 1835 after the ancient tribe of Silures who once inhabited the Welsh Borderland. European geologists have not accepted this system unanimously; the term Gotlandian is used by many Continental geologists for what is Silurian in Great Britian.

B. In America, Lower, Middle, and Upper series are recognized with the place names, Alexandrian (from Alexander Co., Illinois) Niagaran (Niagara Falls) and Cayugan (Cayuga, New York) as equivalents. Latest authoratative survey of the Silurian says this: "It is recommended that North Americans discontinue the futile efforts to use the New York section as a standard for purposes of correlation." Nevertheless no widely accepted substitute is in use.

II. Tectonics and Paleogeography

A. The major tectonic elements - Craton, Transcontinental Arch, and Cordilleran Geosyncline - that were in evidence during the earlier Paleozoic periods continued to dominate sedimentation during the Silurian. The concept of miogeosyncline and eugeosyncline is not as appropriate as for either Ordovician and Devonian as there are no volcanic derivatives in the Silurian.

B. Williston Basin continues to expand and deepen.

C. A variety of sediments, some conglomeratic, in or near the Klamath Mountains (Cal-Oregon) indicate possible enlargement of Klamathonia.

D. Wasatch Line, modified by the Uinta Peninsula-Tooele Arch, marks the eastern edge of preserved Silurian in Utah-Idaho.

E. The possibility of an island-like land mass in the vicinity of the present Raft River Range (Raftriveria?) is suggested.

F. That Ordovician and Silurian seas and sediments once covered wide areas of the Transcontinental Arch is proved by the preservation of rocks of these ages in at least 5 small diatremes in the Precambrian of the Front Range and Laramie Range in Wyoming and Colorado. (See GSA Bull. vol. 80, no. 1)

III. Lithology and Sedimentation

A. The Silurian has an estimated duration of only 20 million years. As a consequence the volume of sediment is the smallest of any system in the Western Interior. It is probably less than 10% of that of the Cambrian. There are relativley good exposures in the Great Basin but none in Arizona, and very little in New Mexico, Colorado and Wyoming. Most of the Williston Basin is underlain by Silurian.

B. Rock types are mainly dolomite (dolostone) in the miogeosyncline and black shale in the transition belt and eugeosyncline. Estimated percentage of rock types: dolomite 75%, limestone 10%, black shale 10%, chert 5%.

C. Stratigraphy reflects very shallow-water deposition during a time of tectonic quiet and slow downsinking in the Cordilleran Geosyncline. Few formal names are necessary to designate the identifiable units, 5 major formations of which the Laketown Dolomite is most widespread, are named. Perhaps as many as 18 members can be recognized. Average thickness is about 1000-1500 feet. Williston Basin deposits (Interlake Group), not exposed in the U.S., are likewise thin and dolomitic.

D. The most obvious characteristic of the Silurian system is the predominance of dolomite. How and why this rock type came into existence is one of the major problems of Silurian stratigraphy. The associated chert is varied in character and also not well-studied or understood.

IV. Biostratigraphy

The Silurian faunas of the Western Interior are vastly inferior to those of the eastern U.S. because of the small volume of sediment, the unfavorable shallow water environments and the extensive dolomitization. Approximate order of decreasing abundance of animal groups as represented in the Western Interior is: Brachiopods, corals, graptolites (in black shales, monograptids mainly). Minor groups include bryozoans, trilobites, ostracods and eurypterids. No land plants have yet been found in the western interior.

Of the above list the corals and brachiopods are frequently found to be silicified in sufficient numbers to be useful as guide fossils. Both groups have been used in zonation schemes. Graptolites are unexcelled in the shale facies. Because of lack of fossils and difficulties in preparing those which do occur many miscorrelations of Silurian dolomites have been made.

V. Boundaries

The Ordovician-Silurian boundary is determined by faunal evidence. It lies in the midst of generally unfossiliferous dolomite beds with many diastems and unconformities. That it is difficult to locate is shown by the fact that for many years and almost without exceptions stratigraphers have been including several hundred feet of Late Ordovician rocks in the Laketown Dolomite, a formation regarded as Silurian in age.

The Silurian-Devonian boundary is likewise located in a dolomite sequence in which diagnostic fossils have not been found. Although the Water Canyon Formation is usually considered to be Early Devonian no paleontologic evidence for the age of the lower part has been found in the northern Great Basin. Late Silurian fossils are known in the Sevy Formation of southern Nevada which is lithologically similar to the Water Canyon. The boundary problem is currently unsolved.

VI. Economic

Dolomite is present in unlimited quantity but apparently unexploited at present. Presence of chert is deleterious.

The Silurian formations are important host rocks in the beryllium-uranium-fluorite ore bodies in the Thomas Range, Juab County, Utah.

VII. Scenic

The Laketown Dolomite forms imposing, dark rugged cliffs in Logan Canyon but only a geologist would be impressed

Bibliography

1972 "Geologic Atlas of the Rocky Mountain Region" Rocky Mountain Association of of Geologists, Denver. The Silurian Chapter is by Frank K. Gibbs, p. 86-89.

1970 "Correlation of the North American S-lurian Rocks" by W. B. N. Berry and A. J. Boucot, Geological Society of America, Special Paper 102, 289 pages.

W. Stokes 1979

1/31/79

Silurian Assignment

The Silurian Period marks the culmination of the Great Dolomite Age of the Great Basin. Read or scan the Stoke's compilation, the newer excerpts from Geology and the U.S.G.S Roberts Mountain Professional Paper. Know something of Osmond's paper on Great Basin dolomitic formations* and any further readings you care to make on the problem:

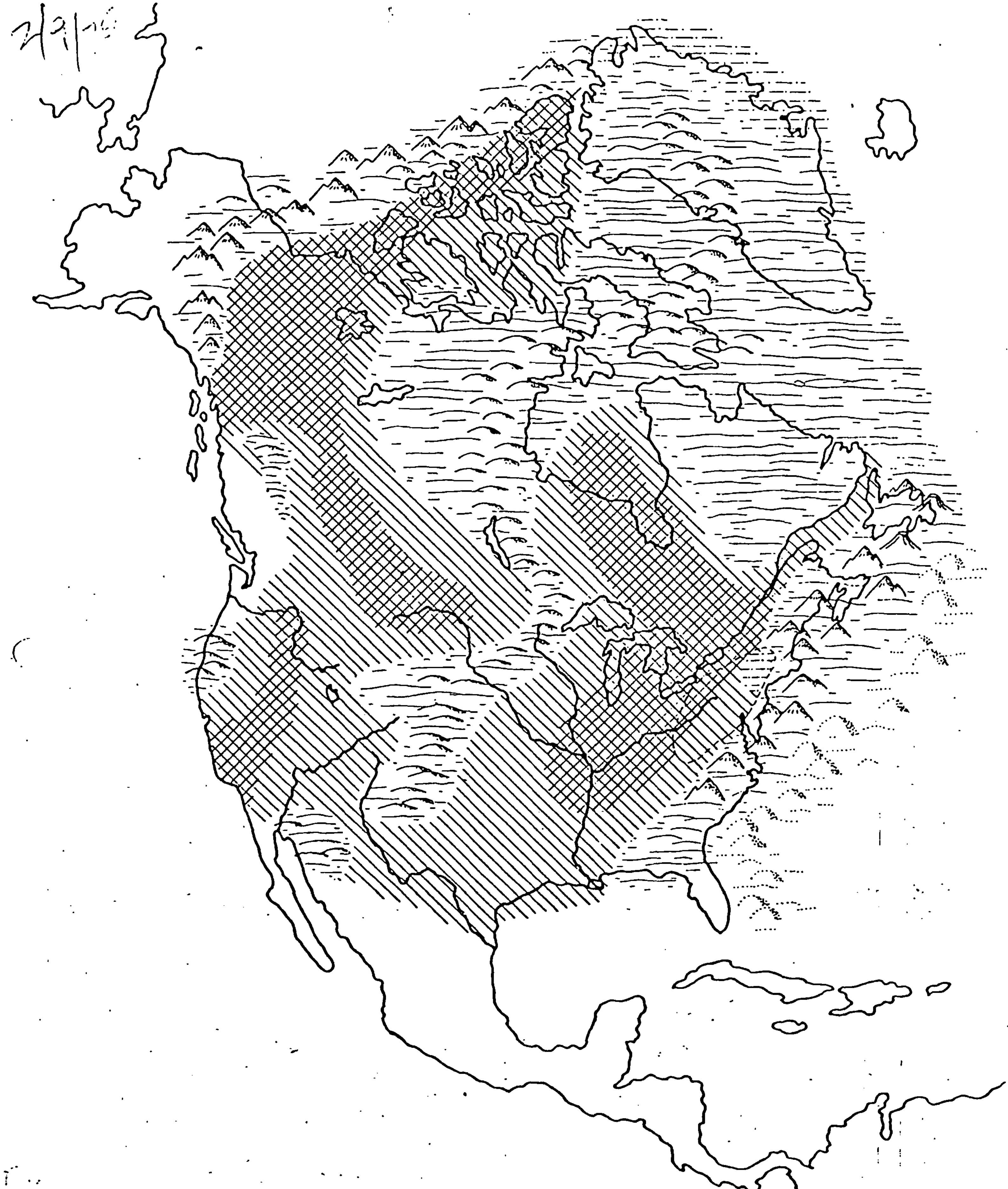
Now:

Write a 2-3 page summary of your own conclusions about the origin of dolomite in Great Basin Paleozoic formations. Agree or disagree or remain neutral but give a critical opinion on the matter.

There will be no alternate problem for this system.

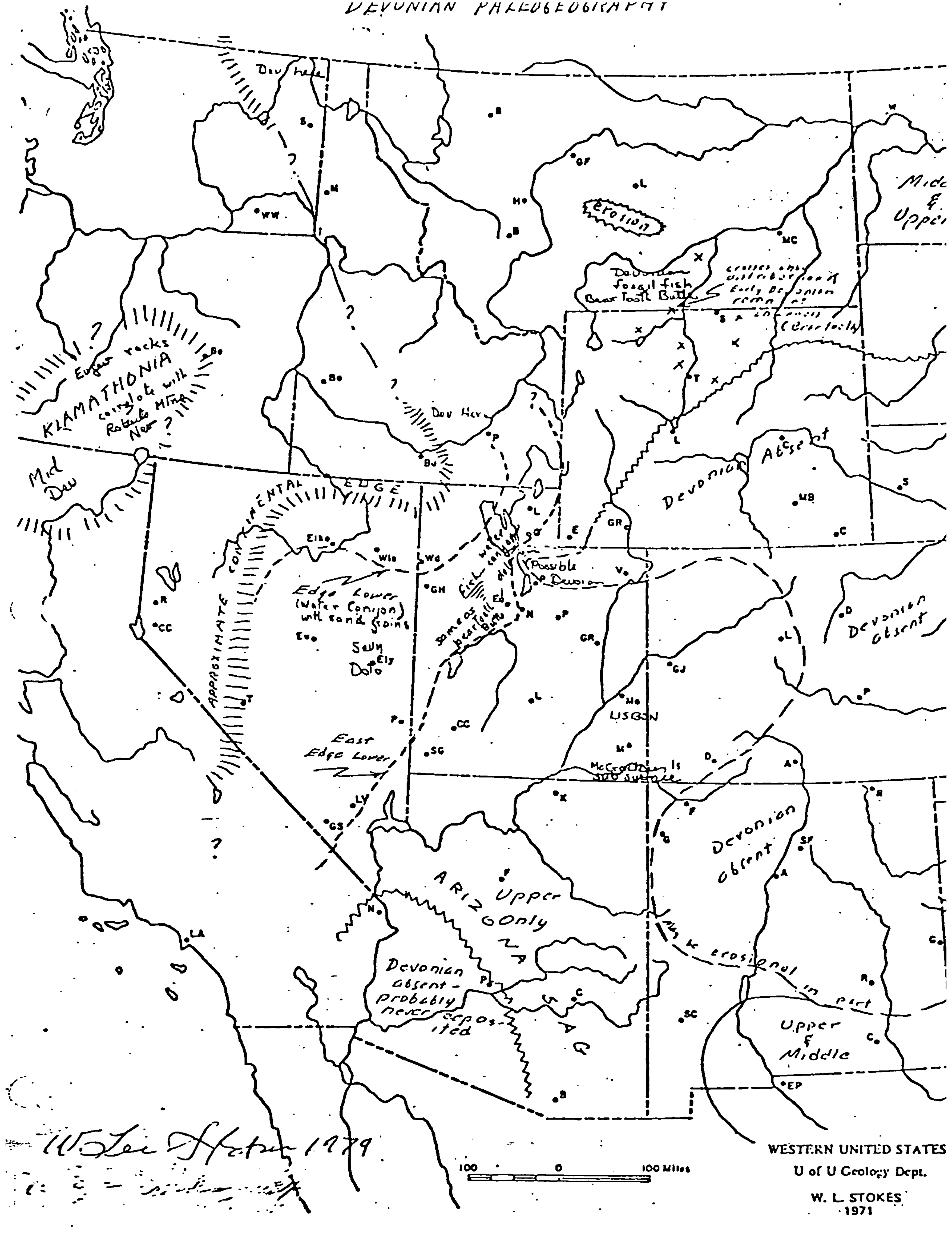
W Stokes, Feb, 1979

* AAPG Vol: 38 #9



DEVONIAN

DEVONIAN PALEOGEOGRAPHY



2/9/79

Generalizations

Paleozoic Stratigraphy of the Western Interior

DEVONIAN

I. Historical

A. The Devonian was proposed jointly by Sedgwick and Murchison in 1839, the name being taken from Devon, England. This type area is largely continental Old Red Sandstone and is not as fossiliferous as corresponding marine rocks in Belgium and Germany. Consequently the latter areas have become the standard for Europe and much of the world.

B. Very thick Devonian formations (more than 10,000 feet) are found in New York. This area has been intensively studied and correlates fairly well with Europe. Possibly the ancestral Atlantic Ocean was closed in the Devonian and Europe and North America were in close proximity. Unfortunately, and mostly because of the slection of unpronouncable Iroquois Indian names for stages the N.Y. section is not being widely used as a reference. The best North American sections appear to be in Canada.

II. Tectonics and Paleogeography

A. Cordilleran Geosyncline, Canadian Shield and Williston Basin dominate the tectonic framework.

B. Transcontinental Arch still much in evidence but now broken by two "sags", one crossing Colorado in a northwesterly direction is the older, the other crossing Arizona with the same trend is a Devonian development, "Mazatzal Land" an island-like remnant of the Mogollon-Defiance positive area is part of the Transcontinental Arch.

C. Williston Basin deepens and expands - now with important evaporites.

D. Wasatch Line is a definate hinge or transition belt.

E. Western continental margin continues to be poorly defined but the section in the Klamath Mountains contains Devonian as does the Metaline Falls district, northeastern Washington.

F. Wyoming shelf was flooded and received a fairly consistent blanket of carbonate, the well-known Jefferson Dolomite

G. The so-called Stansbury Disturbance in north-central Utah records the first mountain building in the state since the Precambrian.

H. The Antler Orogenic Belt crossing Nevada in a north-northeast direction became very active near the end of the Devonian and Early Mississippian. (Roberts, AAPG, Dec, '58).

I. Volcanic activity gives characteristic eugeosynclinal facies in the transition belt and farther west.

III. Lithology and Sedimentation

A. The Devonian Period had an estimated duration of 60 million years and the System has a volume commensurate with this - more sediment than the Silurian, less than the Cambrian. There are many extensive exposures in the Great Basin and Rocky Mountains. Especially noteworthy are transition and eugeosynclinal exposures in Nevada (Roberts Mountains, etc.) Extensive ~~exposures~~ volumes in Williston Basin, feather edge in Black Hills.

B. Carbonates predominate especially in Early and Middle epochs. Black shale to the extreme west and varied clastics - shale, sandstones, conglomerate over much of the northern Great Basin, western Wyoming and Montana late in the period. A phase perhaps of the famous Chatanooga Shale. Estimated percentage of rock types: dolomite, 35%; limestone, 40%; clastics, including sizeable volumes of quartzite, 15%; chert and volcanic derivatives 10%. Although there is little development of reefs in the Western Interior, the Devonian was one of the great reef-forming periods of all time.

C. Stratigraphy reflects shallow water deposition, very like that of the Silurian and dominated by dolomite formations (Sevy, Water Canyon) in the miogeosyncline early in the period. Following this the water seems to have deepened with mixed carbonate deposition (Jefferson, Simonson). Lastly, reflecting tectonism and perhaps uplift of the Transcontinental Arch came clastic beds - some red, over wide areas.

D. Evaporites including potash are of great commercial importance (Prairie Evaporite Formation) in the Devonian of the Williston Basin, chiefly in Saskatchewan, extends into northern North Dakota.

IV. Biostratigraphy

A. The Devonian faunas are numerous, diagnostic, and very diversified. The limestone facies yield abundant silicified and unsilicified specimens. Estimated abundance of animal groups in decreasing importance: Brachiopods, corals ~~conodonts~~, cephalopods and other molluscs, bryozoans, ostracods, conodonts, trilobites, vertebrates, echinoderms. Zonations have been based on many groups - estimated importance in order of decreasing use are brachiopods, conodonts (increasing steadily in use) and corals. Almost too many schemes are in use.

B. First known land plants of western Interior appear (Beartooth), spores and pollen may become important in correlation.

C. Rather abundant and important fish faunas (Water Canyon, Jefferson).

D. By far and away the chief student of Devonian marine faunas of the Great Basin is J. G. Johnson, currently at Oregon State University. A fantastic producer and publisher. He uses the European names and will probably prevail insofar as West is concerned.

V. Boundaries

See Silurian handout for lower boundary

The Devonian-Mississippian boundary has not been located to everyone's satisfaction but will eventually be set at some horizon of continuous sedimentation. Possibly the Great Basin might become a future standard for this important datum plane. According to faunal information, chiefly conodonts, it lies in a clastic interval (Bakken, Chattanooga, Pilot, Percha, Englewood). Hopefully conodonts may largely solve the problem and place western firmations in a world-wide scheme.

VI. Economics

A. The Devonian is a significant oil objective in the Williston Basin, second in importance after the Mississippian.

B. Oil and gas producer in the Lisbon Field of San Juan County, Utah; minor production in N.E. Arizona. Helium in N.W. New Mexico.

C. Potash in Canada, Williston Basin (Prairie Evaporite Formation); bad news for U.S. Pennsylvanian & Permian producers.

D. Host rock for metal deposits in Utah, Nevada and California.

VII. Scenic

Devonian remnants occur beneath the Redwall in the Grand Canyon but are obscure. Cliff-former in Logan Canyon, Nevada Limestone is prominent in many Nevada Ranges. One of the Presidential Ranges (Jefferson) is a prominent type area in Montana.

Bibliography

1942, "Correlation of the Devonian Sedimentary Formations of North America," G. A. Cooper, et. al., Bull. G.S.A. vol. 53, p.p. 1729-1794.

1967, "International Symposium on the Devonian System," D. H. Oswald, Ed. vol. 1, 1055 p.; vol. 2, 1377 p., Published by Alberta Society of Petroleum Geologists, Calgary, Alberta, Canada.

1972, "Geologic Atlas of the Rocky Mountain Region," Rocky Mountain Assoc. of Geologists, Denver, Devonian chapter by Donald L. Baars, p. 90-99. He uses European scheme also.

Wm. Lee Stokes, 1974

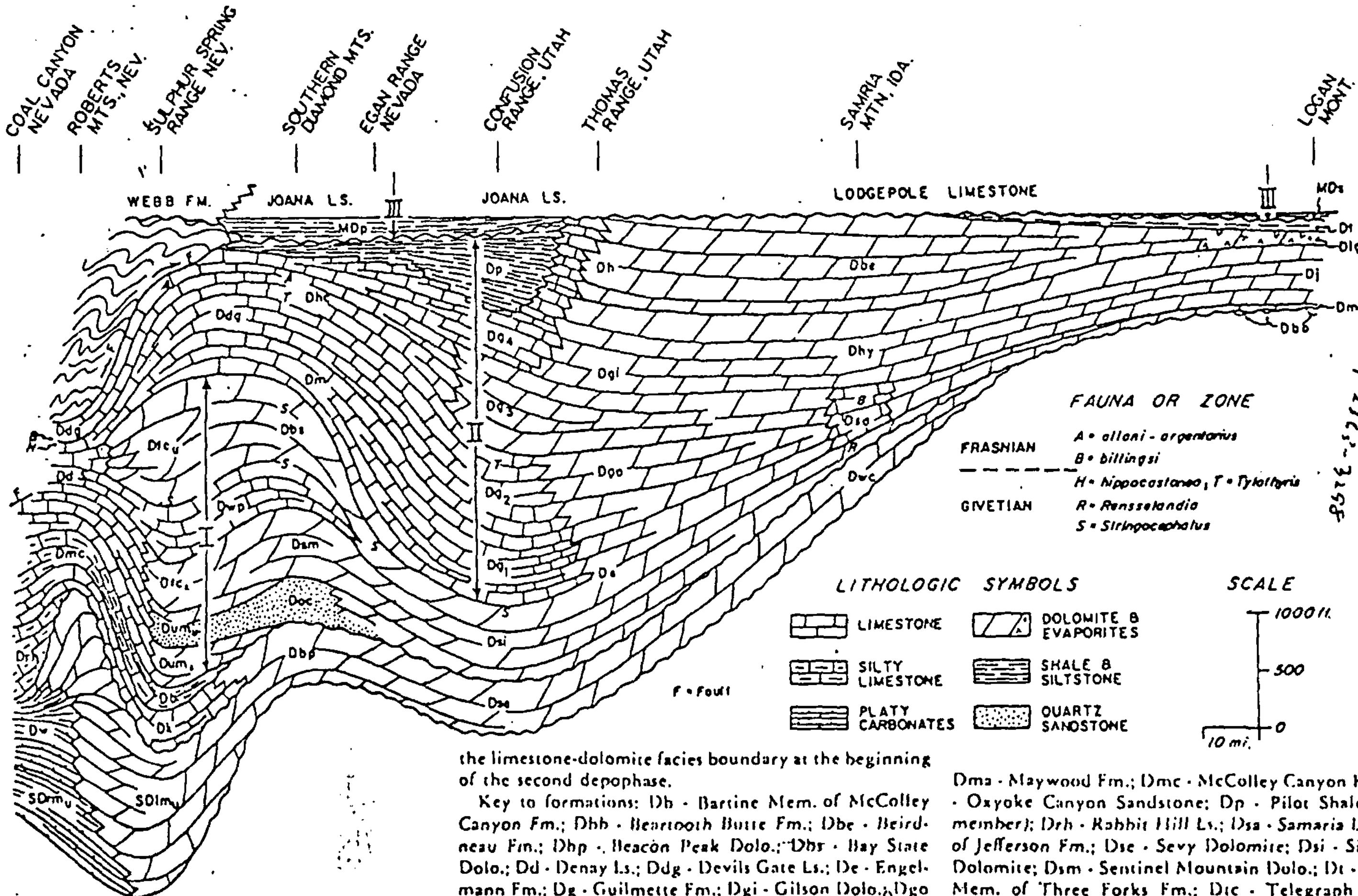


Figure 5. Cross section of Devonian and basal Mississippian rocks from central Nevada to southwestern Montana. Depophases are indicated by Roman numerals. Note the great eastward shift of depocenter and of the limestone-dolomite facies boundary at the beginning of the second depophase.

Key to formations: Db - Bartine Mem. of McColley Canyon Fm.; Dbb - Beartooth Butte Fm.; Dbe - Beirdneau Fm.; Dbp - Beacon Peak Dolo.; Dbs - Bay State Dolo.; Dd - Denay Ls.; Ddg - Devils Gate Ls.; De - Engelmann Fm.; Dg - Guilmette Fm.; Dgi - Gilson Dolo.; Dgo - Goshoot Fm.; Dh - Hanauer Fm.; Dhc - Hayes Canyon Mem. of Devils Gate Ls.; Dhy - Hyrum Dolo. Mem. of Jefferson Fm.; Dj - Jefferson Fm.; Dk - Kobeh Mem. of McColley Canyon Fm.; Dlg - Logan Gulch Mem. of Three Forks Fm.; Dm - Meister Mem. of Devils Gate Ls.; Dma - Maywood Fm.; Dmc - McColley Canyon Fm.; Doc - Oxyoke Canyon Sandstone; Dp - Pilot Shale (lower member); Drh - Rabbit Hill Ls.; Dsa - Samaria Ls. Mem. of Jefferson Fm.; Dse - Sevy Dolomite; Dsi - Simonson Dolomite; Dsm - Sentinel Mountain Dolo.; Dt - Trident Mem. of Three Forks Fm.; Dtc - Telegraph Canyon Dolo.; Dum - Union Mountain Fm.; Dw - Windmill Ls.; Dwc - Water Canyon Fm.; Dwp - Woodpecker Ls.; MDs - Sappington Sandstone; MDp - Pilot Shale (upper member); SDlm - Lone Mountain Dolo.; SDrm - Roberts Mountains Fm.

J.G. Johnson, Timing and Coordination of Orogenic, Epeirogenic and Eustatic Events, G.S.A. Bull. 82, P. 3263-3298

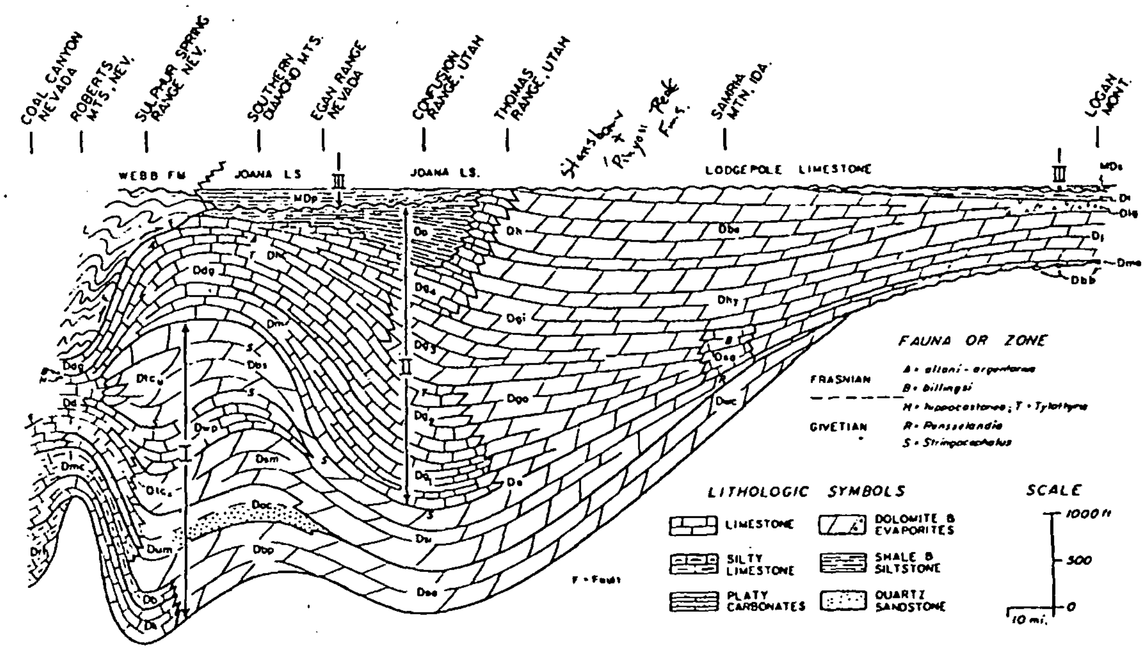
171

4/24/79

Journal of Paleontology, v. 50, no. 6, p. 1113-1128, 2 pls., 10 text-figs., November 1976

LATE EARLY DEVONIAN BRACHIOPODS AND BIOFACIES FROM CENTRAL NEVADA

J. G. JOHNSON and G. W. KENDALL
Oregon State University, Corvallis, 97331 and Teal Petroleum Company, Bakersfield, California, 93309



Text-fig. 5—Cross section of Devonian and basal Mississippian rocks from central Nevada to southwestern Montana. Depophases are indicated by Roman numerals. Key to formations: Coils Creek Limestone is unlabeled above Dmc in the Roberts Mountains. Db–Bartine Mbr. of McColley Canyon Fm.; Dbb–Beartooth Butte Fm.; Dbe–Beirdneau Fm.; Dbp–Beacon Peak Dolo.; Dbs–Bay State Dolo.; Dd–Denay Ls.; Ddg–Devils Gate Ls.; De–Engelmann Fm.; Dg–Guilmette Fm.; Dgi–Gilson Dolo.; Dgo–Goshoot Fm.; Dh–Hanauer Fm.; Dhc–Hayes Canyon Mbr. of Devils Gate Ls.; Dhy–Hyrum Dolo. Mbr. of Jefferson Fm.; Dj–Jefferson Fm.; Dk–Kobeh Mbr. of McColley Canyon Fm.; Dlg–Logan Gulch Mbr. of Three Forks Fm.; Dm–Meister Mbr. of Devils Gate Ls.; Dma–Maywood Fm.; Dmc–McColley Canyon Fm.; Doc–Oxyoke Canyon Sandstone; Dp–Pilot Shale (lower mbr.); Drh–Rabbit Hill Ls.; Dsa–Samaria Ls. Mbr. of Jefferson Fm.; Dse–Sevy Dolomite; Dsi–Simonson Dolomite; Dsm–Sentinel Mountain Dolo.; Dt–Trident Mbr. of Three Forks Fm.; Dtc–Telegraph Canyon Dolo.; Dum–Union Mountain Fm; Dwc–Water Canyon Fm.; Dwp–Woodpecker Ls.; MDs–Sappington Sandstone; MDp–Pilot Shale (upper mbr.).

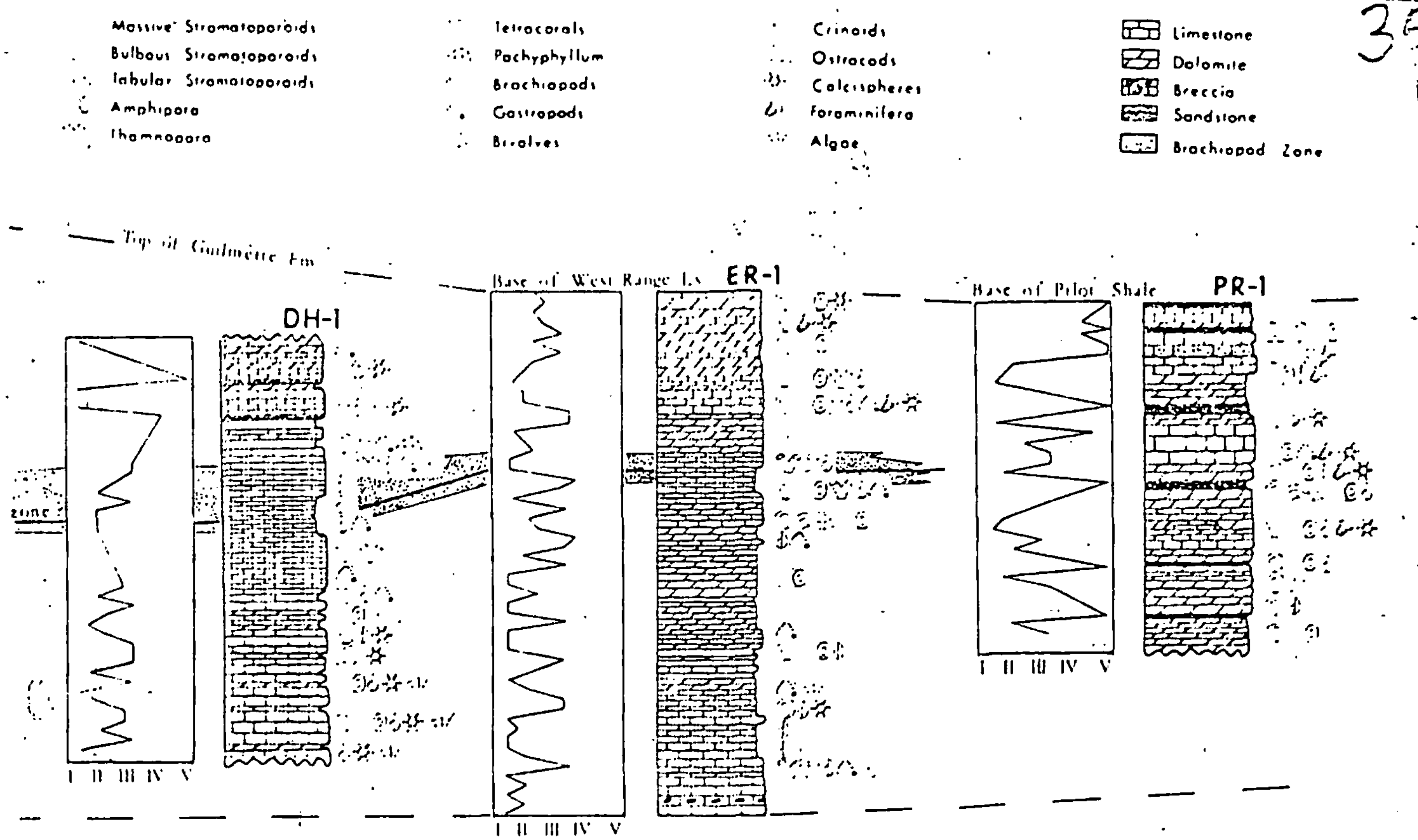
(tongue) and upper Bartine tongue in the Sulphur Spring Range. The sum of this is that the quartzite called Oxyoke Canyon is entirely younger than the *pinyonensis* Zone of the Bartine Limestone rather than a facies of it as has been supposed by Merriam (1963, 1974). It must be emphasized that all of Merriam's cross sections, prepared over the years (1963, p. 52, fig. 8; 1974, p. 16, fig. 7) to show correlations between the platform dolomite sequence (which includes the Oxyoke Canyon Sandstone) and the limestone belt to the west, were based on the false assumption that the Oxyoke Sandstone is a facies of the *pinyonensis* Zone (*i.e.*, the Bartine or "shaly lower Nevada" of Merriam).

There is now no reason to suspect that the base of the Sevy (= Beacon Peak) is regionally any older than the base of the Kobeh Member of the McColley Canyon Formation, although that conclusion has been suggested (Nichols & Silberling, 1975). Physical evidence for an unconformity beneath both units has been reported and the oldest rocks above this unconformity, now included in Kaskaskia Depophase I, are of latest Helderbergian age at the Devonian continental shelf margin in central Nevada. Wherever a basal sandstone and transgressive shallow water carbonates occur with fossils they are of late Helderberg to Oriskany age and correspond in sedimentary mode and timing with the base of the Kaskaskia sequence originally defined elsewhere in North America (Sloss, 1963).

SYSTEMATIC PALEONTOLOGY

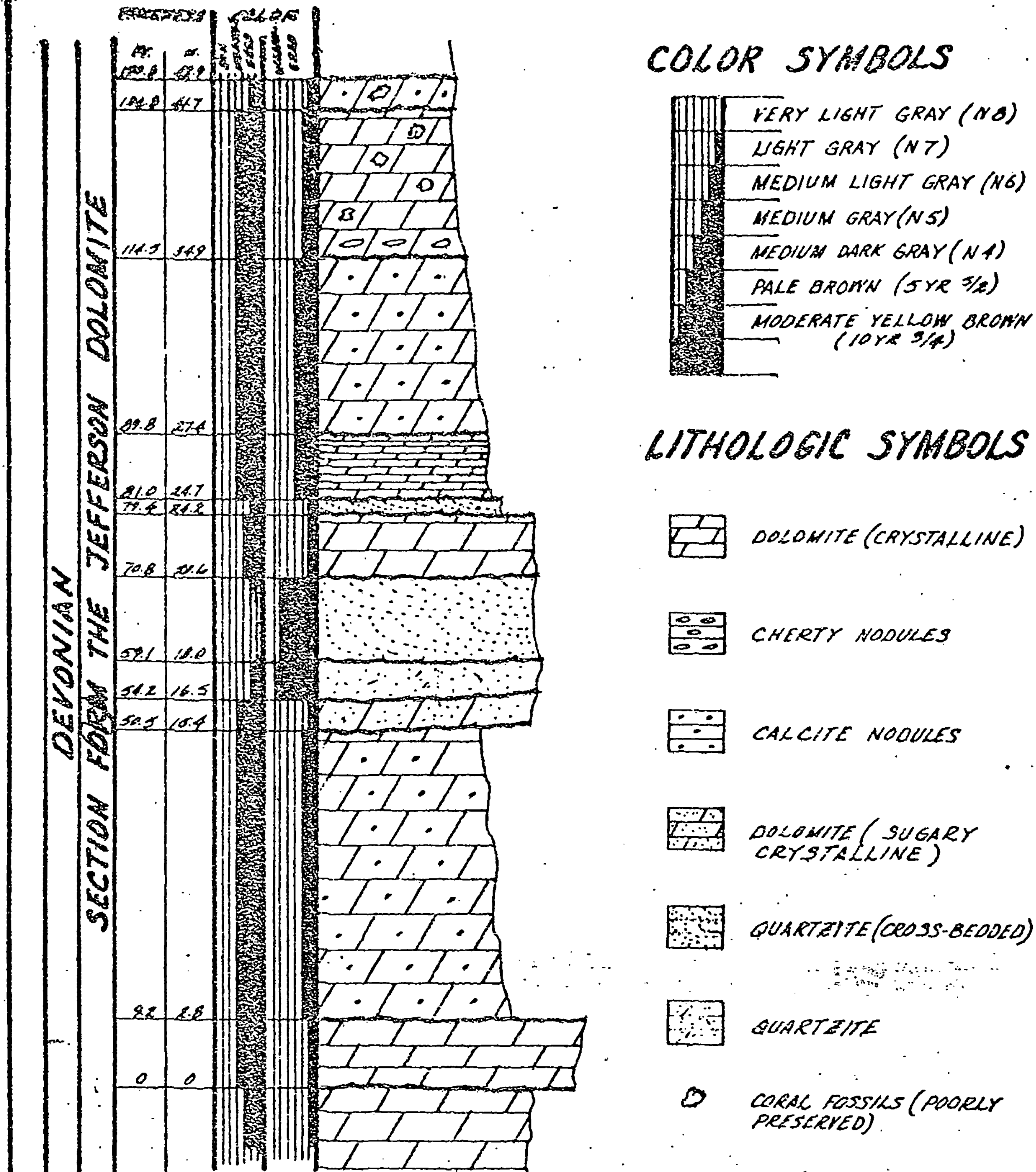
Phylum Brachiopoda
Family Gypidulidae Schuchert & LeVene
Subfamily Gypidulinae Schuchert & LeVene
Genus Pentamerella Hall

Discussion.—The type species of *Pentamerella* is *P. arata* (Conrad) from the Schoharie



TEXT-FIGURES 5a and 5b.—Energy fluctuations, faunal distribution, and regional correlation of the Guilmette Formation in the study area.

9/2/1174 PARTIAL SECTION-JEFFERSON DOLOMITE



SECTION FROM THE JEFFERSON DOLOMITE

Section which was measured is located 13.4 miles north and 6.2 miles east of Delle, Utah. Location was scaled from "Geologic Map, Lakeside range, Tooele County, Utah" by John C. Young.

November 1977 Measured by: Byron Parker, Lowell Howlett

Submitted by: Byron Parker

2/9/79

Devonian Assignment

With not more than 3 average pages complete ONE of the following. Include a one page illustration or a page of illustrations of your own making.

A. Discuss and illustrate the Stansbury Uplift (or Stansbury Disturbance). Refer to Rigby. G.S.A. Bull. v. 70, p. 207-218 and Utah Geological Society Guidebook No. 13

B. Discuss and illustrate the "Arizona Seq." Consult U.S.G.S. Prof. Paper 233-D; and Johnson, Geol. Soc. Am. Bull. v. 81 p. 2077-2015.

W. Stokes
Feb 7, 1979

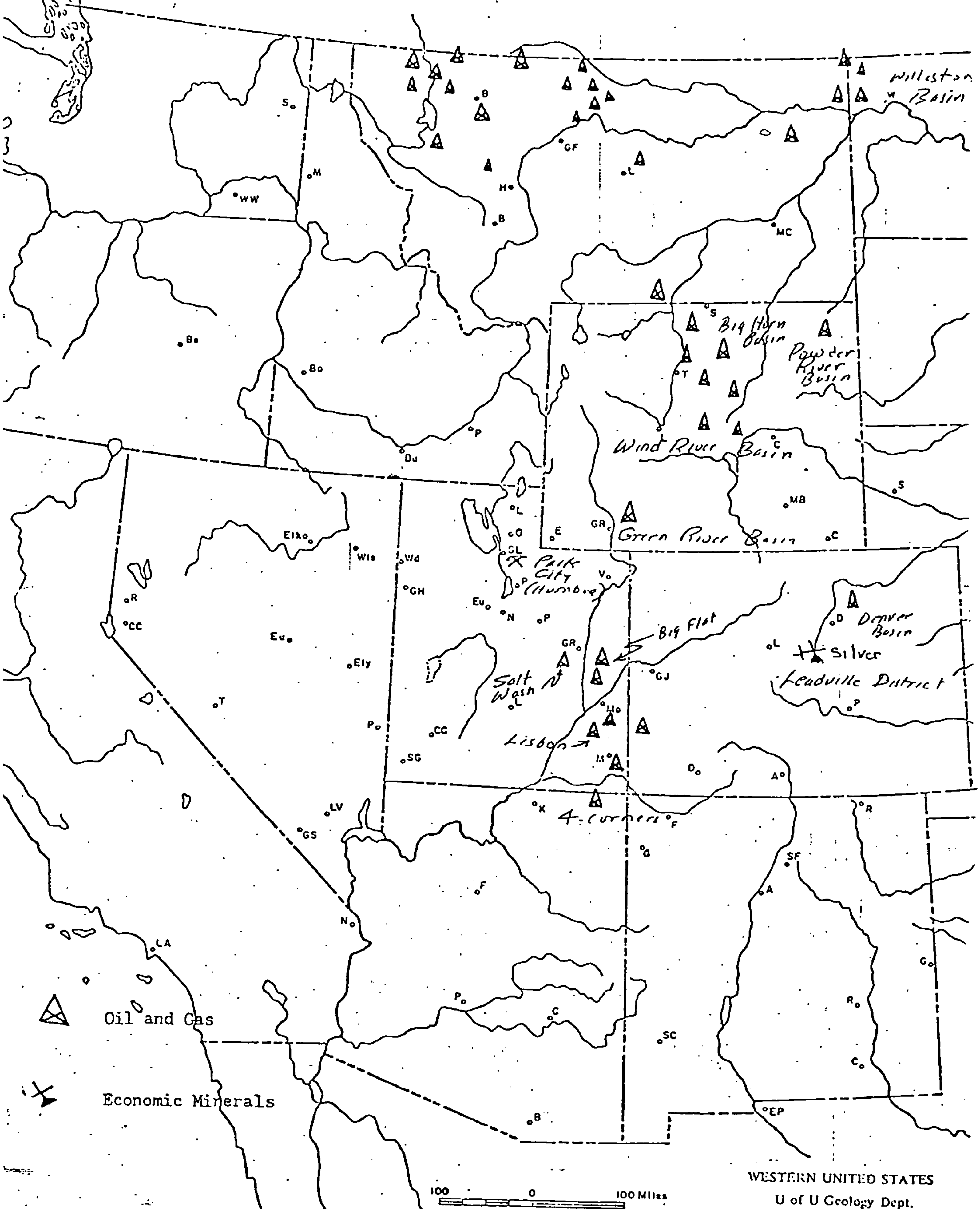
P.S. Please do not take these out of the library unless there are other copies remaining.

12/16/79

Stokes '79 Mississippian

7/16/79

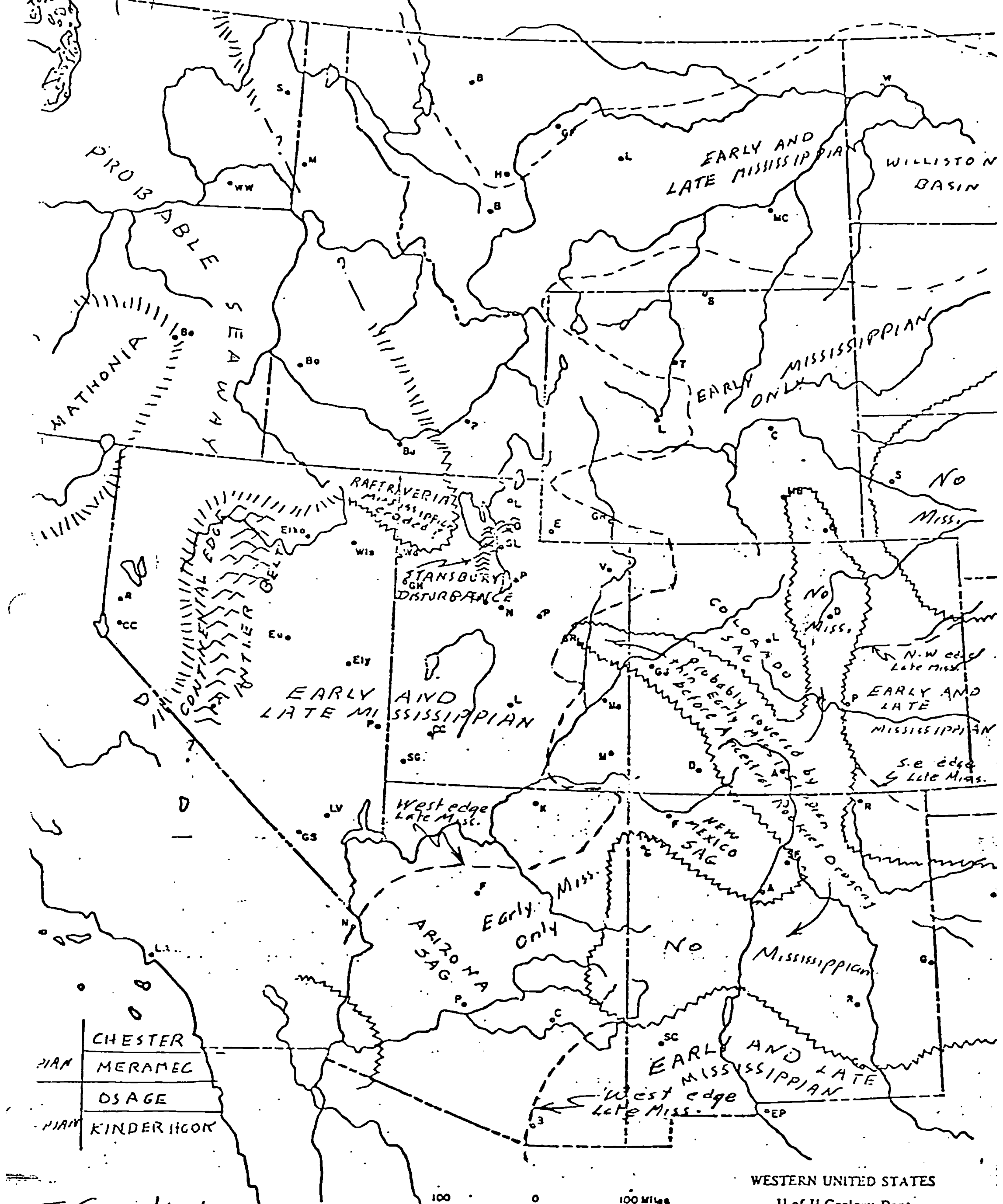
Mississippian System



Compiled from: Geologic Atlas of the Rocky Mountain Region U. S. A.
By Rocky Mountain Association of Geologists, 1972.

A 10/79

MISSISSIPPIAN PALEOGEOGRAPHY



WESTERN UNITED STATES
U of U Geology Dept.
W. L. STOKES
1971

1979

2/16/79

Generalizations

Paleozoic Stratigraphy of the Western Interior

MISSISSIPPIAN

I. Historical

A. The term Carboniferous, a natural derivation from the important coal resources associated with the correspong time period in Great Britain and Europe,was introduced by W. D. Conybeare and W. Phillips in 1822 in their study of British Geology. The term is still standard in Europe and many other countries even though a basic twofold subdivision is accepted by most. Belgium is a standard for European usage.

B. The economic importance of coal in the Appalachian Region led to early study of associated geology. The state geological survey of Pennsylvania was established in 1836 and in 1891 Henry Shaler used the term Pennsylvanian in preference to Upper Carboniferous. In the same report he introduced the term Mississippian for the limestone-rich beds below the Pennsylvanian. The USGS did not accept the Mississippian and Pennsylvanian until 1953. Many American geologists seem inclined to use European series and stage names for the Mississippian but some employ terms from Missouri and Illinois.

II. Tectonics and Paleogeography

A. Cordilleran Geosyncline, with Antler Orogenic Belt; Canadian Shield; and a very distinct Williston Basin dominate the tectonic framework.

B. Transcontinental Arch becomes much less positive and is broken by distinct sags in Colorado and Arizona and extensivley flooded by shallow seas. Correspondingly narrowing of "islands".

C. Pronounced downsinking of Williston Basin which has rather narrow but persistent connections westward across Montana.

D. Probable enlargement of Klamathonia as evidenced by Mississippian strata in central Oregon (Coffee Creek).

E. Orogenic activity in west-central Nevada (Antler Orogeny) creates additional land mass and highlands. Extent of orogeny northeasterly and southwesterly is debated.

F. Emergence of northwestern Utah into low land mass may record enlargement of Raftriveria. No Mississippian in Newfoundland Range.

G. Pronounced and extensive downsinking in the Cordilleran Geosyncline in Late Mississippian causes seas to retreat westward leaving extensive tracts of Early Mississippina exposed on the shelf areas. Downfold probably associated with Antler compression.

III. Lithology and Sedimentation

A. The Mississippian Period has an estimated duration of 25 million years. There are many extensive exposures in the Western Interior. Notable are Early Mississippian formations on the shelf (Madison, Leadville, Redwall). By contrast, and as a result of the Antler Orogeny, sediments of the western Great Basin are clastic (White Pine, Chainman). Subsurface Mississippian is extensive in the Williston Basin, chiefly carbonates.

B. This was a great lime-producing period - world wide manifestations. Estimated percentage of rock types: limestone 50%, fine-clastics 20%, quartzite and sandstone 15%, dolomite 10%, coarse clastics 5%.

C. A very cherty system. Many beds are over 50% chert.

D. Mid-Mississippian was a major turning point in sedimentation, again world wide. Before this was mainly carbonate deposition, afterward increasing clastic contributions building up toward the present. This resulted form the exposure of older quartz-rich terranes by the Hercynian-Appalachian disturbances. Removal of the early Paleozoic carbonate "armon plate" was under way.

E. Stratigraphy reflects mainly moderate to shallow water deposition in clear seas during Early Mississippian time and east of the Utah-Nevada line. West of this and in Confusion Basin, Utah, orogenic, shallow water to subaerial deposits record the eastward spread of detritus from the Antler Belt.

F. Evaporites characterize Williston Basin and are expectable from its interior position and climatic situation.

G. Various "islands" of Transcontinental Arch were supplying limited amounts of clastic components into surrounding seas. Wind action may be indicated.

H. The first notable expression of extensive, subaerial weathering of a carbonate terrane is found on the Early Mississippian over wide areas of the shelf (Molas, Belden).

IV. Biostratigraphy

A. Mississippian marine faunas are prolific and varied. Many groups were more than numerous, they were true rock-builders- crinoids, corals, foraminifera. Many excellent specimens are readily available. Less silicification than in previous periods. Estimated abundance of obvious animal groups in decreasing abundance: brachiopods, corals, bryozoans, echinoderms (encrinites), protozoans, (endothyrids) molluscs, conodonts. Zones have been proposed on the basis of several groups; most used are corals (especially colonial forms), brachiopods (especially spirifers and productids), cephalopods (certain restricted facies, goniatites especially), conodonts, bryozoans.

B. Probably because of the barrier of the Transcontinental Arch it is difficult to correlate western faunas with the mid-continent. Many good correlatiwns with Eurasia and almost zone for zone along strike into Canadian Rockies.

C. Strange lack of land plants and animals. Paleontologists should look more closely at the shallow water and continental beds in Nevada.

D. The Great Basin may be one of the world's best repositories for late Mississippian Marine faunas. Many opportunities here.

V. Boundaries

See Devonian summary for lower boundary.

The upper boundary is again a very difficult and problematical one. Considering those areas not covered by Late Mississippian rocks the boundary is not as difficult to locate although it does lie in a zone of weathering and disintegration in many places. In the Great Basin, especially Oquirrh and Confusion Basins there was continuous sedimentation and the datum lies in a number of formations: basal Oquirrh Manning Canyon, Diamond Peak, Chainmon. Faunal evidence, probably from brachiopods, will determine the boundary ultimately. There seems to be no good tectonic or paleogeographic criteria.

VI. Economic

Chief petroleum objective in Williston Basin (Mission Canyon is big producer)

Oil in Big Flat field in Utah.
Limestone for flux, aggregate, and cement; Flux, Tooele Co., Utah

Uranium in Wyoming-associated with solution effects
Host rock in mining districts - Leadville, Colorado
Ground water in Black Hills - northern plains
Potential phosphate in Brazer Fm. Great Basin.

VII. Scenic

Timpanogos Cave and lower slopes of Mt. Timpanogos, Wasatch Range; Brush Creek Cave, near Vernal, Uinta Range; Chinese Wall, Logan Canyon and vicinity; Red Wall in Grand Canyon.

Bibliography

1956, "Correlation of the Mississippian Formations of North America", Bull. Geol. Soc. Amer., Vol. 59, pp. 91-146. J. Marvin Weller, Chairman.

W Lee Stokes, 1979

5/11/77

AGE	LITHOLOGIC TYPES	WHITE PINE DISTRICT Hague 1883	EUREKA DISTRICT Hague 1892	EUREKA DISTRICT Nolan et al. 1956	ELY DISTRICT Spencer 1917	PIOCHE DISTRICT Westgate & Knopf 1932	LANGENHEIM 1962	THIS PAPER Sadlick
MISSISSIPPIAN	Fine brown clastics	Diamond Peak Quartzite	Diamond Peak Quartzite	Diamond Peak Quartzite	Diamond Peak Quartzite not recognized	Scotty Wash Qtzt.	Diamond Peak or Scotty Wash formations	Chainman Formation: Jensen Mem.
	Yellowish sandstone and conglomerate							Willow Gap Ls. Mem.; Donner Mem.
	Dark ripple laminated sandstone	White Pine Shale	White Pine Shale	Chainman Shale		Peers Spring Formation	White Pine Group: Hamilton Canyon Formation	Hamilton Canyon Mem.
	Dark shale				Chainman Shale		Chainman Shale	Camp Canyon Mem.
	Siltstone						Peers Spg.	Needle Mem.
	Siliceous limestone			Joana Limestone	Joana Limestone	Bristol Pass Limestone	Joana Limestone	Joana Limestone
DEVONIAN	Calcareous shale			Pilot Shale	Pilot Shale	Not recognized	Pilot Shale	Pilot Shale
	Devonian limestone	Nevada Limestone	Nevada Limestone	Devils Gate Formation	Nevada Limestone	West Range Limestone	Devils Gate Limestone	Devils Gate Limestone

Joana [illegible]

FIGURE 2. HISTORY OF NOMENCLATURE OF MISSISSIPPIAN STRATA OF EAST-CENTRAL NEVADA. NO CORRELATIONS ARE IMPLIED.

7/16/79

THE ANTLER OROGENY

Historical

Ferguson and Cathcart (Jour. Wash. Acad. Sci. V. 42, No. 3, 1924) first reported Permian rocks overlapping folded and eroded strata of Ordovician age in the Toquima and Toyabe ranges and Candelana Hills. From this Nolan postulated that a north-northeasterly trending geanticline had formed in central Nevada during the late Paleozoic time (AJS N. Ser., Vol. 16, No. 92, 1928). Eardley named this the Manhattan geantacline from the Manhattan mining district, about 40 miles north of Tonopah. (Jour. Geol. vol. 55, p. 309, f 1947). As evidence from many ranges, particularly in northcentral Nevada steadily accumulated, the term geantacline was found unsatisfactory and R. J. Roberts suggested replacing it with the term "Antler orogenic belt" (USGS open-file report, p. 95). From this came the term Antler Orogeny and the gradual acceptance of this event as a major disturbance equal in importance to the Taconic or Acadian of eastern geology.

Date of Orogeny

From original evidence the structural effects could be dated no closer than late Paleozoic - post Ordovician to pre-Permian. Detailed work in critical areas showed that rocks as young as Late Devonian are disturbed and that strata as old as Middle Pennsylvanian are generally not affected. "The orogeny probably comprised several distinct pulses, taking place at different times throughout the region." (Roberts, Hotz, Gilluly and Ferguson, AAPG, vol. 42, p. 2813 f., 1958).

Area of Structural Disturbance

Geologists have drawn the Antler Orogenic Belt in somewhat different ways. The maximum effects seem to lie in west central Nevada with extensions north and south. In Nevada the axis lies a few miles west of Manhattan, passes north northwesterly 30 - 40 miles west of Eureka, and then between Carlin and Winnemucca not far from Battle Mountain where Antler Peak, the type locality, is located. The axis then swings easterly curving north of Elko to pass through or near Mountain City. From here it is "anyones guess" as to where the Antler Belt actually lies. It may be buried under the Snake River Plain or may actually be in northwestern Utah. According to Churkin the Antler Orogenic Belt is well displayed north of the plain in central Idaho south of the batholith (AAPG. Vol. 46, No. 5, p. 569f). The sedimentary facies and geologic structures that are in alighnment from Nevada into central Idaho are so similar that a continuation of the Antler Belt seems to be highly probable.

The southern continuation of the Antler Belt into California is much less certain. Some investigators on the basis of coarse clastics in the Mississippian of the Inyo Mountains have postulated a continuation of the Antler orogenic effects into that region. In a recent paper Stevens and Ridley report that these deposits are probably caused by slumping on the steep west-facing continental slope and could not be derived from uplifted land to the west. They conclude that there is no evidence that the Antler Orogenic Belt ever extended into eastern California (Stevens, Calvin H. and Ridley, Albert P., Middle Paleozoic Off-shelf deposits in Southeastern California: Evidence for Proximity of the Antler Orogenic Belt, GSA Bull. vol. 85, #1, Jan. 1974, p. 27-32.).

Special Terminology

Sturctural and stratigraphic complications in and adjacent to the Antler Orogenic Belt made it necessary for workers in the region to use specialized terminology that may not be familiar elsewhere. This was described by Silberling and Roberts as follows: "Roberts and others used the term "assemblage" to designate major groupings of Paleozoic rocks which are representative of a particular sedimentary and tectonic environment in northern Nevada. The term assemblage is used similarly for rocks along the eastern and southern margins of the one considered here (Northwestern Nevada).

"A different kind of subdivision, however, is required in northwestern Nevada for the upper Paleozoic and lower Mesozoic rocks. The subdivisions adopted are lithologically and geographically discrete units of major rank termed "sequences" that are set apart from underlying or overlying sequences by unconformities. The sequences differ from assemblages in being discrete, vertically delimited rock units, some of which, though lithologically distinct, were deposited under much the same environmental conditions." (GSA Special Paper 72, p. 6).

Chief Tectonic Effects

The Antler Orogeny was basically a compressional event with thrust faulting as the chief effect. Many thrust faults and attendant thrust sheets have been mapped. Chief of these is the Roberts Mountain Thrust from Roberts Mountain, Eureka County, Nevada. The thrust sheets travelled eastward and there was much imbrication depending on local conditions. From the cross-sections that have been drawn one must conclude that the thrust planes were at shallow depths and the plates thin. There is little evidence of movement along deep levels of Precambrian rocks as was the case in the later more easterly disturbances.

Igneous Effects

The Late Devonian to Middle Pennsylvanian was a time of little volcanic action in the western United States. Igneous rocks in the Devonian and Mississippian of the Klamath Mountains have no proven genetic relationship to the Antler Orogeny. An important fact is that no granitic masses older than Permian have so-far been located in the western United States.

Rocks Disturbed

The section uplifted and eventually displaced by thrusting are of early Paleozoic age - specifically Cambrian, Ordovician, Silurian and Devonian. The Valmy Formation of Ordovician age apparently had the greatest volume of any formation effected and is the one most frequently encountered in the thrust sheets.

With few expceptions the disturbed rocks are of the eugeosynclinal and transitional facies. Roberts et al. gives the following estimate of rock types in the eugeosyncline: shale, 20 - 40%, sandstone, graywacke, and quartzite 10 - 30%, chert with shale partings as much as 30%, and volcanic deriviatives from a few to 30 percent. These are the materials making up the uplifts and thrust sheets and eventually dispersed by stream action to distant sites of deposition. The attached chert by Kay indicates the names of formations transported in the thrust silices.

Sedimentary and Stratigraphic Effects

The Antler Orogenic Belt was certainly topographically high enough to be called mountainous but was much inferior to the Sierras or Rockies of the present time. Debris spread out from the central zone and from the adjacent thrust sheets furnishes excellent clues to the type of material made available to erosion and also to the slope and configuration of the adjacent surfaces both exposed and submarine.

Chert fragments of the Vinini Formation are very distinctive and unusually durable. Black color is dominant but shades of red, green, black brown, white, yellow and pink are found. The formation tends to break into grit or granule-size fragments, pieces 1/8 to 1/2 inch are extremely abundant in most Vinini debris and there is little of larger or smaller size. Such chert pebble or chert granule conglomerates are common in the Diamond Peak Formation, the Chainman Formation and the Tonka Formation.

The best treatment of the sedimentological and stratigraphic effects of the Antler Orogeny beyond the actually disturbed areas is that of Sadlick (unpublished Ph.D. dissertation, U. of U.). He has named the Donner Member of the Chainman Formation from exposures in the Silver Island Mountains and included in it a variety of coarse and fine clastics most characteristic of which is conglomerate made up of chert pebbles one to three inches in diameter. The chert is confidentally assigned a source in the Ordovician Valmy Formation. The depositional edge of the Donner Member as defined by Sadlick is about 70 miles east of the orogenic belt. He visualizes it as probably accumulated "chiefly as a series of coalescing deltaic plains."

Possible Related Disturbances

What has just been described is the axis of maximum deformation. This is usually referred to and shown on maps as the Antler Orogenic Belt. Several other areas many miles from the central belt were distrubed at this same time. The Stansbury Uplift appeared in north central Utah in late Late Devonian time. It was a north-trending structure along which early Paleozoic miogeosynclinal rocks were exposed and deeply eroded. (Stokes and Arnold, 1958). An angular unconformity between the Chainman Formation and older rocks seen in the Silver Island Range near Wendover, Utah was dated by Sadlick and Schaeffer as middle Mississippian. (GSA Vol. 70, No. 12, p. 1786). They call this the Wendover Phase of the Antler Orogeny.

Some geologists (see Johnson GSA Bull, vol. 82 p. 3263) apply the term Antler Orogeny to all Late Devonian-Middle Mississippian disturbances throughout North America including the Franklynian Geosyncline in the Arctic.

The Antler Effect

J. G. Johnson in an article titled Timing and Coordination of Orogenic, Epeirogenic and Eustatic Effects, (GSA Bull. vol. 82, p. 3263) concludes that "orogeny, epeirogeny, and eustasy act in concert in response to the same driving mechanism, because the timing of orogeny in the geosynclines and the spread of marine waters to their maximum on the cratonic interior is found to coincide with remarkable accuracy during three of the four orogenies, and to be permissively in accord with this correspondence of timing in the fourth.* The name "Antler Effect" is suggested for this fundamental relation," p. 3263.

The Other Side of the Mountain

Before the Antler Orogeny the Cordilleran Geosyncline was unbroken by internal uplifts and marginal seas spread unimpeded from the open proto-Pacific across the interior lowlands to the center of the continent. In the middle Paleozoic the Antler Belt became a barrier and a source of sediment that for a time divided the marginal geosyncline into two parallel troughs each with its peculiar sedimentary contents. Eastward, shallow-water marine deposition continued throughout the Paleozoic and into the Triassic. Westward in northwestern Nevada a new basin, the Pumpernickle-Havallah basin was formed (see Bissell, 1962 section). In west-central Nevada, the Diablo sequence was deposited beginning in Permian time. Both of these sequences rest with profound angular unconformity on lower and middle Paleozoic rocks and overlap structures developed during the Antler Orogeny (Silbering and Roberts, 1962, p. 35).

South of a line between Reno and Eureka evidence for reconstructing the structure and stratigraphy of regions west of the Antler Belt is uncertain due to cover and igneous effects.

*He refers to the Taconic, Acadian-Antler, Sonoma-Appalachian, and Nevadan. See attached diagram.

Economic Geology

No mineral deposits seem to have been emplaced during or as a result of the Antler Orogeny. Mention must be made of the famous Carlin Nevada gold deposits that are mined for fine colloidal gold disseminated in leached carbonate strata at the Roberts Mountain Formation (Devonian) in a window of the Roberts Mountain thrust fault. Although the thrust fault and Antler Orogeny predaté and have no genetic relation to the ore there has been a great deal of prospecting of Nevada thrust sheets in hope of finding another Carlin.

Relation to Later Events

The Antler Orogenic Belt lies east of the Sonoma Orogenic Belt, created in Permian-Triassic time and west of unnamed Mesozoic disturbed areas that are chiefly of Jurassic-Cretaceous age. Effects of the Antler event are difficult to untangle from both of these because (a) the Sonoma Orogeny brought rocks from an unknown distance to the west and in the process deformed the sediments shed from the west flank of the Antler Belt, and (b) the mid-Mesozoic thrusts in many instances carried with them large segments of the Antler slices well beyond their previous positions. To complicate matters much of the Mesozoic movement in northern Nevada was south and southeasterward. This may have brought material out of western Idaho or Oregon that had not been previously displaced or it may have simply given an extra eastward component to the already-existing Antler displacement.

John Riva concludes from his work in the HD Range, Nevada, almost the last exposure to be seen before the lava cover: "The regional significance of this [southward-southeastward] direction of movement athwart the prevalent eastward direction, remains to be evaluated and understood" (John Riva, Thrusted Paleozoic Rocks in the Northern and Central HD Range, northeastern Nevada, GSA Bull. vol. 81, p. 2716f, 1970). Kay and Crawford working in central Nevada conclude that "...prospects are that thrusts of great lateral displacement developed through a long span in the late Paleozoic and early Mesozoic Eras" (Marshall, Kay and John P. Crawford, Paleozoic Facies from the Miogeosynclinal to the Eugeosynclinal Belt in Thrust Slices" GSA Bull., vol. 75, p. 425f, 1964).

Mechanics of the Antler Orogeny

It is tempting to try to fit the Antler Orogeny into a frame work of plate tectonics and to see in the resulting mountains a result of plate collision and compression. This impulse is heightened when it is observed that the period of most intense action, from Late Devonian to Middle Pennsylvanian was a time of known compressive reactions along the eastern borders of North America. The closure of the proto-Atlantic and the juxtaposition of the opposing continents was certainly in the middle Paleozoic coincident with the Antler Orogeny. Johnson's theory (see Antler Effect) has a fundamental truth.

Any efforts to relate the Antler Orogeny to plate interactions encounters immediate difficulties because:

1. There is an almost total lack of igneous rocks of both intrusive and extrusive kinds such as accompany subduction in more recent collisions.

2. There is no sign of the ophiolitic suite of rocks that accompanies subduction or underthrusting in most more recent plate collisions and even Appalachian-Caledonian collision zones of eastern North America.

3. Nowhere have basement rocks or oceanic crustal rocks been found in the associated thrust slices.

4. If there were plate tectonic effects the shortening or compressive effects were expressed as obduction rather than subduction which is always a difficult process to bring about in straightforward collisions of plates.

5. Assuming as anyone must that there were eastward directed forces these cannot have been related to converging land masses or closure of a marginal sea because the belt to the west remained a basin of deposition which received normal sedimentary deposits for 4 successive periods (Penn., Perm., Triassic, Jurassic).

6. If there were no converging blocks or plates either marine or submarine to exert east-directed pressure on the eugeosynclinal deposits their movement must have been caused by gravity which would require uplift on the order of 10-30,000 feet. If there were such an offshore welt or ridge did it move? And what happened to it later? And why did none of the thrust slices cut into it so as to move and preserve deep-seated rock in the orogenic belt?

Faced with these difficulties the best alternative in my opinion is some sort of "accordian plate tectonics" a la Churkin, 1974. Here is a possible mechanism whereby pressure may be exerted by an offshore buttress which immediately drifts away or subsides to leave the continental margin orogenically deformed while at the same time creating another basin on the same site as was once occupied by the displaced sediments in which further sedimentation could take place.

W. L. Stokes, 1976

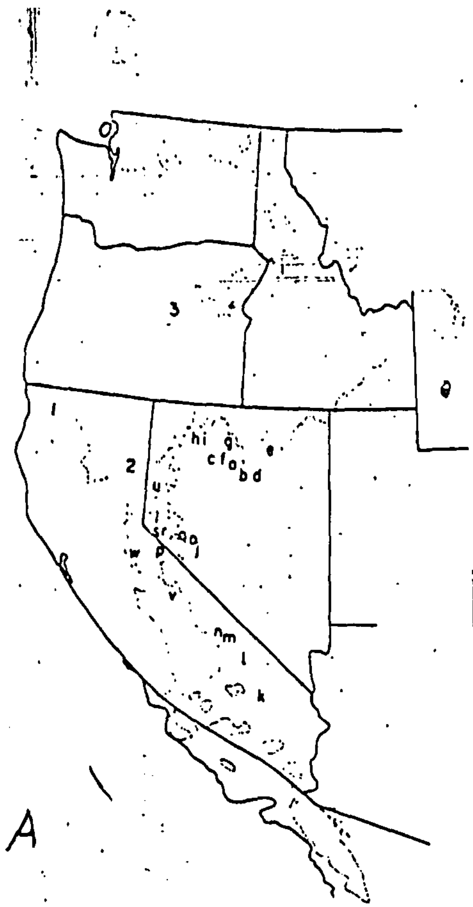


Figure 3. Locations of studied suites. Map constructed as described in caption for Figure 1. a. Scott Canyon Formation, Antler Peak quadrangle. b. Shwin Formation, Shosone Range. c. Valmy Formation, Water Canyon, Sonoma Range. d. Valmy Formation, Shoshone Range. e. Schoonover Formation, Jacks Creek, Independence Range. f. Pumpernickel Formation, Sonoma Range. g. Farrel Canyon Formation, Osgood Mountains. h., i. Happy Creek Formation and overlying Triassic(?) rocks, Jackson Mountains. j. Excelsior Formation, Hawthorne area. k. Soda Mountains Formation, Soda Mountains. l. Warm Springs Formation, Butte Valley. m. Unnamed unit, northern Inyo Mountains. n. Unnamed unit, Alabama Hills. o. Gillis Formation, Hawthorne area. p. Unnamed unit (Excelsior?), Bodie Hills. q. Wassuk Range, no samples. r. Unnamed unit southwest of Yerington. s. Triassic-Jurassic, Pine Nut Range. t. Unnamed units of Triassic-Jurassic(?) age near Genoa and Carson City. u. Peavine sequence, Reno. v. Volcanic rocks of Shadow Creek and Mammoth Crest, Devil's Postpile area. w. Unnamed Triassic-Jurassic(?) unit in San Andreas quadrangle. 1. Klamath Mountains. 2. Devonian island-arc assemblage of northern Sierra Nevada (from Hietanen, 1973). 3. Triassic keratophyres of central Oregon inlier. (from Dickinson and Vigrass, 1964). P. 1914

Figs A & B "Paleozoic and Lower Mesozoic Volcanism and Continental Growth in the Western United States" Rogers J.W., et. al. GSA Bull. vol. 85 # 12, 1974.

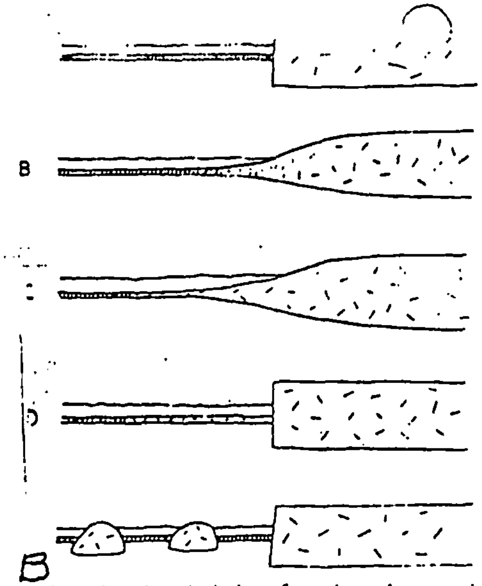


Figure 2. Hypothetical configurations along continental margins. Granite and basalt patterns represent continental and oceanic crust, respectively, in relationship to sea-level lines. Gradational transition between continental and oceanic material is shown in section B.

P. 1916

Figs C & D "Silurian (Llandovery) Downdropping of the western margin of North America" J.G. Johnson & E.C. Potter Geology, vol. 3, No. 6, 1975

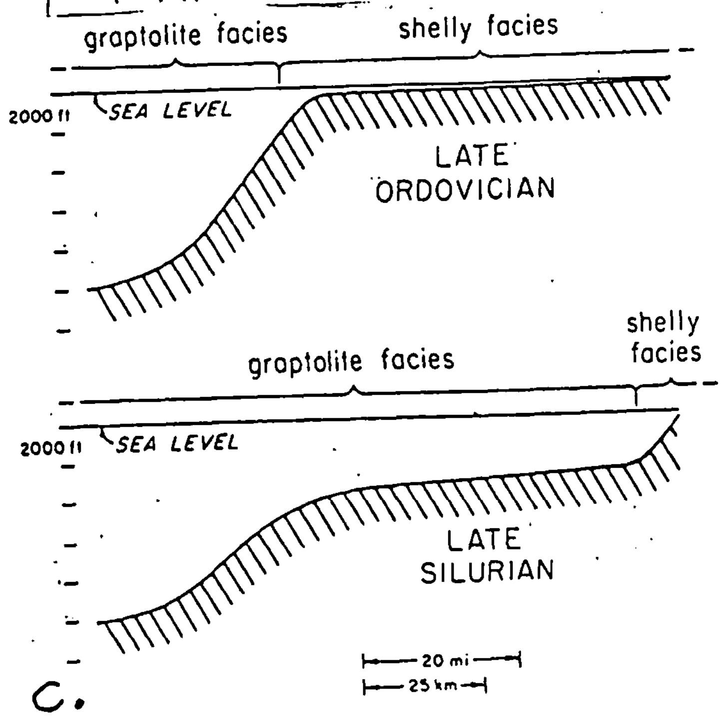
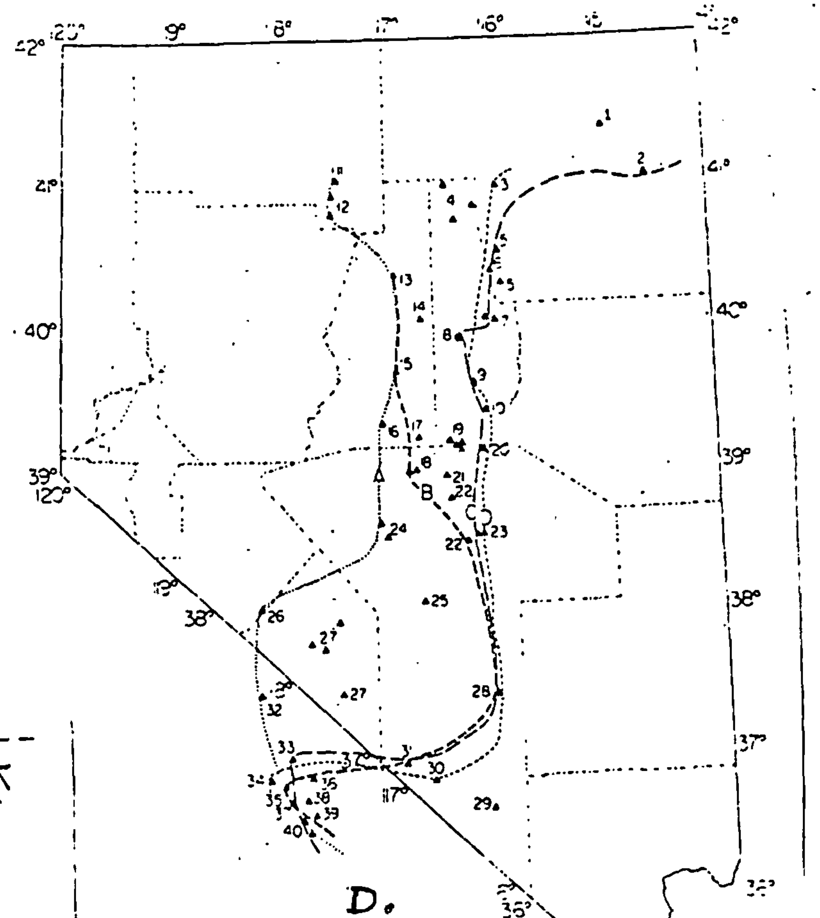


Figure 2. Two east-west profiles at 40th parallel, to show altered bathymetry due to downdropping of 64-km-wide strip of continental margin. B and C are plotted in Figure 1. We recognize that graptolite and shelly facies ordinarily overlap. Shelly facies is plotted here to include its supposed maximum extent and depth. P. 233



P. 232

Figure 1. Index map of Nevada, showing county boundaries. A = minimum westward extent of Lower Cambrian cross-bedded quartz sandstone, B = minimum westward extent of Upper Ordovician-Lower Silurian shelf carbonate, C = boundary between graptolite facies and shelf carbonate facies of Upper Silurian rocks, D = boundary between open-marine limestone facies and shelf dolomite facies of lower Middle Devonian rocks. References for numbered localities: 1–Johnson, 1970; 2–Thorman, 1970; 3–Evans and Ketner, 1971; 4–Roberts and others, 1967; 5, 6–Smith and Ketner, 1972; 7–Roberts and others, 1967; 8–Winterer and Murphy, 1960; 9–Merriam, 1940; 10–Merriam, 1963; 11–Ferguson and others, 1952; 12–Ferguson and others, 1951; 13–Gilluly and Gates, 1965; 14–Gilluly and Masursky, 1965; 15–Stewart and McKee, 1968; 16–Stewart and Palmer, 1967, and Washburn, 1970; 17–McKee and Ross, 1969; 18–Kay and Crawford, 1964; 19–Merriam, 1973; 20-23–Kleinhampl and Ziony, 1967; 24–Ferguson and Cathcart, 1954, and Kleinhampl and Ziony, 1967; 25–Kleinhampl and Ziony, 1967; 26–Ferguson and others, 1954; 27–Albers and Stewart, 1973; 28–Cornwall, 1972; 29–Burchfiel, 1964; 30–Cornwall, 1972; 31–Cornwall, 1972, and Cornwall and Kleinhampl, 1964; 32–Nelson, 1962; 33–Ross, 1967a; 34, 35–Stevens and Ridley, 1974; 36–Ross, 1967b; 37, 38–Ross, 1967a; 39–McAllister, 1956, and Ross, 1967a; 40–Ross, 1967a, and Stevens and Ridley, 1974.

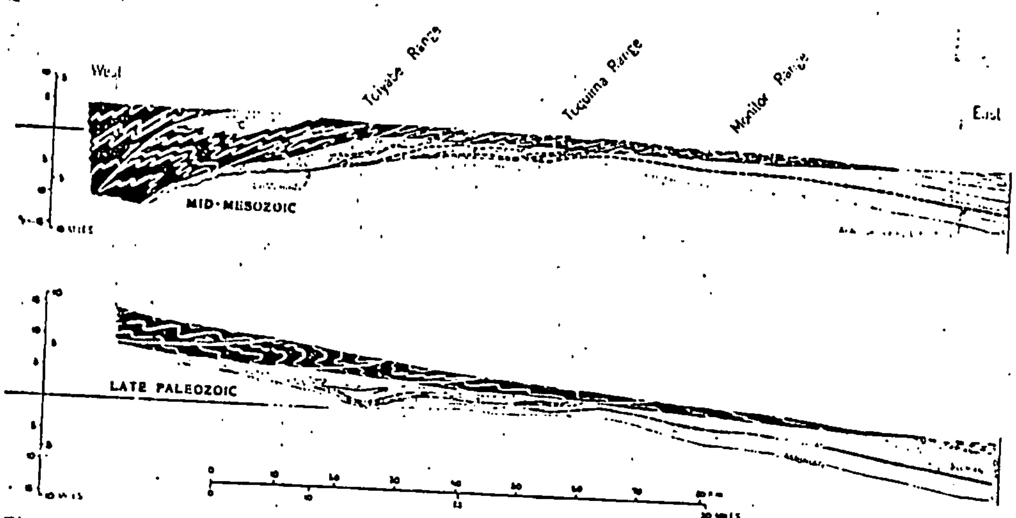


Figure 9. Restored structure sections across central Nevada prior to Basin-Range faulting. A. Restored structure section subsequent to deformation of Carboniferous sediments and early Mesozoic strata to the west, and prior to Nevadan intrusions. B. Restored section—early Carboniferous time on the hypothesis that the allochthon moved from a tectonic belt farther west in the Antlerian Orogeny, the thrust sheet gliding on an eastward-descending plane

From "Paleozoic Facies from the miogeosynclinal to eugeosynclinal facies in thrust slices, Central Nevada" Marshall Kay & J.P. Crawford, GSA Bull. Vol 75, p 425f, 1964

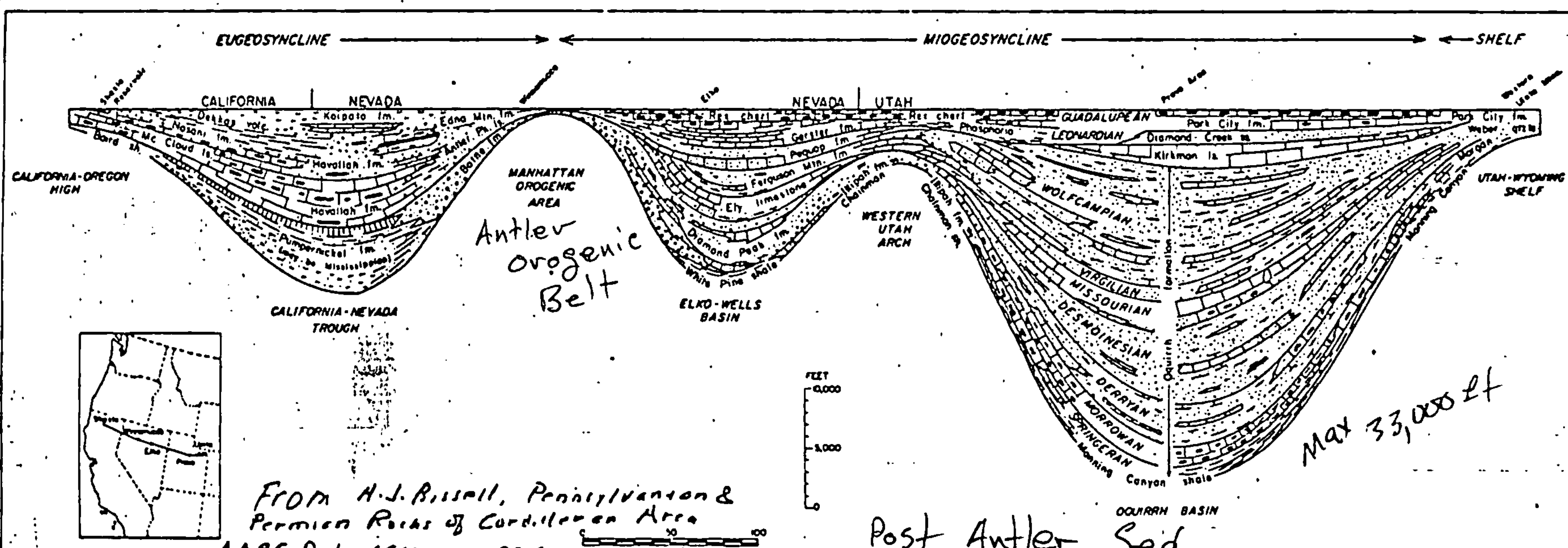
PALEOZOIC ROCKS ADRIFT IN SOUTH-CENTRAL IDAHO

Skipp, Betty, U.S. Geological Survey, Denver, Colorado 80225; Hall, Wayne E., U.S. Geological Survey, Menlo Park, California 94025

All Paleozoic rocks exposed north of the Snake River Plain between the Fish Creek Reservoir area and the Big Lost River are allochthonous, having moved east or northeast tens of kilometres on folded and faulted thrusts of probable Sevier age. Older thrusting of Antler age may have telescoped middle Paleozoic rocks exposed in the Fish Creek Reservoir area. The parautochthon throughout the area consists of folded and faulted Devonian miogeosynclinal shallow-water carbonates. At Fish Creek Reservoir, the parautochthon is overridden by five structural plates. The lowest two contain Devonian rocks of similar ages but of continental margin and deep-water oceanic facies; telescoping of these facies demonstrates tectonic transport of at least 48 km. The remaining three plates are Mississippian flysch of the Copper Basin Formation and two plates of Permian and Pennsylvanian interior basin deposits of the Wood River Formation, the highest structural element. The Copper Basin allochthon, some 19 km wide, consists of ...

conglomerates, sandstone, and impure limestone composing a Mississippian flysch sequence more than 7,000 m thick. This allochthon is bounded above and below by folded and faulted thrust faults with traces which trend N30°W. On the west, the Wood River Formation overrides the Copper Basin. On the east the Copper Basin is thrust over an intricately folded Mississippian through Upper Permian miogeosynclinal carbonate sequence which also has been thrust eastward several kilometres. Left-lateral tear faults trending N70°E offset the Copper Basin and upper Paleozoic carbonate allochthons.

Abstract. GSA Boise, Idaho Meeting May. 1975, P. 643

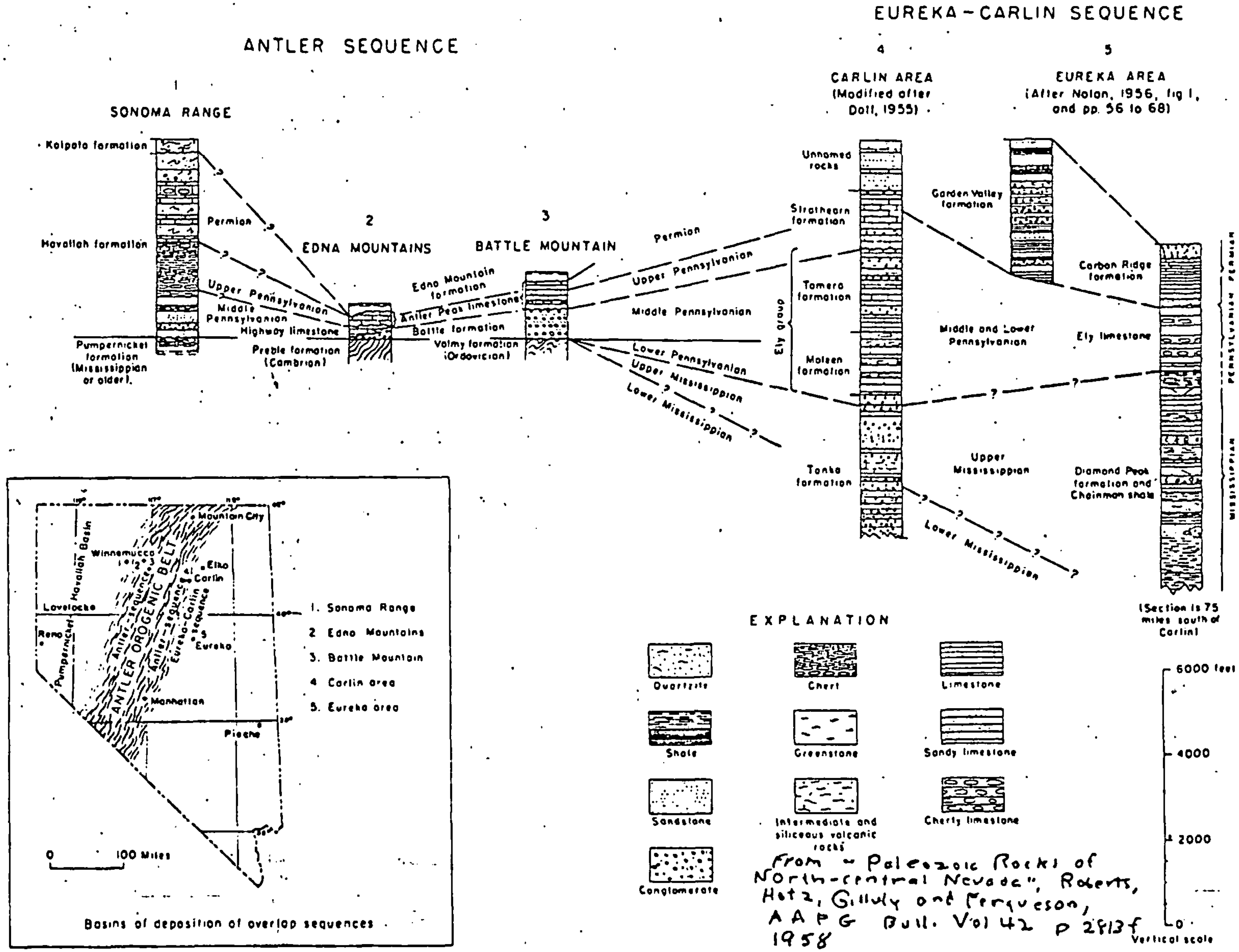


Antler Orogenic Belt

Max 33,000 ft

From H.J. Bissell, Pennsylvanian & Permian Rocks of Cordilleran Area AAPG Bull. 1962, p 1885.

Post Antler Sed.



From ~ "Paleozoic Rocks of North-central Nevada", Roberts, Hotz, Gilluly and Ferguson, AAPG Bull. Vol 42 p 2813f 1958

FIG. 6.—Stratigraphic sections of overlap assemblage in north-central Nevada.

2/16/79

Mississippian Assignment.

Do ONE

1. Write 1-2 pages (plus one simple illustration on the topic "Events and processes in the interval from Early Mississippian to Pennsylvanian sedimentation in Colorado Plateau-Wyoming Shelf area"

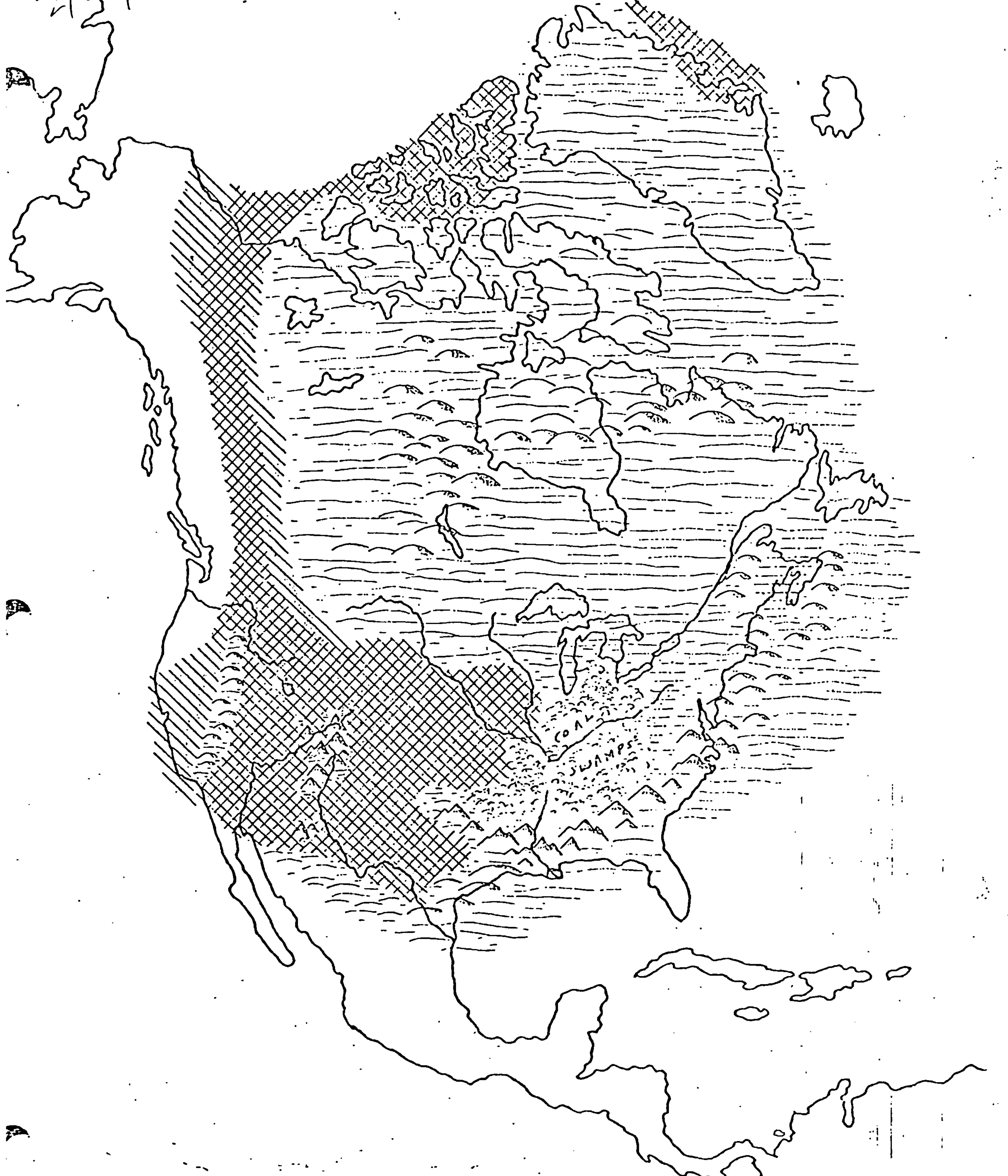
See Rocky Mountain Assoc. Geol. Guidebook on "Deep Drilling Frontiers of the Central Rocky Mountains" 1975. (Mississippian Drawer)

2. Comment on the presence of Mississippian on Precambrian with no intervening strata along the Uinta Range. See Williams, A.A.P.G. Bull v 37, p 2734-2742 and U.S.G.S. P. Paper 490., and Crittenden, "Miss. stratigraphy of the central Wasatch and western Uinta Mountains, Utah" — in Miss. drawer. Mayer Paper in Intermountain Geologist

Let ~~one~~ reference lead to others. The problem appears not to have been discussed specifically.

W. Stokes.
Feb 16 '79

2/23/79



Stoker, 1979

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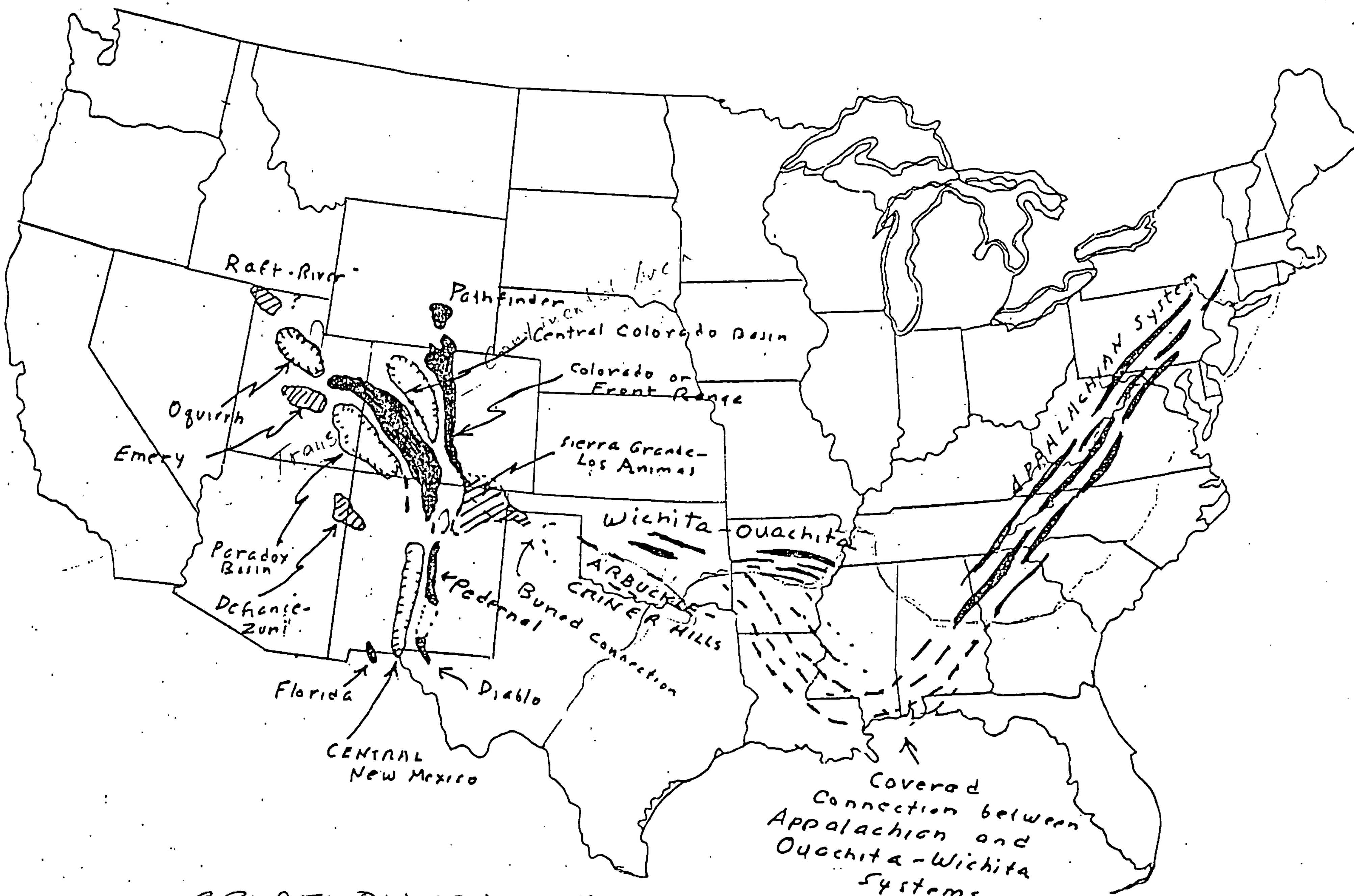
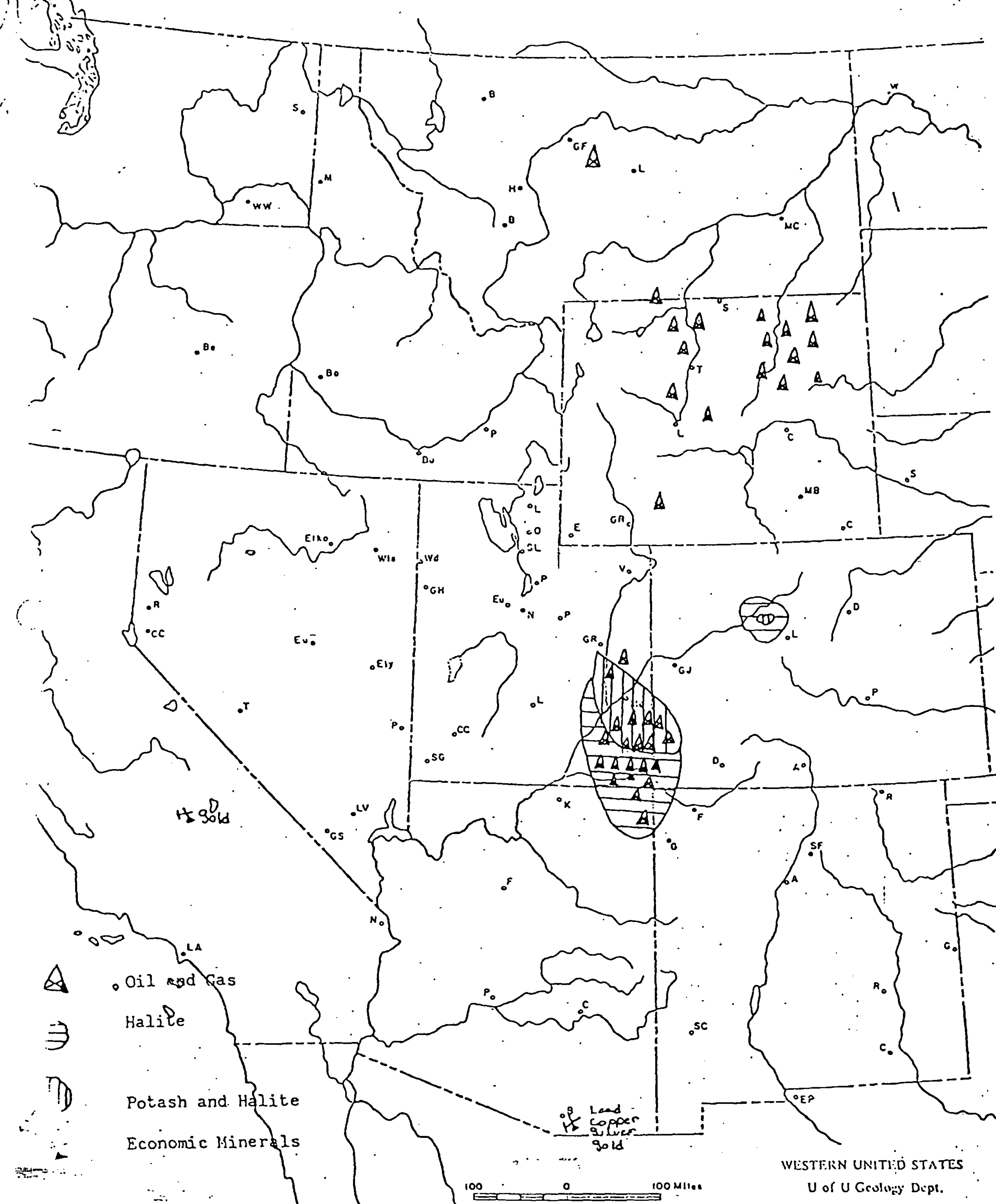


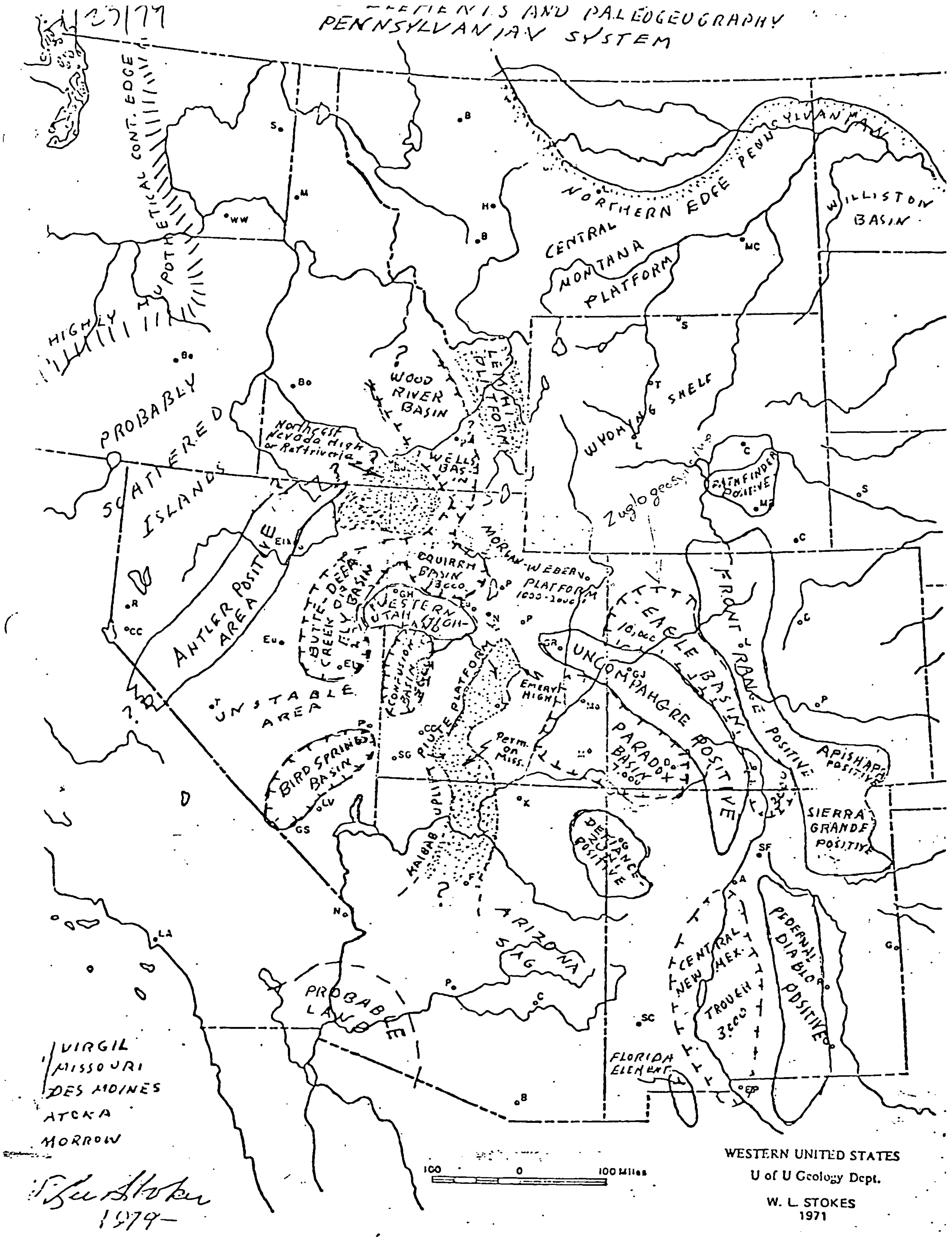
CRUDE DIAGRAM OF
APPALACHIAN-ANCESTRAL
ROCKIES TREND

Appalachian-Hercynian Trend

Pensylvanian System



Compiled from: Geologic Atlas of the Rocky Mountain Region U. S. A.
By ... Mountain Association of Geologists, 1972.



25/77
ELEMENTS AND PALEOGEOGRAPHY
PENNSYLVANIAN SYSTEM
HYPOTHETICAL CONT. EDGE
HIGHLY
PROBABLY
SCATTERED ISLANDS
NORTHERN EDGE PENNSYLVANIAN
CENTRAL MONTANA PLATFORM
WILLISTON BASIN
WOOD RIVER BASIN
LEHI PLATFORM
Northeast Nevada High or Rettrivaria
WELLS BASIN
WYOMING SHELF
PATHFINDER POSITIVE
Zuglogeosyncline
MORGAN-WEBER PLATFORM
1000-2000
OQUIRRH BASIN
3000
WESTERN UTAH HIGHLAND
BUTTE-DEEP CREEK
ELY BASIN
ANTLER POSITIVE AREA
UNSTABLE AREA
CONFUSION BASIN
PIUTE PLATFORM
EMERY HIGH
EAGLE BASIN
10,000
UNCOMPAHGRE POSITIVE
FRONT RANGE POSITIVE
APISHAPA POSITIVE
Perm. on Miss.
PARADOX BASIN
7,000
BIRD SPRINGS BASIN
KAIBAB UPLIFT
DEFIANCE POSITIVE
SIERRA GRANDE POSITIVE
ARIZONA SAG
PROBABLE LAND
CENTRAL NEW MEX. TROUGH
3000
PEDERNAL DIABLO POSITIVE
FLORIDA ELEMENT
VIRGIL
MISSOURI
DES MOINES
ATOKA
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U of U Geology Dept.
W. L. STOKES
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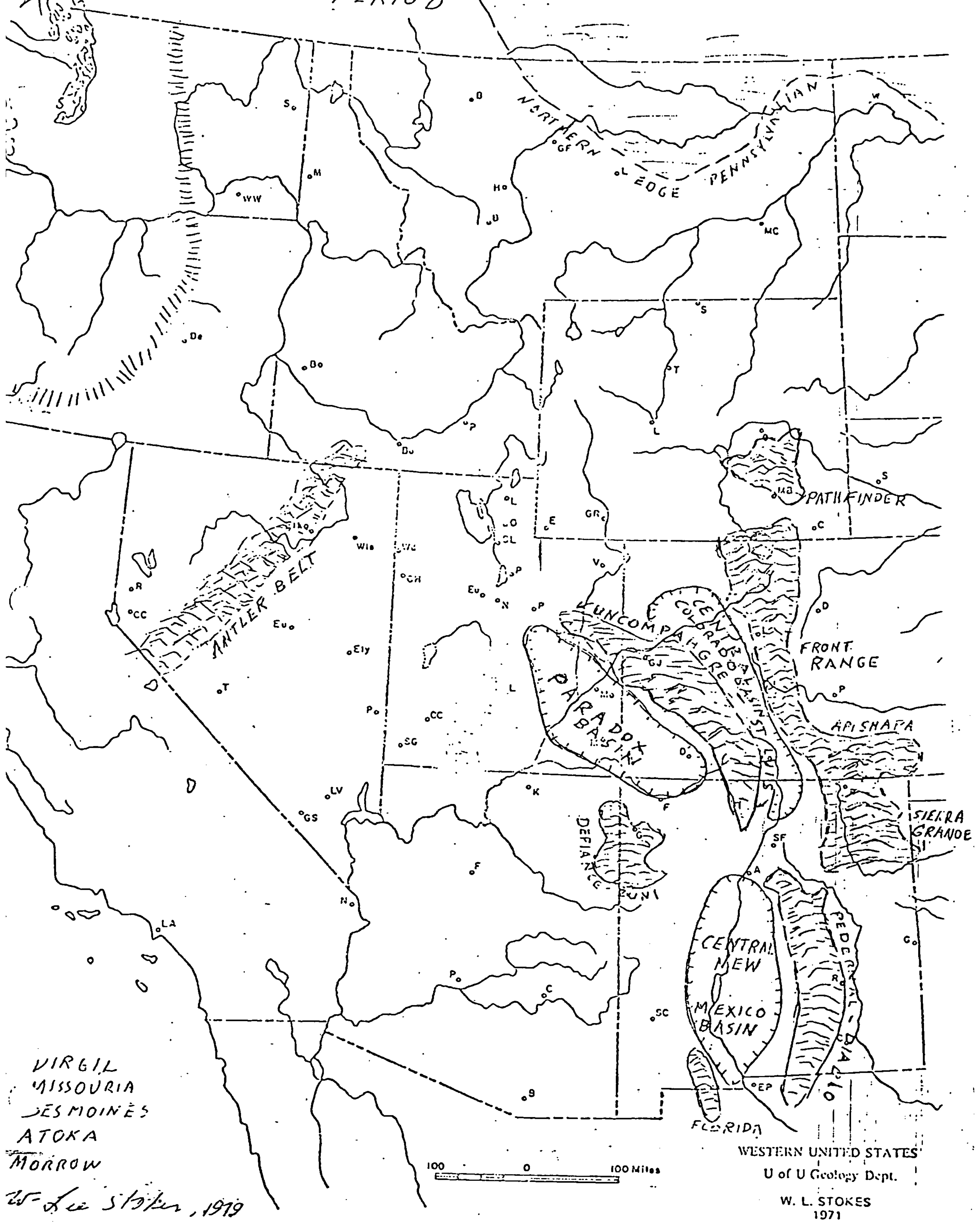
1979

... GEOGRAPHY — PENNSYLVANIAN PERIOD



W. Lee Stokes, 1979

2/23/79

Generalizations

Paleozoic Stratigraphy of the Western Interior

PENNSYLVANIAN

I. Historical

The Pennsylvanian Period or System of American usage is the approximate equivalent of the Upper Carboniferous of European geologists. The term Pennsylvanian was first used by Henry S. Williams in 1891. The name is taken from Pennsylvania where the system is prominent and economically important for the great coal deposits. However, eastern exposures are not typical and because they yield few fossils other than plants they have not proven satisfactory as standards of reference. Mid-continent Pennsylvanian exposures (Kansas, Missouri, Iowa, Oklahoma) are very fossiliferous and provide fairly good standard sections for North America.

II. Tectonics and Paleogeography

A. A major orogenic event which produced the Ancestral Rockes in Colorado, Utah, Wyoming and New Mexico took place in the mid-Pennsylvanian. This event completely altered previous tectonic patterns and left a permanent impression on subsequent geologic events. A great number of impressive features originated with this orogenic event. (See tectonic map attached). Maximum deformation was achieved where the orogenic belt crosses the Transcontinental Arch.

1. Positive high-standing elements capable of furnishing sediment, influencing drainage and modifying climate include: a) Uncompahgre Element, b) Frontrange Element, c) Pathfinder Element, d) Defiance-Zuni Element, e) Pedernal Element.

2. Positive low-lying elements furnishing little or no sediment and acting as shoals or low islands include: a) Raftriver High, b) Lemni Platform c) Piute platform.

3. Negative elements sinking so as to receive large volumes of sediment include: a) Paradox Basin, b) Eagle Basin, c) Oquirrh Basin, d) Central New Mexico Basin, e) Wood River-Wells Basin, f) Butte-Deep Creek Basin, g) Bird Springs Basin.

4. Areas that received sediment but were not notably negative or basin-like include a) Wyoming Shelf, b) Arizona Sag, c) Morgan-Weber Platform, d) central Montana Platform.

5. Many straits, bars, sills and spillways between and among the positive and negative elements have been named but there is little uniformity of usage. Authors have put forth names and interpretations but only time will tell which of these will prove to be of practical use.

B. The Antler Orogenic Belt continued active and the resulting highlands supplied sediment eastward into central Nevada. The northern and southern continuations of the Antler Belt are not clear.

C. The Williston Basin lost much of its former individuality and is scarcely identifiable in Pennsylvanian time.

D. The Wasatch Line is poorly marked and is not particularily discernable on isopach maps.

E. Evidence of increasing land masses in California, Oregon, and Washington, Klamathonia may have been welded to the continent during the Pennsylvanian.

III. Lithology and Sedimentation

A. The Pennsylvanian System contains the greatest variety and diversity of sedimentary rock types of any geologic period. The volume of sediment produced is probably greater than that of any other Paleozoic period. Exposures are widespread in the Great Basin and Rocky Mountains. Subsurface formations are extensive and varied.

B. Generalizations about rock types are difficult to make. Commonly, formations are mixtures that follow well-defined to vague cyclic patterns. Extimated percentages of rock types: sandstone and quartzite, more or less arkosic, 50%; limestone and dolomite, including mixtures, 20%; fine-clastics, including black shale, 10%; conglomerate, mostly arkose, 10%; salt, including sodium and potassium bearing types, 10%; coal, fractional percent.

C. There is surprisingly little volcanic material, either extrusive or intrusive considering that this was an orogenic period.

D. "Red beds" begin to appear (Morgan, Fountain, Madera). These are mostly traceable to the Ancestral Rockies.

E. The Paradox salt section is one of the most voluminous on earth. 29 cycles of deposition, each with potash salts, have been counted.

F. Cyclic repetition of lithologic types is common and must relate to the cyclothemic sedimentation of the mid-continent and elsewhere. Western "cycles" were largely submarine and hence lack coal.

G. The Oquirrh Formation (Penn-Perm) with a maximum thickness of 25 to 30 thousand feet is one of the thickest if not the thickest formation in the United States.

H. Reef-like accumulations, termed "reefoid" by Pennsylvanian stratigraphers are locally important.

IV. Biostratigraphy

A. Pennsylvanian faunas present a very orderly outgrowth from those of the Mississippian and are varied and prolific. The lithology seems especially favorable for preservation of specimens easily freed of matrix; silicification is locally important but not widespread. Estimated decreasing abundance of obvious animal groups: brachiopods, protozoans (fusulinids), molluscs, bryozoans, corals, echinoderms (crinoids), sponges and conodonts. Three groups have been particularily useful in defining workable zones: fusulinids, brachiopods, and conodonts.

B. Widespread flooding and break-up of the Transcontinental Arch permitted wide migration and easy interchange of marine faunas between the mid-continent and western areas. Same species in Kansas, Colorado, Utah, and Nevada. Canada has very little Pennsylvanian.

C. There are a few coal beds with plant remains, chiefly in New Mexico. Palynology could be helpful but is not much used in the Western interior as contrasted with the eastern coal basins. Salt contains well-preserved pollen.

D. With one exception there are few vertebrate fossils. The exception is the fauna from the Bear Gulch Member of the Tyler Formation, Early Pennsylvanian, Central Montana. A particularily productive site near Becket has yielded 27 genra of ray-finned fish and five sharks. Even more important are the first fairly good fossils of the conodont animal. (GSA Special Paper 141). It appears to be a vertebrate.

E. The most prolific writer and long-time student of Penn-Perm stratigraphy and fusulinid biostratigraphy of the western interior is Prof. H. J. Bissell, Brigham Young University. He has published many important papers, chiefly in the Bulletin of the A.A.P.G.

V. Economic:

The Pennsylvanian is probably of greater economic importance than any other geologic system in the Western Interior.

1. Many important fields in Montana (New Sumatra, Stensvad), Wyoming (Lander-Hudson, Winkleman), Utah (Aneth, Ashley Valley), Colorado (Rangely) and Arizona (Boundary Butte).

2. Great reserves of sodium and potassium-bearing minerals. One of the world's largest potash mines is operated by Texas-Gulf Sulfur at Cane Creek, Grand Co., Utah. Possibility of bromine from associated brines.

3. Host rock in several of the world's greatest metal mining districts: Bingham Utah; Ely, Nevada.

4. Building store locally. The great accumulations of gravel and sand along the shorelines of Lake Bonneville are chiefly derived from Oquirrh Formation; example, Point of the Mountain Spit.

VI. Scenic

A. The higher elevations of the southern Wasatch from American Fork Canyon to Mt. Nebo are carved from the Oquirrh Formation; also, most of the Oquirrh Range, particularily that visible from Salt Lake City.

B. Deeper parts of the Canyon of the San Juan River where it crosses the monument Upwarp, specifically, the Goosenecks section. Several small outcrops in the Cataract Canyon of the Colorado, stretches in the deeper canyons of the Green as it crosses the Uinta Range, Split Mountain and much of Yampa Canyon.

C. The Garden of the Gods, near Denver, is in upturned sediments of the Fountain Formation. Many other scenic spots along the Front Range are carved from this colorful formation.

Bibliography:

1944, "Correlation of the Pennsylvanian Formations of North America", R. C. Moore, et al., Geol. Soc. Amer. Bull., vol. 55, p. 657-706.

The Geologic Atlas of the Rocky Mountain Region (1972) contains the following papers on the Pennsylvanian System: "Regional Synthesis of the Pennsylvanian" compiled by William W. Mallory, p. 111-127; "Tyler Formation in the Subsurface, Central Montana", by F. S. Jensen and Kendall P. Carlson, p. 128-129; "Pennsylvanian Arkose and the Ancestral Rocky Mountains", by William W. Mallory, p. 131-132; "Pennsylvanian Rocks and Salt Anticlines, Paradox Basin, Utah and Colorado", Robert J. Hite and Fred W. Cater, p. 133-137; "Pennsylvanian Stratigraphy, Southern Sangre de Cristo Mountains, New Mexico", by Patrick K. Southerland, p. 139-143.

W. L. Stokes, 1979

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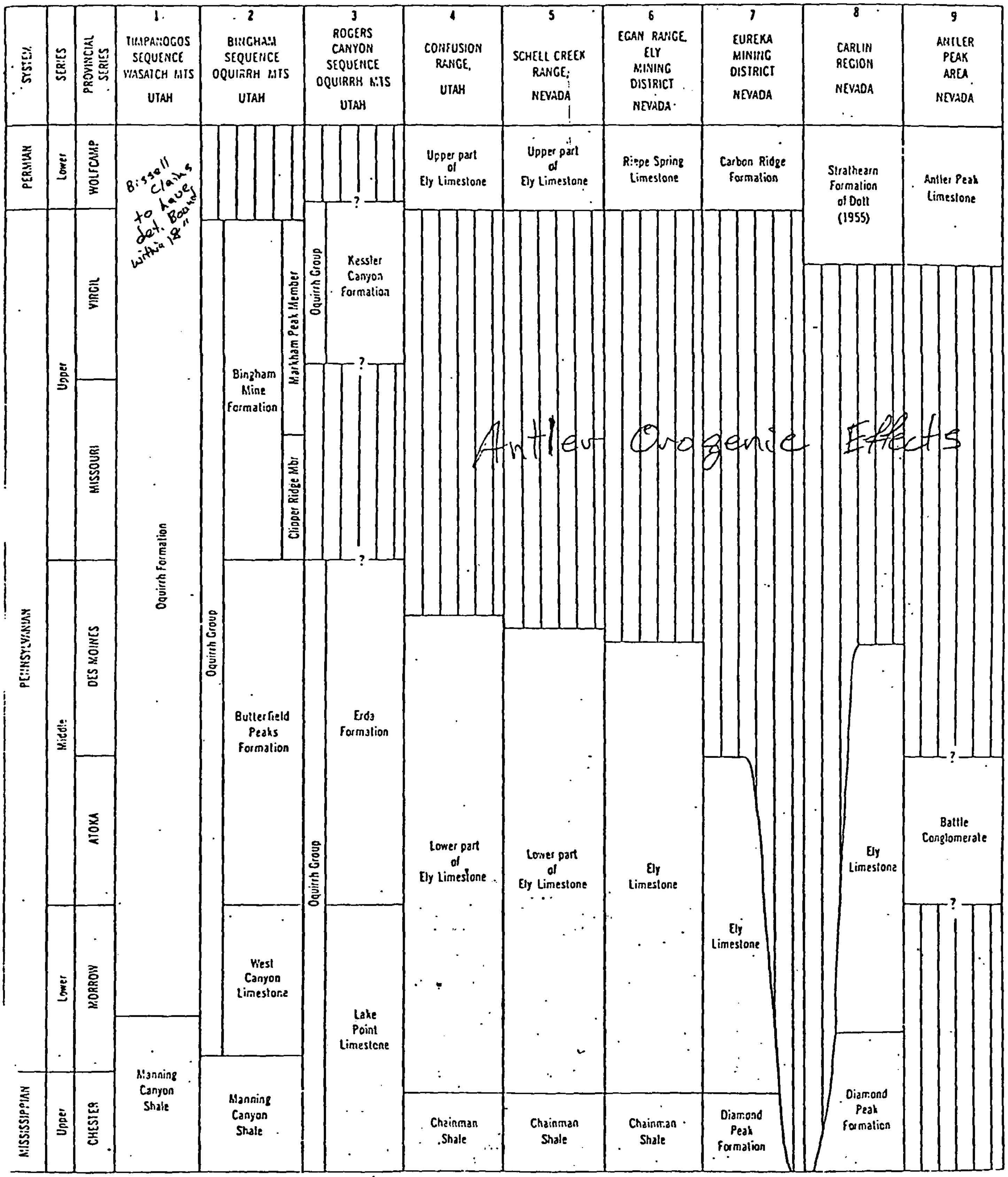

FIGURE 13.—Correlation chart of sections in the central Great Basin.

USGS P. Paper 450-E

Oquirrh of Bingham Can

3/24/74

ANCESTRAL ROCKIES

Historical

The name Ancestral Rockies was applied by Willis T. Lee an astute student of Rocky Mountain geology. He was thinking of the great major uplifts of Colorado and could not know the geographic and geologic extent of what he was naming.

Following Lee a number of smaller uplifts were located and named in New Mexico. Chief worker was C. B. Reed. His paper "Distribution of Pennsylvanian Rocks in Late Paleozoic Sedimentary Basins of Northern New Mexico" (Jour. of Geology, Vol. 55, 1947) revealed the existence of the Pedernal and Sierra Grande uplifts.

Late additions to the family of Ancestral Rockies were made by Mallory (USGU Prof. Paper 450-E, 1950) who named the Pathfinder Uplift of southern Wyoming and by Heilmun who attempted to annex the Raft River Uplift in northwestern Utah (Intermountain Assoc. of Petrolleum Geologists, Guidebook to 10th Annual Field Conference 1959).

It seems strange that the latest general review of tectonics, "The Tectonics of North America - A Discussion to Accompany the Tectonic Map of North America..." U.S.G.S. Prof. Paper 628 by P. B. King does not mention the Ancestral Rockies by name and gives only one sentence to them by any name whatsoever (p. 64).

Date of Orogeny

The elevation of the Ancestral Rockies appears to have been a geologically rapid event. Where evidence is best, as in central and western Colorado, it can be dated as Early and Middle Pennsylvanian with pulses of varying intensity shifting from place to place. Mallory (Geol. Atlas of the Rocky Mountains, 1972, p. 131) states: "In southeastern Wyoming, the oldest arkosic conglomerate in the Fountain Formation is Atoka... Near Castle Rock (S. of Denver), however, a calcareous zone near the base of the Fountain Formation contains a Morrow fauna..., and in wells near Pueblo, Colorado, arkosic sandstone projects as tongues into marine glauconitic rocks of Morrow age. The oldest arkosic conglomerate in the Minturn Formation.... is either of latest Atoka or early Des Moines age... The oldest arkosic conglomerate in the Paradox Basin is of Des Moines age..."

Area of Structural Disturbance

No one has defined the physical limits of the Ancestral Rockies. Greatest disturbances are in central Colorado but effects continue in both directions, into Utah and Wyoming on one hand and New Mexico, Oklahoma and Texas on the other. Considering the time of this disturbance it seems advisable to consider the Ancestral Rockies Orogeny a part of the Appalachian Orogeny and in even broader, almost global context as part of the Hercynian Orogeny of Middle Paleozoic time which affected Europe and possibly parts of Asia as well.

Considering major aspects of the Hercynian Orogeny and making appropriate comparisions with the preceeding Caledonian episodes of Ordovician, Silurian, and Devonian time the following characteristics emerge: 1) the disturbances were mainly east-west or latitudinal in effect, and 2) they seem to be due to compressive effects between the southern and northern continental masses, that is, by convergence between Laurasia and Gondwana. The Hercynian had the effect not only of compressing and elevating geosynclinal fillings where these existed but also of breaking across cratonic areas not previously covered by sediments.

Insofar as North America is concerned there is an almost unbroken line of major disturbances from New England on the east to Idaho on the west which includes as major elements the Appalachian System, the Ovachita-Wichita Systems and the Ancestral Rockies System. It seems significant that the major structural expression of the Appalachian System is folding, of the Ouachita it is overthrusting, and of the Wichita-Ancestral Rockies it is rifting or normal faulting.

Special Terminology

No special terms have been found necessary to describe the structure or stratigraphy of the Ancestral Rockies. But great diversity has come about in naming the major positive areas. Thus the Uncompahgre has been referred to as an Element, an Uplift, an Anticline, an Axis, an Anticlinorium, a Highland and as Uncomphagria. The word uplift seems to be most popular.

Less prominent positive areas have been called by other names; thus: Piute Platform, Emery High and Zuni Upland. Only time will tell which terms will survive.

Chief Tectonic Effects

The Ancestral Rockies Orogeny was basically a tensional event with normal faulting as the chief effect. The exact location of the faults bounding the various uplifts and basins can seldom be exactly located because of blanketing effect of debris shed across them from the uplifts. Internal faults are difficult to date and differentiate from dislocations of later, mostly Laramide age. Breaks in the basins or grabens are obscured by later deposition.

Rocks Disturbed

The section uplifted in most of the structural blocks of the Ancestral Rockies was basically a great and unknown thickness of Precambrian crystalline basement rocks and an overlying thin veneer of marine sedimentary formations of Pre-Pennsylvanian age. Beyond the area of most intensive deformation in Colorado, New Mexico, south central Wyoming, and eastern Utah the thickness of sedimentary rocks is greater but such areas are not included by all geologists in the Ancesteral Rockies.

The thickness of Precambrian crystalline rocks brought to the surface was probably at least 12,000 feet along the Uncompahgre Front (southwest flank of Uncompahgre Uplift). The overlying cover of Cambrian, Devonian and Mississippian in the same general area of greatest disturbance ranges from a feather edge to 2,500 feet or so. The sedimentary section is chiefly limestone, dolomite and shale.

Ancestor or Descendant?

Many geologists have objected to the term Ancestral Rockies because there seems to be no genetic relationship with the Laramide Rockies of later time. But the name sticks in the mind and grows in the literature and there is little hope of replacing it with anything else. After all it was and is a great feat to untangle the effects of two orogenies that cross each other as do the Ancestral and Laramide Rockies. Two giants of geology, W. T. Lee and T. S. Lovering disagreed over whether or not the Cretaceous seas ever crossed or covered the Rockies. This was because of evidence of both pre-Cretaceous and post-Cretaceous signs of orogeny and the Pennsylvanian uplifts were not understood at the time.

What we are coming to realize is that the Ancestral Rockies themselves may have formed along paths dictated by still older patterns. The heart of Colorado is a crossroads, not of two but of three orogenies. The ancient Transcontinental Arch, trending northeasterly across North America, is the result of an orogeny 1.5 - 1.8 billion years ago. (The famous Colorado Mineral Belt follows this trend.) The Ancestral Rockies broke across this trend about 300,000,000 years ago in the form of a conjugate fracture pattern. They traversed the resistant core of the Arch at right angles so as to expend the least energy and then resumed a more northwesterly course in Utah. Finally, in the Tertiary 60,000,000 years ago, a north-south system was superimposed as the Laramide Rockies rose across the previous trends.

The subject of rejuvenated tectonic features is a large one and the "crossroads" area of central Colorado is a prime area for its investigation.

Mother of Red Beds

From the viewpoint of stratigraphy and sedimentation the greatest effect ofthe Ancestral Rockies was to supply raw material for the famous red beds of the west. This role has been emphasized in an important paper by T. R. Walker (Formation of Red Beds in Ancient and Modern Deserts, G.S.A. Bull. 78, p. 353-368). This was followed by other papers the same worker. The source of the iron atmos, according to Walker are common iron-bearing minerals such as hornblende, chlorite, biotite, ilmenite and magnetite.

The iron-bearing minerals contained in the gneiss and schist of the Precambrian cores of the Ancestral Rockies were ideal sources of the hematitic pegment which eventually developed in many cubic miles of sediment derived from these uplifts. The effects of climate on the origin of red beds has been argued but no one has brought forth any serious objections to the time-honored theory that an arid environment is very favorable because the absence of near-surface water allows penetration of the oxygen that is essential to the process. Other evidences of aridity such as evaporites, wind-blown sands and scarcity of plants and animals during the reign of the Ancestral Rockies tell the same story.

Consider the spacial distribution of the red beds. Commencing at the north they are know to extend into northern Wyoming (Chugwater Group), the Black Hills (Spearfish), subsurface of Kansas and Nebraska (Dockum or Jelm Oklahoma, west Texas and eastern New Mexico (Dockum), northwestern New Mexico (Cutler Group, Chinle, San Rafael Group), northern Arizona (Cutler Group, Moenkopi, Chinle, San Rafael Group), eastern and south-eastern Utah (Cutler Group, Moenkopi, Chinle, Glen Canyon and San Rafael Groups), southern Nevada, southeastern Utah, and northwestern Arizona, (Hermit, Moenkopi, Chinle), northern Utah (Morgan, Moenkopi tongues, Chinle, Entrada), western Wyoming (redbed tongues of Triassic Thaynes and Permian Phosporia ormations).

This distribution centers in central Colorado and even if there were no other evidence, the case for deriviation from the Ancestral Rockies would be strong. The depositional edge can be drawn without much doubt on all sides except the southeast where the Dockum disappears under Gulf Coast sediments. The great lobe that trails across northeast New Mexico, western Oklahoma, and west Texas seems to be headed for a depositional basin in the ancestral Gulf of Mexico.

The distribution in time reflected in the stratographic section is also significant. The first appearance of red beds coincides with the introduction of arkose in the early and Middle Pennsylvanian. In Utah the first red beds (discounting those of the Precambrian and possibly Late Devonian) are in the Morgan Formation, seen extensively in the eastern Uinta Mountains. From this time red beds proliferate in the geologic sections of the Rocky Mountain states. The Four-Corners, shared by Utah, Colorado, New Mexico and Arizona is referred to as the Red-Rock Country even in the tourist brochures. The colorful Fountain Formation is an eye-catcher along the Funt Range and the "race track" topography of many of Wyoming's picture-book anticlines is due to the red beds. As a matter of fact the earlier geologists of the west used the term Red Beds as a useful synonym and map symbol for the undifferentiated continental Permo-Triassic.

After reaching a height of importance near the Permian-Traissic boundary red beds declined. There were fewer in the Jurassic than in the Triassic and fewer in the Middle Jurassic than in the Early Jurassic. By the close of the Jurassic there were none of significance. This distribution in time correlates exactly with the prominence of the Ancestral Rockies as topographic features and as source areas. It is surely no coincidence that the mountains were essentially submerged in their own debris during the Triassic and that encroachment of stream deposits from distant sources to the west was underway in Morrison (Late Jurassic) time. It may be argued that small remnants were still visible in the Cretaceous but this has not been proven.

Morrison is the end of the red beds

The history of Paleozoic-Mesozoic red beds in the western United States in essentially a history of the Ancestral Rockies.

Salt Basins Among the Uplifts

The Ancestral Rockies were discovered because of the prominence of the great uplifted segments. The true significance of the intervening basins came later and was not fully appreciated until much deep drilling had been done.

Many of the Pennsylvanian basins proved disappointing as producers of oil and gas but an accidental by-product of great economic importance was the discovery of salt in some of them. Drilling near Cisco, Utah encountered potash salt and led to the eventual understanding of one of the greatest reserves on earth. Salt and gypsum were also discovered in the Eagle Basin of central Colorado. Evaporatic deposits of less concentrated chemical types also took place elsewhere.

The special problems of dealing with salt-generated tectonics and stratigraphy will be covered in a later lecture.

Arkose

The arkosic sandstones derived from the Ancestral Rockies are among the most notable of the United States. Two other arkosic provinces are the Triassic of the Newark Series and the Eocene of California. Arkose is defined as sandstone with 25% or more feldspar in the nonclay detrital fraction.

Arkose when subject to reworking and disintegration gives many other rock types. Purer grades of sand represent one product. The stratigraphy of sandstones related to the Ancestral Rockies will be treated later.

A few geologists are thoroughly sick of Ancestral Rockies arkose. Several wells have gone through as much as 10,000 feet of it in test wells without reaching a hoped-for underlying marine section.

Climatic Clues

The effects of mountains as barriers to atmospheric circulation is a major topic in meteorology. Not much can safely be said about the climatic influences of ancient ranges because not only are the configurations and heights of the mountains largely hypothetical but also other climatological factors such as winds and water bodies are poorly understood.

It is, therefore, interesting to find a serious and quite believable attempt to relate the Ancestral Rockies to the paleobiology of Periman life in the Four Corners region (Vertebrates from the Cutler Group of Monument Valley and Vicinity, P. P. Vaughn, N. Mexico Geol. Soc. Guidebook, 1973, p. 99-f). Vaughn notes that both Permian plants and animals of the of the Four Corners - Grand Canyon region hint at arid, rigorous climates and are much less prolific than faunas and floras of the same age in Texas.

"Could these differences have been due to a rain shadow cast over parts of the Four Corners by the Ancestral Rocky Mountains? Opdyke (1961) has noted that the directions of slope of Permian aeolian deposits in the western United States point to winds from the north and northeast and has suggested that what is now the southwestern United States may have lain, during Early Permian time, in a belt of prevailing easterlies near the equator, in accord with paleomagnetic data. The crest of the Uncompahgre highland may have reached, in Wolframp time, a mile or greater above the surrounding country..." This would have been similar to the effects of the Sierra Nevada in Tertiary time which created a rain shadow and desert east of it.

Fits the arkose problem too.

The Ancestral Rockies and Global Tectonics

Any tectonic event as powerful and extensive as the Ancestral Rockies Orogeny must eventually be examined in the light of the new global tectonics. No one seems to have done this. What is offered here are a few problems that come to mind in thinking about the gross relationships of this mountain system. Bear in mind a similar summary offered in connection with the Antler Orogeny.

1. There are no evidences of either subduction or obduction in this system. On a smaller scale there are few evidences of thrust faulting or even gravity sliding. Unless there are unsuspected deep-seated underthrusts as in the Laramide Rockies this system shows no significant sign of compression and cannot be related to continental collision.

2. No igneous rocks are genetically associated with this system. The same may be said of mineral deposits and even of hydrothermal activity. Nothing seems to have been operating to yield either heat or deep seated gaseous, liquid,or soild products.

3. Tensional effects are evident from the great normal faults such as that on the Uncompahgre Front with an estimated throw of at least 20,000 feet. Such effect could scarcely be attained without relaxation of crustal forces or actual extension at a lower level. Only in connection with rift zones are such effects compatable with global tectonics.

4. The Ancestral Rockies are not parallel with or adjacent to the continental margin as it then existed or as it now exists. As a matter of fact they are oblique to the Gulf Coast and to the Pacific Coast.

5. There are no obvious relationships to former zones of sedimentation or downfolding. The trend of disturbances cuts through a region of minimal sedimentary section, in places perhaps with no sedimentary cover at all.

6. Other sections or systems of the great Hercynian-Appalachian disturbance can be related to reactions between the southern and northern continents. Thus, the European sections could reflect the pressing of Africa against Europe and the Appalachians the late Paleozoic collision of northwest Africa and eastern North America.

The only feasable way of affecting the southern and southwestern parts of North America by plate movements at the proper time would be to bring South America against this part of North America. An illustration of how this might be arranged is presented (attachment) but it is an awkward solution to say the least and to bring it about requires that Central America and most of Mexico be absent.

But assuming that the weight of South America could be brought to bear briefly against the southwest edge of the solid transcontinental arch it seems possible that a tension-creating situation conductive to rifting along the lines of the Ancestral Rockies could be created.

W. L. Stokes, 1976

3/2/54

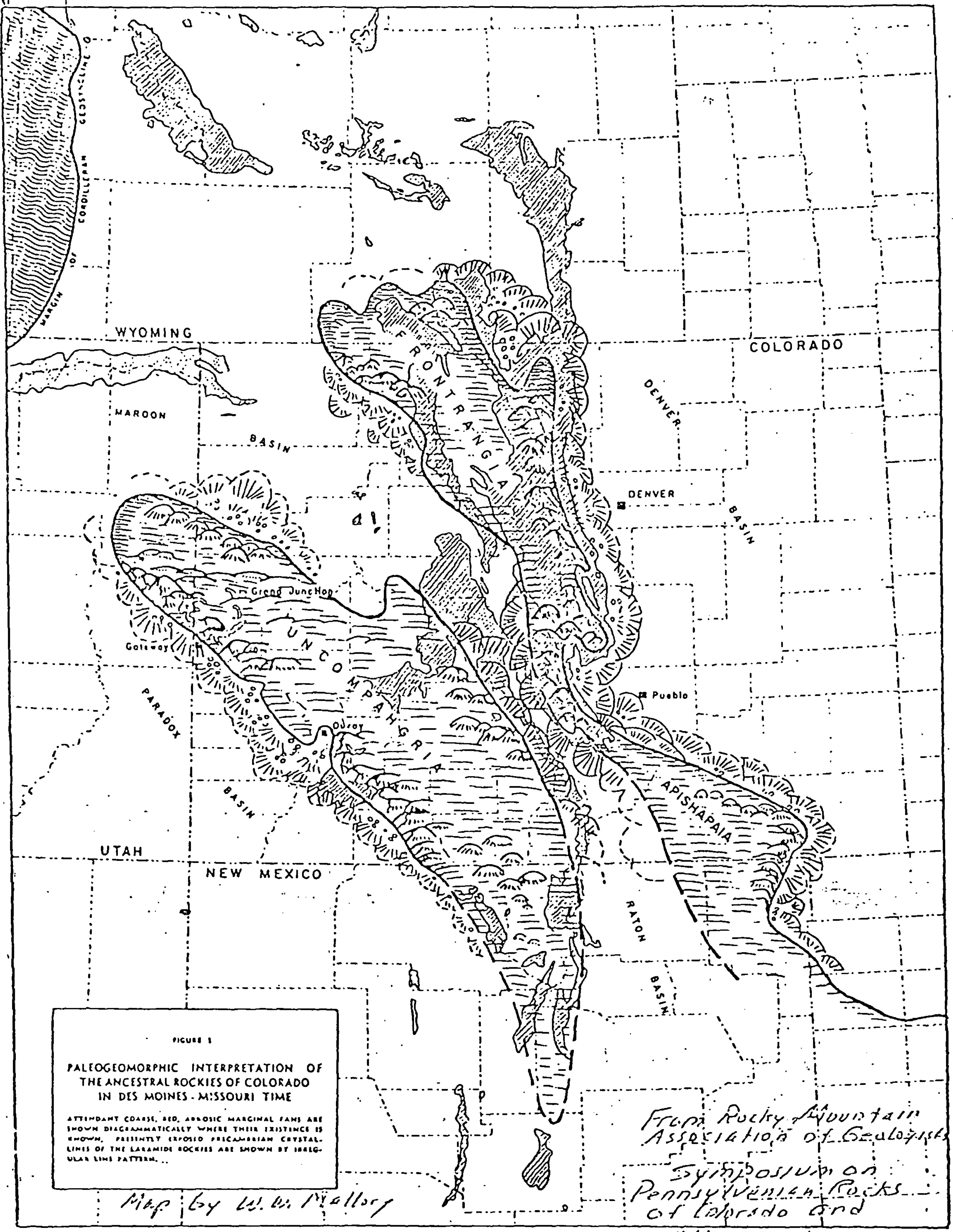

FIGURE 1

PALEOGEOMORPHIC INTERPRETATION OF THE ANCESTRAL ROCKIES OF COLORADO IN DES MOINES - MISSOURI TIME

ATTENDANT COARSE, RED, ARKOSIC MARGINAL FANS ARE SHOWN DIAGRAMMATICALLY WHERE THEIR EXISTENCE IS KNOWN. PRESENTLY EXPOSED PRECAMBRIAN CRYSTALLINES OF THE LARAMIDE ROCKIES ARE SHOWN BY IRREGULAR LINE PATTERN. ..

Map by W. W. Mallory

From Rocky Mountain Association of Geologists
Symposium on Pennsylvanian Rocks of Colorado and Adjacent Areas
1954

14 Four Corners Geol. Soc. Guidebook, 8th Field Con. Canyonlands 1975

PENNSYLVANIAN CORRELATION DIAGRAM

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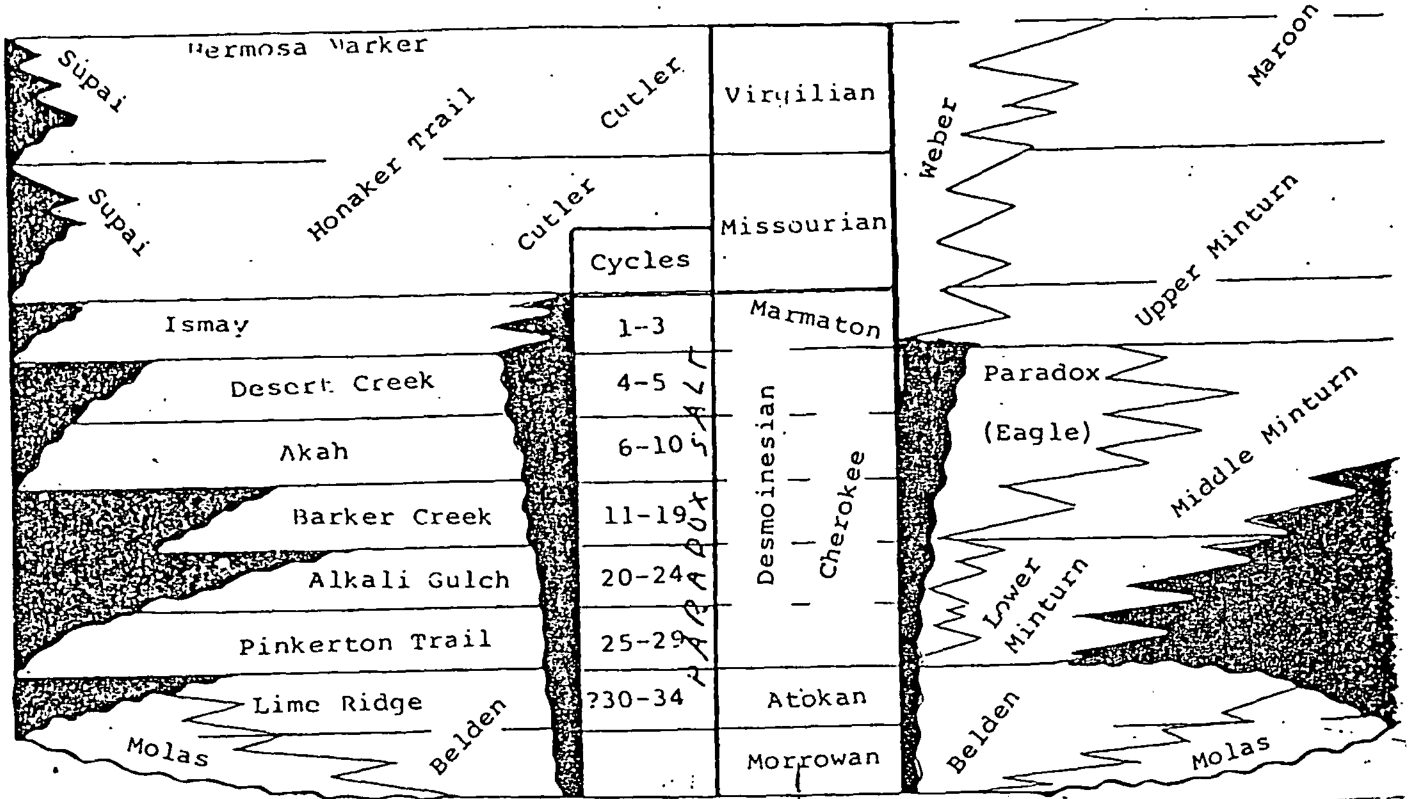

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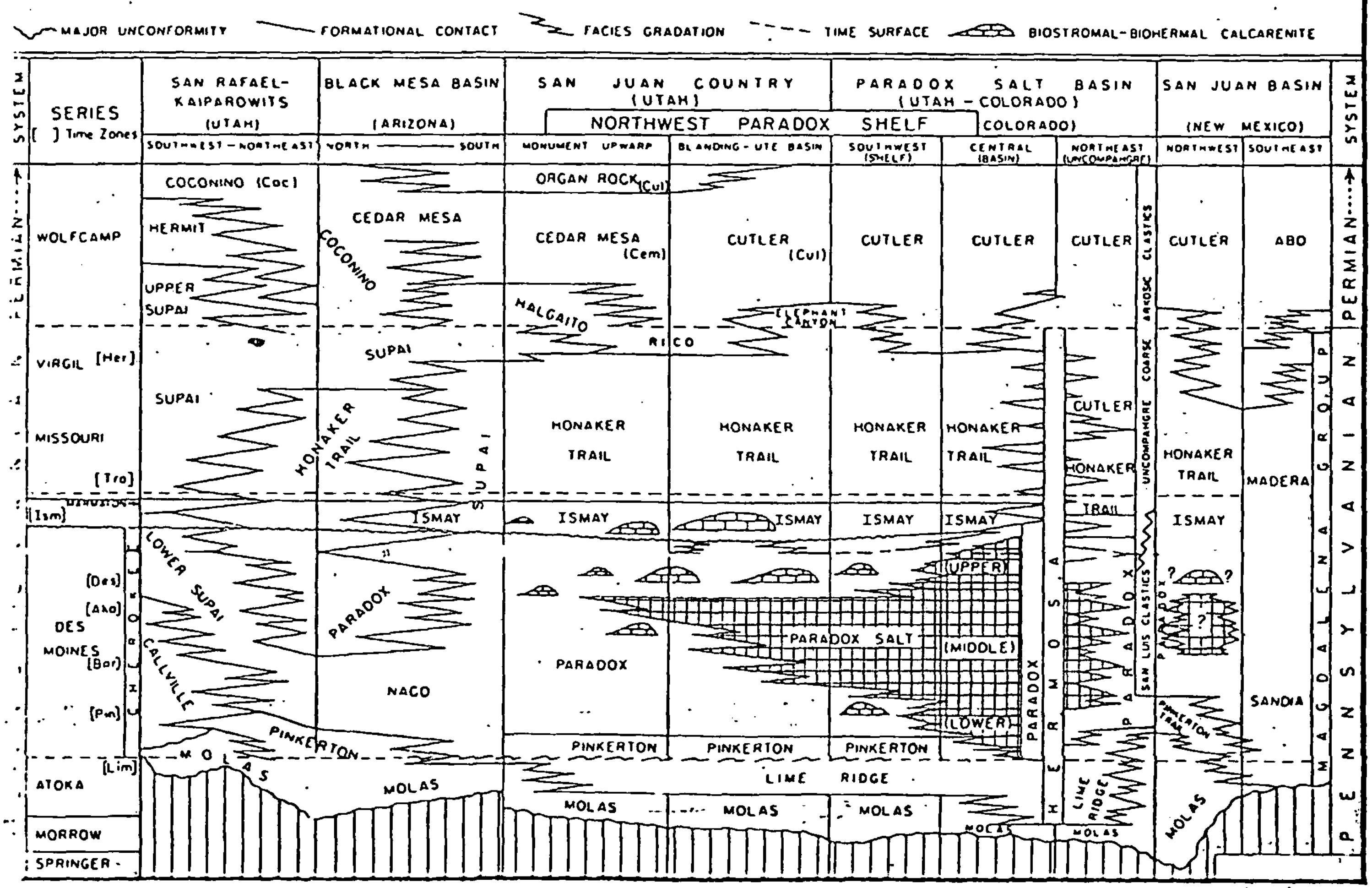

FIGURE 6—Pennsylvanian correlations—Four Corners region.

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Pennsylvanian Assignment.

Do One

1 Discuss in a historical way the evolution of understanding the Pennsylvanian deposits of the Grand Canyon. Start with the earliest age determinations and continue to the present. Stress the evidence that has brought the change. A chart showing the progressive nomenclature will do for an illustration. A good starting reference is
Geology of the Grand Canyon ; McKee paper 1975

2. Discuss briefly the tectonic relations and related stratigraphy of the Eagle or Central Colorado Basin (same thing). Express an opinion as to why this basin contains no significant potash salts. The Big Red Book gives references.

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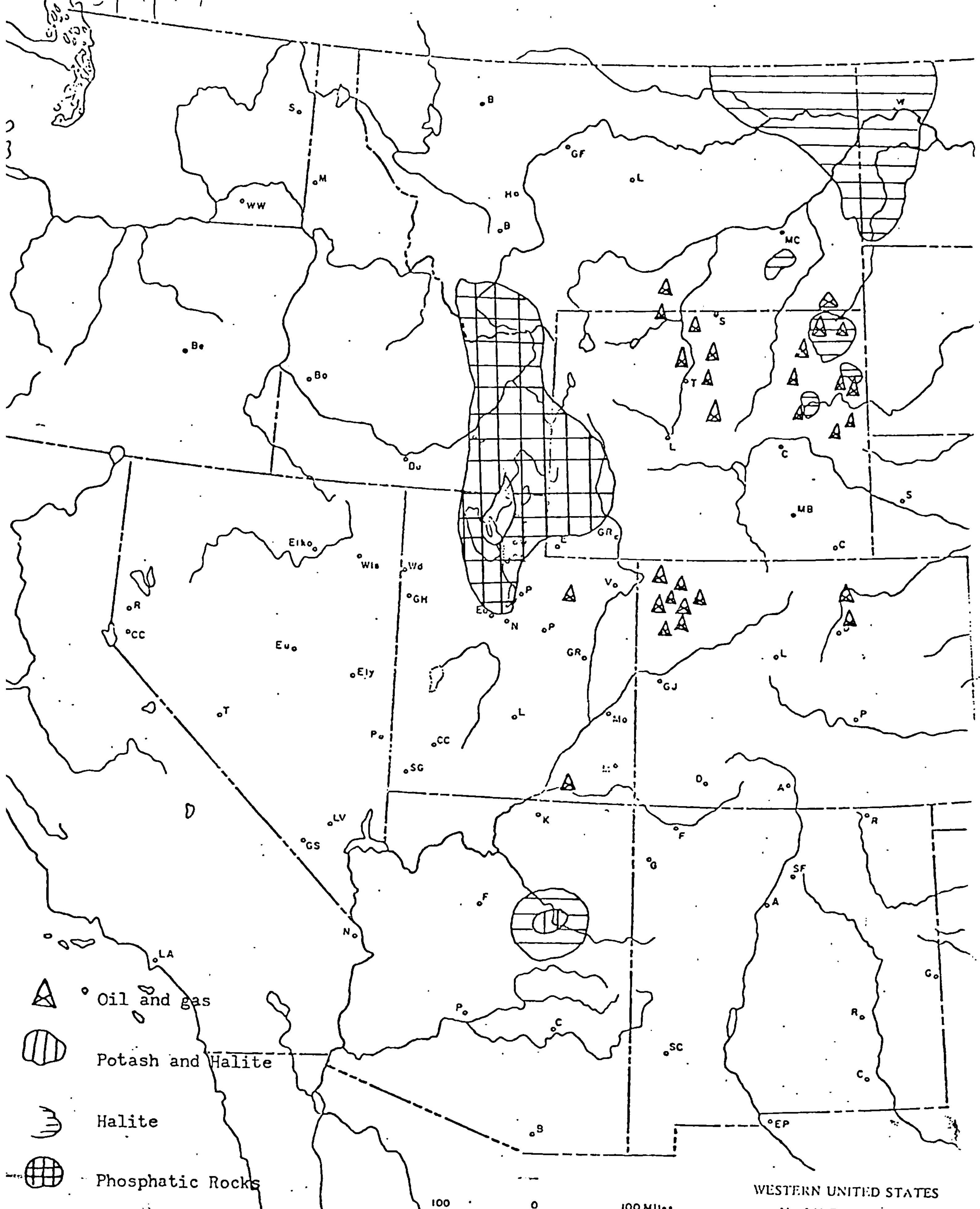
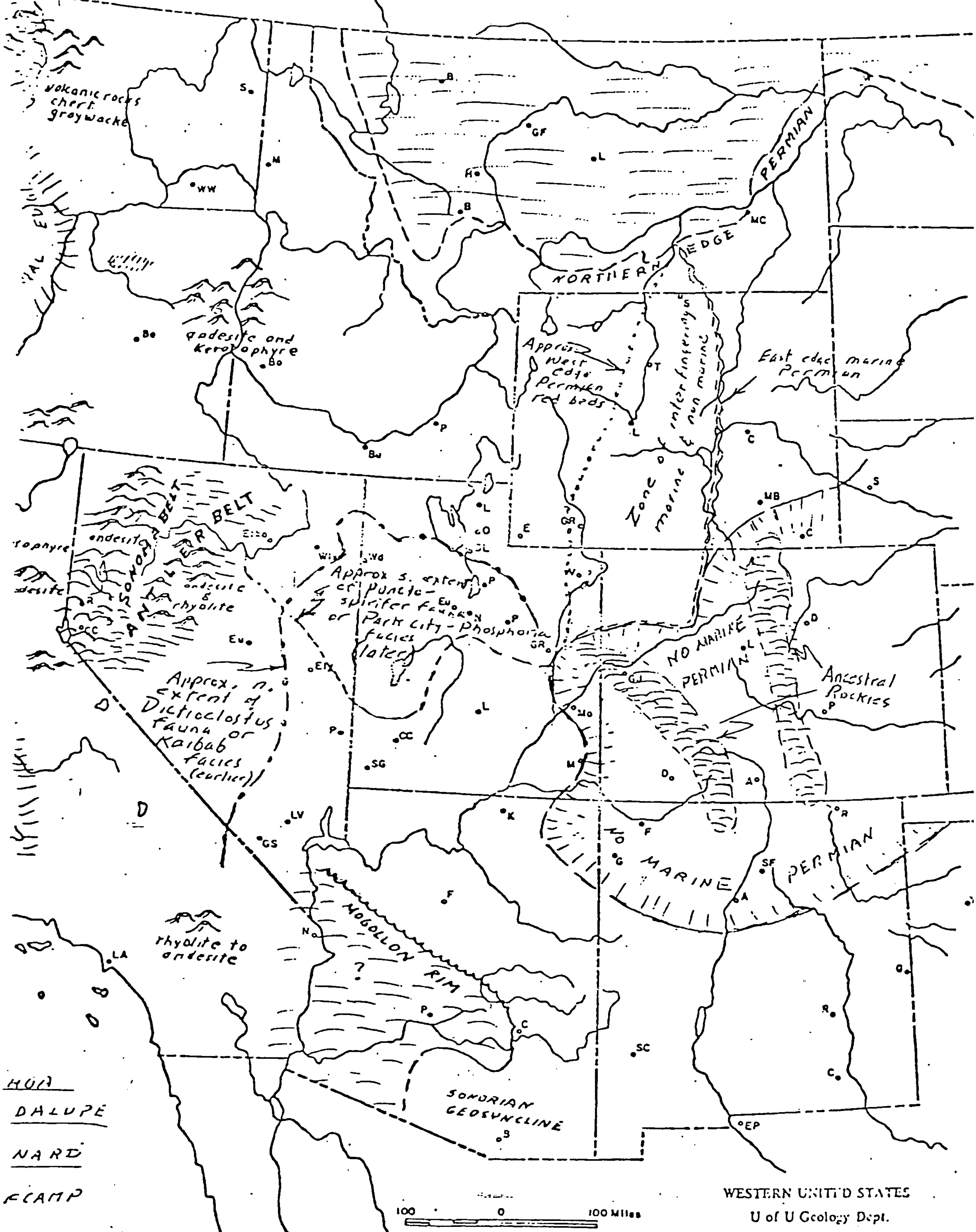
Permian System



Compiled from: Geologic Atlas of the Rocky Mountain Region U. S. A.
By Rocky Mountain Association of Geologists, 1972.

AND OTHER FEATURES



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7/9/79

Generalizations

Paleozoic Stratigraphy of the Western Interior

PERMIAN

I. Historical

The Permian System was established by Sir Rodrick Murchison (founder of the Ordovician) in 1841. The standard world section is on the west flanks of the Urals, European Russia. The Permian of Great Britian and Western Europe is sparcely fossiliferous continental red beds and sandstone. Since access to U.S.S.R. localities has been restricted not many non-Russian geologists have been able to study the type Permian

Knowledge of North American Permian has been slowly attained because the best exposed and most complete sections are in the Southwest and were not explored until relatively later than other Paleozoic outcrops. Texas has provided the standard U.S. sections and names - Wolfcamp, Leonard, Guadalupe, Ochoan - the last named is gradually falling into misuse as it appears to be represented only in the Oklahoma, West Texas, east New Mexico area and may span an insignificant part of the period. In any event, western sections can be dealt with satisfactorily in terms of the 3 older epochs.

II. Tectonics and Paleogeography

A. Dominating tectonic elements of the Permian were: the Ancestral Rockies in process of steady erosional destruction; the Canadian Shield, now expanding somewhat southward; the Cordilleran Geosyncline, with the Antler Belt undergoing erosion and the Sonoma Orogeny creating mountains in western Nevada and eastern California late in the period; the Transcontinental Arch which was rather low with a widening "sag" in Arizona and essentially obliterated in eastern Colorado, Nebraska and Kansas.

B. Negative and positive elements that originated in the Pennsylvanian continued into the Permian but highlands were greatly lowered and basins were essentially filled to create a less diversified landscape. The Paradox, Eagle, Oquirrh and other basins were gradually obliterated and not much in evidence after Wolfcamp time, but the Uncompahgre and Frontrange elements continued to be emergent throughout the period.

C. Great volumes of clastic sediments derived from the Ancestral Rockies in addition to filling adjacent basins pushed back the epicontinental seas and created wide zones of interfingering marine - non marine facies.

D. Many scattered outcrops of volcanic rocks usually associated with metasediments and clastic derivatives are known in the Pacific States. These suggest archipelagic islands and a rapid accretionary action in the Pacific coastal area. Much of this Permian material may rest on oceanic crust. Little intrusive material or granitic debris is known, however.

E. Growing, but still not positive evidence for a late Paleozoic. Early Mesozoic mountain-building episode, the Sonoma Orogeny, is accumulating. (Date-240-220 my). Chief effects are eastern California, almost overlaping into the present Sierra Nevada and into Northwestern Nevada. (GSA, Vol 83, p. 1996).

II. Lithology and Sedimentation

A. The Permian Period has an estimated duration of 55 million years during which a large volume of diversified rocks was produced. Estimated percentages of dominant sedimentary rock types in decreasing abundance are: Sandstone, both marine and non-marine,45%, limestone and dolomite,20%; fine clastics, much of it red, 15%; phosphorite, 5%; chert, 5%; evaporites, mostly gypsum, 10%. This tabulation ignores the volcanic derivatives that make up large but unknown fractions in Nevada, California, Washington and Oregon. Exposures are widespread in the Colorado Plateau, Rocky Mountains and Great Basin and are relatively more abundant than previous systems in the Pacific states.

B. Cyclic deposition of the Oquirrh-Ely type continues approximately to the end of the Wolfcamp. The Wolfcamp portion of the Oquirrh is not significantly different from the Pennsylvanian part in apprearance, lithology and biostratigraphy.

C. Deposition of "red beds" reaches a maximum and reflects acellerated erosion of the Ancestral Rockies and deposition in alluvial plains and near-shore marine environments around them (Upper Fountain, upper Maroon, Cutler, Supai).

D. Extensive blanket sandstone formations mostly aeolian are characteristic of later Permian time (Weber, Cedar Mesa, Coconino, DeChelly). An exception is the White Rim which seems to be a barrier bar formation. Wind directions were predominantly to the south.

E. Gypsum is locally abundant (western North Dakota, adjacent Montana; northwestern Texas, adjacent New Mexico, Arizona, and southern Nevada) and there are halite deposits in North Dakota, Montana, Wyoming, Nebraska, Colorado, Kansas, Oklahoma, New Mexico and Texas. Potash is found in Arizona, New Mexico and Texas.

F. Dolomite again is an important rock type especially in the Leonard and Guadalupe formations of Nevada, and southwestern Utah.

G. Chert is very abundant, possibly constituting a greater relative volume than in any previous period. This may or may not correlate with a strong resurgence of sponges in the late Permian or with the strong volcanism in the nearby Pacific belt.

H. Conditions were favorable, in the late Leonard-Guadalupe for the accumulation of great volumes of phosphate-bearing rock (phosphorite) in the late Permian of western Montana, western Wyoming, northern Utah and northeastern Nevada (Park City and Phosphoria Formations). Upwelling of Pacific currents across shallow shelving banks in connection with the right marine organisms seem to have been responsible for the phosphate deposition.

I. Important breaks in the sedimentary record, as is so often the case, occur within rather than between the series. Thus there are important intra-Wolfcamp, intra-Leonard and intra-Guadalupe unconformities of wide extent.

IV. Biostratigraphy

A. No significant faunal break exists between the Pennsylvanian and Permian of the Western Interior but there is a distinct and abrupt change within the Permian at about the level of the Wolfcamp-Leonard boundary. Below this level the approximate order of decreasing abundance of obvious animal fossils is: fusulinids, brachiopods, bryozoans, molluscs, echinoderms. Above the discontinuity the approximate order is molluscs, brachiopods, bryozoans, corals, sponges and echinoderms. The displacement of brachiopods by molluscs marked an important and permanent change -- brachiopods never regained their previous status. Fusulinids declined almost to the vanishing point in north America but continued to prosper and evolve in Asia.

B. For the early Permain the faunas show connections with the mid-continent. Later there are migrations from the southeast (Texas-New Mexico) and from the northwest (Siberia-Canada). The southeast biofacies is marked by the large productid Dictioclostus, the northwest facies by the spiriferoid Punctospirifer pulcher.

C. Much interest attaches to the faunas of the phosphate-bearing beds. Brachiopods such as Lingula and Orbiculoidea, with phosphatic shells are very abundant. Many areas yield a peculiar depauperate fauna of small gastropods, pelecypods and cephalopods. All these plus sharks and perhaps numerous soft-bodied animals contributed to the phosphate content of the Park City and Phosphoria.

D. With the spread of continental deposits in the vicinity of the Ancestral Rockies, a large number of reptiles, amphibians, and fresh-water fish became fossilized. Numerous animal tracks are found in the Coconino and DeChelly but little else. Bones of a variety of vertebrates come from the Halgaito and Organ Rock members of the Cutler, chiefly in the Monument Valley area. The age is late Wolfcamp to early Leonard.

In the marine beds of the Great Basin and northern Rocky Mountains vertebrates are represented by a variety of sharks including the peculiar Helicoprion which left spiral tooth rows in relative abundance.

E. The record of Permian plant life in the Western Interior is faily important. A large flora from the Hermit Shale is regarded as one of the earliest to show semi-arid adaptations. Smaller collections from the Halgaito and Organ Rock near the Four Corners area show need for a fair amount of moisture while the plants from Texas and Oklahoma Permian seem to have grown under well-watered conditions. It has been suggested that there paleobotanical differences were due to orographic influences of the Ancestral Rockies.

V. Economic

1. The Phosphoria and Park City Formations are major producers of phosphate rock in Montana, Idaho, Wyoming and Utah. Utah, for example has an estimated 29,000,000 tons of ore with a grade of 31%+ P_2O_5 and 3,700,000,000 tons of lower grade, cut off point is about 18% P_2O_5.

2. Oil is produced from the Permian in many fields especially in Wyoming. Oil fields seem to be related to the phosphate-rich facies. Although secondary to Pennsylvanian in Utah there is Permian production in the southern high plateaus, western San Rafael, Rangely, Ashley Valley.

3. Non-combustable gas including carbon dioxide has been found in Permian sandstone of the Colorado Plateau--this could be a great future resource.

4. Asphaltic (tar) sand in large amounts occurs in the Permian of the Colorado Plateau. The White Rim, Coconino, and Cedar Mesa are impregnated over wide areas.

5. Gypsum is present in Permian of almost all Western Interior states. Halite is only slightly less abundant. Potash is mined from the Texas-New Mexico region.

6. The Park City Formation is the host rock for over 75% of the metalliferous deposits of the Park City mining district and there are ore bodies in the Permian part of the Oquirrh at Bingham and in the Elv Limestone at Ely. Very extensive deposits of sedimentary copper are found in late Permian strata of Oklahoma.

7. Local building stone, especially Lykins of Front Range.

VI. Boundaries

See Pennsylvanian summary for Mississippian-Pennsylvanian boundary problem.

The proper placement of the Triassic-Permian boundary is a classic of geology because it was a time of great physical and biological change and transition. Almost everywhere there seems to have been a complete withdrawal of the seas so that some part of the very late Permian of very Early Triassic is missing. At no place is it certain that we have continuous deposition across the boundary. Fossils of most importance in reference to the problem are conodonts, ammonites, and brachiopods. Great Basin sections may be of importance in locating the proper horizon.

VII. Scenic

Permian rocks are widely exposed in the Colorado Plateau where they have attracted wide fame because of their striking coloration and unusual erosional forms. Among the areas specifically designated as scenic or scientific attractions are:

Canyonlands National Park: The topographically lower and more southerly sections of Canyonlands N.P. show exposures of the Cutler, White Rim and Cedar Mesa formations. Banded gray and red rocks of continental origin carved in bare-rock cliffs and monuments dominate the landscape.

Canyon DeChelly: This is the type area of the DeChelly Sandstone. Vertical canyon walls with "cliff dwellings" are the attraction.

Natural Bridges N.M. Three great national bridges - Kachina, Sipapu, and Owachomo -- in west-central San Juan Co., Utah are carved by meandering incised streams from the Cedar Mesa Sandstone.

Monument Valley: Higher members of the Cutler Formation are carved into a variety of massive mesas, buttes, and spires. The shafts of these are DeChelly sandstone, the flaring bases are Organ Rock Formation.

Upper slopes and rim rock of Grand Canyon: Three strikingly different formations of Permian age rise above the Redwall cliffs of Grand Canyon - Supai, Coconino, Toroweep-Kaibab.

The block U on the hillside above the campus is on the Park City Formation. also the rocks at the mouth of Mill Creek are in this same formation.

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W. Lee Stokes, 1974

3/2/79

Non-Marine Stratigraphy in the
Late Paleozoic of the Western Interior

HISTORICAL

Western Europe and the British Isles were well suited as training grounds for the earliest generation of geologists. Few other areas of similar size could have furnished adequate types for the divisions of the standard geologic column that eventually came into world wide use. Strange as it seems, the British section even provides a foretaste of the non-marine late Paleozoic and early Mesozoic sedimentation and stratigraphy that dominate so much of the western United States. There are wind-blown sands, red beds and fluvial deposits that match on a reduced scale counterparts thousands of miles away.

But geology and geologists did not progress directly from western Europe to western North America. There was a lengthy interlude when American geologists were getting their feet on the ground on the eastern seaboard. True, the rocks and fossils they found in the Appalachians weren't too different from those of the Old World. And well they might be for continental drift had managed to leave pieces of old Europe attached to North America and pieces of old America attached to Europe. What was lacking in eastern geology, except for the entirely atypical filling of the Triassic basins, was a good suite of non-marine rocks. So a generation or two of geologists were trained in looking at, mapping and correlating the wonderfully fossiliferous, limy formations of the Appalachian geosyncline.

When these marine-oriented geologists came into the west, particularily the Colorado Plateaus, they cannot be blamed for looking at the great non-marine sections in the light of their training in the east. The mistakes they made were natural enough but have been difficult to correct once embedded in official reports and perpetuated in textbooks.

Basic Difficulties

Fewness of Fossils: It is difficult for a geologist accustomed to working in the Jurassic and Cretaceous of western Europe to adjust to the fact that many formations elsewhere are nearly or entirely barren of fossils. In the Colorado Plateau great thicknesses of rock, even entire formations, have not yielded a single fossil of any kind. The remains that come to light in other formations are not the kinds on which reliable correlations can be based. To be more specific, the detailed subdivisions of late Paleozoic and early Mesozoic rocks that have been made are based chiefly on abundant marine fossils occurring in well-documented lineages and assemblages; fossils found in equivalent continental beds are naturally of land living types that may be totally unfamiliar ar at least not yet assigned any specific value in geologic dating.

This paucity of fossils is a fact of life with which many workers in continental beds are not prepared to deal. Their college training was with too many well-formed brachiopods, glistening trilobites, three-dimensional pelecypods and fresh-looking gastropods. Most of these must be forgotten when one sets foot in the continental deposits of the western wastelands. Now, our hero must look for ostracods that may appear like grains of sand, for charophytes that look like nothing he has seen before, for limonite encrusted bones and scattered teeth that seldom if ever come close to making up a complete animal. He must look for tracks and trails and ought to be an expert in trace fossils. A branch, a twig, a chip of wood may solve his correlation problem but he must recognize wood that has structure and wood that has none. He must learn which types of sediment may have spores or pollen and be willing to work long and hard to see these through to final determinations. He must learn to recognize in the distance those protruding ledges of limestone among the red beds underfoot which are likely to yield fossils. And to observe those lenses of light-colored marlstone deposited in temporary lakes among ancient dunes or flood plains and even for mere streakes of carbonaceous material in otherwise barren silt or sand. And he must learn to be wary of the abundant pseudo-fossils that nature has scattered in the non-marine rocks to deceive the unwary.

Once a geologist learns what to expect and what to look for he may find that his formations are not really unfossiliferous and his correlations not impossible.

Scarcity of Key Beds: With or without fossils the geologist working in continental sediments must never forget key beds. Every bed and every discontinuity between beds has a potential value in correlation. Its use depends on how far it can be traced. Every bed and every bedding plane must have finite lateral dimensions but little is known about how far either of these may extend. The thicknesses of uncounted numbers of sedimentary beds are known but a complete 3-dimensional study of a single bed is probably not to be found. This is, of course, excusable because the data for such a study would be almost impossible to obtain.

Faced with the multiplicity of beds and bedding planes seen in random cross sections it is small wonder that most geologists wouldn't recognize one of these if he should see it again at another time or place. Only a few beds will make an impression strong enough to be retained for later recall. But an effort must be made. The best field stratigrapher I know, literally memorizes as much of a section as he can. Another takes innumerable photographs. Something like this is the only way that the less-than-extraordinary bed or bedding plane can be used as a marker. After all, isn't this the way fossils are used, by search, memorization, and comparison. The geologist referred to above has had great success in recognizing types of chert in the marine sections, these can be traced over great distances.

The best key beds are in marine or at least subaqueous sections. Above all a key bed, once laid down, must remain intact or it is of no value. Thus the bentonite beds of the Ordovician of the Appalachians and of the marine Cretaceous of the Western Interior are proving of surprising value in correlation. All peculiar, eye-catching, features such as chert, unusual mineral or rock fragments, odd bedding, weathering characteristics and striking colors should be evaluated in working out correlation problems.

This adds up to a dismal picture for finding key beds in most continental sections. How can a key bed maintain continuity under subaerial erosion and deposition? Fluvial deposits originate through the reworking and shifting of mud and sand while aeolian deposits are built chiefly of dunes by a constant forward shifting of sand with almost daily changes.

Volcanic ash, the raw material of the best key beds may fall and blanket sea bottom and land surface alike but the chances of its remaining intact in the former environment are naturally much better than in the latter. It requires little imigination to visualize the rapid deflation and destruction of an ash bed under strong winds in the desert environments.

Lack of Radiometric Dates: Sedimentary rocks yield relatively few radiometric dates compared to igneous and metamorphic varieties. Considering the late Paleozoic and Mesozoic rocks of the Western Interior the chances for obtaining satisfactory dates from non-marine beds are less than they are for marine beds. This is chiefly because so many good dates have come from the Cretaceous bentonites. No good dates seem to have been obtained from the late Paleozoic sections associated with the Ancestral Rockies. This may be because no one has seriously searched for suitable materials.

FAVORABLE METHODS

In spite of a general lack of f ssils, key beds, and material for radiometric dating there are several favorable materials and techniques available. One of these is paleomagnetism. Workers with this method search out the clastic red beds and have reported good results from the Colorado Plateau and adjacent regions. Best results have so far been obtained in the Moenkopi Formation. It is known from work elsewhere that the Permian was almost if not entirely reversed in polarity but there are numerous reversals in the Triassic and Jurassic. Hopefully these polarity epochs can be traced between various red bed basins to give better correlations than now exist. For example the Colorado Plateau section can possibly be related to the Wyoming section, the position of the Wingate Sandstone can be determined, and the exact horizon of the Permian-Triassic boundary can be fixed. The time values of unconformities below the Shinarump and Higham needs to be settled also.

It appears that reversals in the Triassic if properly evaluated and traced well separate time planes equal to these established by paleontology.

Another method with potential value in correlation is that of paleocurrent indicators. Included here would be cross-bedding of both aeolian and aqueous origin as well as lineations, ripple marks and any other features that reveal the direction of depositing currents or simply of the paleoslopes. Most marine beds, particularily the carbonates are devoid of good directional indicators but the continental red-beds and light-colored sandstones of the late Paleozoic are literally full of them. No good illustrations of their use in correlation come to mind but situations can be imagined where paleo-directional data would be of value. It might, for example, prevent a correlation of units with opposing flow directions and favor correlation of those with parallel ones. A perceptive geologist might be able to discern a wind-blown sand from a beach sand and thus not correlate them.

Finally correlation of key unconformities seems to have as much or more value than correlation by key beds. One result of subaerial erision, as any geomorphologist knows, is the creation of various types of level or gently-sloping surfaces. Thus these are extensive gravel-coated or armor plated surfaces, paleopediments, duracrusts, and deflation-controlled surfaces (such as Great Salt Lake Desert). These may cover thousands of miles and are more extensive than beds either above or below. See W. L. Stokes, 1968, Multiple parallel-truncation bedding planes - a feature of wind-deposited sandstone formations: Jour. Sed. Pet. vol. 38. no. 2, and G. N. Pipiringos and R. B. O'Sullivan, Chert Pebble Unconformity at the top of the Navajo Sandstone in Southeastern Utah, Four Corners Geol. Soc. Guidebook 8th Field Conf. 1975, p. 149. Latest: "Principal Unconformities in Triassic and Jurassic rocks, Western Interior, United States" Pipiringos and O'Sullivan, USGS Prof Paper 1035-A, 1978

Example - the Cutler Problem

The Cutler Formation, extensively exposed in Utah, Colorado, New Mexico, and Arizona is a complex assemblage of diverse depositional units the mapping and understanding of which calls upon all the techniques of non-marine stratigraphy. Donald C. Baars foremost student of the section characterizes the Cutler Formation as a "battleground" of marine and non-marine influences and says: "If the rocks were not to well exposed to view they would be a stratigraphic nightmare." To justify this assessment he points out that every formation in the Permian System either pinches out or changes facies 100 percent within Canyonlands.

The stratigraphic complexities of the Cutler Group arise from the paleogeography of Permian time. The prominent northwest trending Uncompahgre Uplift lying to the east and northeast of the Paradox Basin shed large volumes of arkosic, coarse grained sediments into the near-by basin where the underlying Paradox salt was in process of restless motion. Once delivered to the lowlands the heterogeneous mixture of clastic material was subject to a variety of influences well reflected in the complex facies of the Cutler. Fluvial action of the type characteristic of arid or semi-arid climates distributed material from the mountains across alluvial fans into flatter areas with gradual reduction in the size of the component fragments. Parallels with modern arid basins which display sloping alluvial fans grading into central playa-lake basins come to mind. Wind was an ever-present agent of erosion and deposition and probably had maximum effects in the lower reaches of the alluvial fans where sand and the finer grades of sediment came to rest. The analogy with interior basins breaks down for the earlier and later parts of the Permian when the ultimate base level was a shallow shelving sea to the south and west. The near-shore and coastal bands provided yet another anergy-charged environment for Culter sediment.

In summary then the Cutler contains 1) Coarse bouldery alluvial fan deposits, 2) fluvial deposits ranging from course to fine, 3) playa or mud-flat deposits of shallow water origin, 4) coastal dunes or offshore bars, 5) shallow marine deposits, and 5) deeper marine deposits mainly calcareous. Not all of these can be fully discussed but the problem of the massive sandstones deserves further notice.

Wind or Wave?

No textbook is complete without an illustration of cross-bedded sandstone from the Colorado Plateau. The Navajo Sandstone is a favorite subject but there are others, no less spectacular. All of the thicke sandstone members of the Cutler formation are cross-bedded througnout and no description or discussion of them is complete without at least a comment on its origin and paleogeographic significance. Only two modes of origin seem possible and opinions of the experts seem to be well divided between them. One group favors an aeolian origin, the other an aqueous one. Is the cross-bedding formed by wind or by water? One thing is certain, there is a consistency of dip direction over many thousands of square miles and throughout many successive units that must weigh heavily in any considerations of the problem. This consistancy is shown by the compilation of cross-bedding directions for the Weber, Cedar Mesa, White Rim, DeChelly and Coconino by Stewart and Poole (see attachment). There is no doubt that they favor the wind origin. On the basis that the wind direction is indicated these workers postulate a position near the equator at the time of formation.

I cast my vote for the aeolian hypothesis for the bedding of <u>most</u> of the Permian blanket sandstones. My reasons are not so much because of the Cross-bedding as because of the planes between the cross-bed sets. These I have called truncation bedding planes because their formation requires the removal of material from the tops of previously thicker beds. These planes are so flat and parallel that they can, in my opinion, be explained only by the removal of material down to a level determined by the water table, a process impossible to carry out in the aqueous environment.

Colleagues disagree. Chief proponent of an aqueous origin is D. L. Baars. He believes the cross-bedding "supplemented by the ripples and the convolute bedding" strongly suggest a subaqueous origin. He makes an almost air-tight case for the White Rim being an "off-shore" bar by citing the discovery in it of marine fossils. It is most difficult, however, to imagine a water origin for the Coconino and DeChelly with locally abundant animal tracks (very long-legged if wading in water!). The truth may lie between the extremes. If nearby marine intercelations or equilivants can be proven the sands may also be marine but if there were no seas anywhere in the near vicinity why fight the aeolian hypothesis?

Salt Stratigraphy

Salt is the most tricky material with which the structural geologist and stratigrapher has to deal. Being easily soluble it seldom appears at the surface and has thus been passed over in most surface mapping. The great salt basins have mostly been discovered in the course of exploratory drilling and exact structural and stratigraphic details of many of them have been revealed by mining. Whenever encountered the stratigraphy of salt deposits has been surprising. Examples are the Polish-German plains underlain by Permian salt, the Gulf Coast Salt Dome Province of the United States and the Utah-Colorado Paradox Basin.

The salt section of the Paradox Basin was discovered in drilling for oil near Crescent Junction, Grand County, Utah. The fact that the salt section is highly disturbed and that internal movement and subsequent solution are together responsible for large northwest-trend valleys came with continued drilling and mapping in all parts of the Basin. Potash salt had been discovered in some of the early cores and this led to mining ventures chief of which is the 2700-foot shaft of Texas Gulf Sulfur Company at Potash, Utah. Incidentally, the shaft is no longer operated as a regular mine, the potash is extracted by solution methods.

Although exploration of the salt structures has been disappointing there were important discoveries on the southwest margins of the basin in the Four-Corners area. Actually there are Pennsylvanian wells in each of the four adjoining states which derive oil from the Pennsylvanian rocks. The oil pools are directly associated with the Paradox salt section in that they are in rocks equivalent in age to the salt and are in reefoid sediments on a sill or threshold leading into the salt basin. This relationship has stimulated exploration for other similar environmental situations around the margins of the salt.

The great central salt deposit of the Paradox Basin is characterized having originated by cyclic deposition and each cycle has been rightly or wrongly termed a cyclothem. The ideal cycle consists of nine members limited above and below by black shale. Thickness of an average cycle is about   feet. There are 29 recognizable cycles in the center of the basin, each has been given a number and correlated widely using gamma ray-neutron logs. What are almost certainly yearly rings or varves are found in the salt beds. Fossils are not lacking but are obscure; included are conodonts, small molluscs, brachiopods, pollen and plant fragments. The cyclic replenishment of saline brines must have been by way of certain inlets or sags around the southern margin of the basin and the whole problem of repetitions sedimen cannot be treated independently of the whole great problem of Pennsy vanian sedimentation best typified by the coal bearing cyclothems. (Best review is Stratigraphy and Tectogenesis of the Paradox Basin, by Ernest Szabo and Sherman A. Wengard in Four Corners Geol. Soc. Guidebook, 8th Field Confer. Canyonlands, 1975.)

Interaction of Salt and Sedimentation

Basic to understanding all salt stratigraphy is the fact that salt has a lower specific gravity than ordinary sediments and therefore tends to rise to a position of equilibrium. Gravity is thus the motive force for salt domes and anticlines. The question as to why some salt deposites tend to develope flow structures while others do not has not been answered. Some initiating trigger such as differential heat flow faulting or igneous intrusions probably operate in the subsurface while differential loading or unloading may operate at the surface. These factors if properly understood will probably explain why certain deposits remain static over long periods while others begin to move almost immediately. The problem of dating salt movements is not greatly different from dating mountain building pulses. Unconformities are created and can be dated and in some cases fragments from uplifted beds are incorporated in the flanking sediments.

From structural-stratigraphic evidence it is known that the Paradox salt began to push upward in Pennsylvanian time probably when the overburden of limestone was less than 300 feet. Faulting in the basement could well have been the triggering mechanism. In any event the rise was rapid enough to keep pace with sedimentation. There is no evidence that the salt anticlines became more than low hills but they did act as obstructions which diverted fluvial action and created semi-independent basins of deposition for along period of time. Basining was acellerated by downsinking between the anticlines as the salt slowly drained into the uplifts. This action was particularily strong during late Pennsylvanian and Permian time as the Culter red beds were deposited. Thus one well drilled in a salt core may go through 12,000 feet of salt while one a few miles away will go through an equal thickness of red beds.

W. Lee Stokes
1976

3/3/77

Ancient southern margin of North America

Philip B. King, U.S. Geological Survey, 345 Middlefield Road, Menlo Park, California 94025

What was the configuration of the ancient (pre-Paleozoic) southern margin of North America? This is shrouded with uncertainty, because we must peer under the blanket of Mesozoic and Cenozoic deposits that cover the Gulf Coastal Plain (especially beyond the point where the Early Jurassic Louann Salt develops coastward) and thence to the Paleozoic orogenic rocks and structures beneath them. The speculations offered here are an extension of several regional reviews of the Ouachita orogenic belt that I have made recently (King, 1975, 1976).

The Paleozoic southern margin of North America is dominated to the southeast by the Appalachian orogenic belt and to the south by the Ouachita orogenic belt that extends across the south-central states as far as Mexico, and perhaps beyond (Fig. 1). As we shall see, these orogenic belts exhibit important variations along their courses, but the continuity of the whole orogen during Paleozoic time is attested by the chain of foredeeps that lie along the edge of the craton for the entire distance: the Black Warrior basin in the angle between the Appalachian and Ouachita belts in Alabama and Mississippi, the Arkoma basin north of the Ouachita Mountains in Arkansas and Oklahoma, the Fort Worth basin to the southwest in north-central Texas, and the Val Verde basin north of the Marathon region in western Texas.

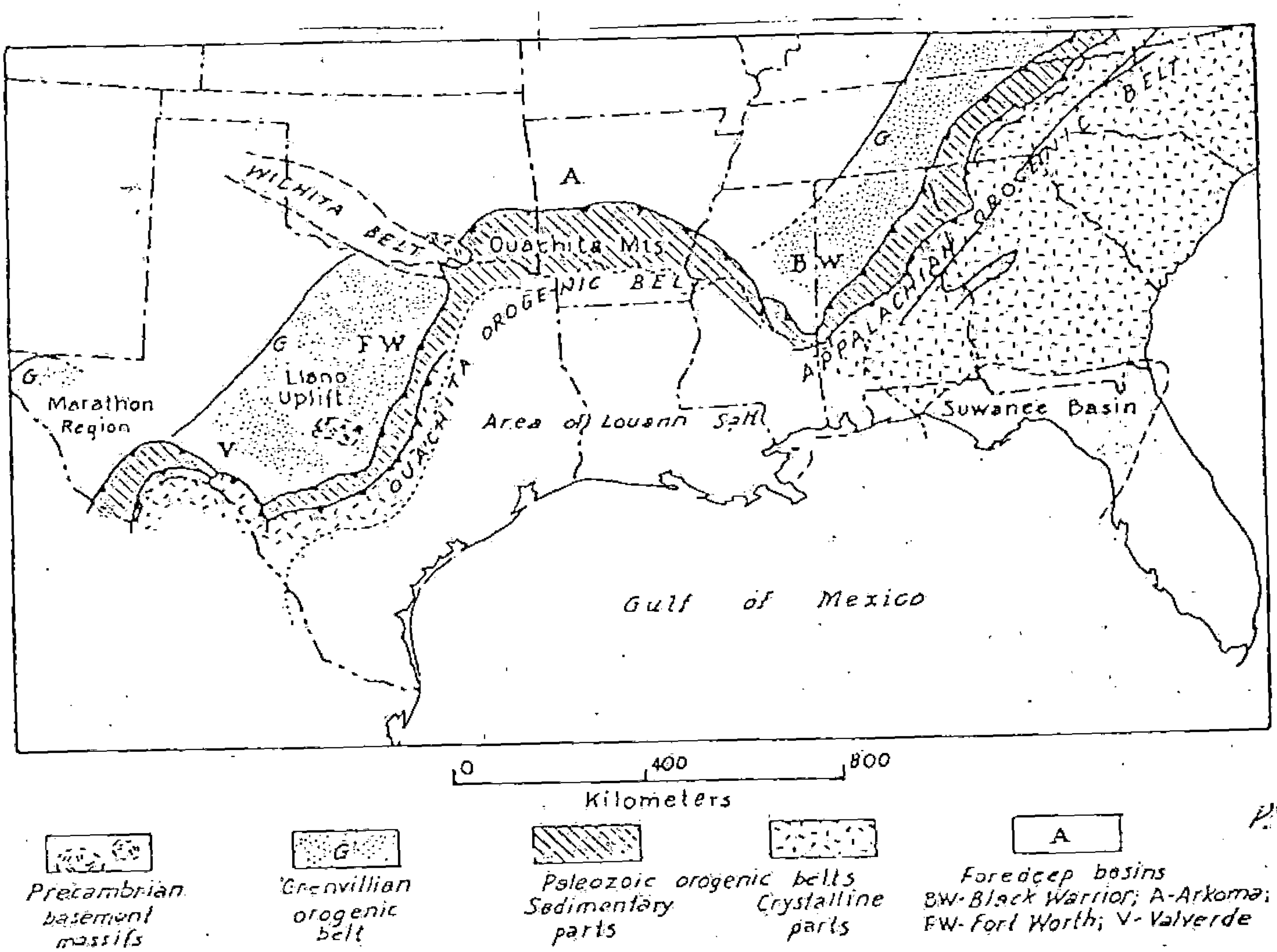

Figure 1. Map showing pattern of Paleozoic orogenic structures in southern United States.

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867-C. R11 Smith & Ketner READ IT!

formational events are removed, the Roberts Mountains allochthon appears to consist of shuffled platelets separated by thrusts that are not greatly folded. In places these thrusts are subparallel to the bedding within the platelets, and in other places they bound platelets having highly crumpled and tightly folded beds. Much, if not most, of the major folding of these rocks can be attributed to Mesozoic compression. It is unlikely that these broad thin sheets were pushed great distances against friction and the force of gravity. On the contrary, they must have been drawn by gravity.

No mechanism known to us, however, appears to be completely satisfactory for placing these rocks into a position from which they might have moved by gravity. Any plate tectonics model that involves development of melanges or of high pressure–low temperature metamorphism and the incorporation within the allochthon of extensive ophiolite suites is unlikely, as little evidence of these effects has been found. Poole and Desborough (1973), however, have reported small elongate bodies of alpine-type serpentinites which they infer to be detached fragments of Paleozoic or older upper mantle incorporated into oceanic crustal and sedimentary rocks during lithospheric plate convergence. Any model that requires extensive lateral push to raise the rocks into a position for sliding is also unlikely because of the observed condition of the thin plates. If obduction (a little understood process) could have placed the rocks into position without excessive lateral push and could have peeled off the top of the eugeosynclinal deposits so as not to leave remnants of extensive ophiolite suites and high pressure metamorphic rock, it might serve as a helpful method, as recently suggested by Poole (1974) and Burchfiel and Davis (1975). We are reluctant, however, to accept any current plate tectonics model as affording a completely satisfactory solution to all the problems involved in the movement of the Roberts Mountains allochthon.

We are equally reluctant to accept the notion that the allochthonous rocks simply slid, under the influence of gravity, off a static uplifted area. Using this hypothesis, the area tectonically denuded of Ordovician, Silurian, and Devonian rocks that rode eastward on the Roberts Mountains thrust would have to be at least as wide as the upper plate and probably much wider, because repetition of stratigraphic units is characteristic of that thrust plate. The combined width of the allochthon and the area of tectonic denudation from which it came must be at least 220 km (136 mi), double the provable minimum width of the allochthon. A static sloping surface that is steep enough to permit the movement of the platelets by gravity alone from the crest of the area of denudation to the east edge of the upper plate entails an almost impossibly high elevation of the crest. The existence of such an eastward-sloping surface would imply a corresponding west slope of equal width and another, western allochthon corresponding to the Roberts Mountains allochthon somewhere west of the area of denudation. More liberal (and more plausible) estimates of the east-west extent of the Roberts Mountains thrust correspondingly imply an even more liberal width of the area of denudation and a completely unrealistic elevation of the area of denudation. We therefore feel that this concept is untenable.

We find it more plausible and more in accord with the evidence to envision a succession of crustal waves, each of modest proportions, that migrated from the west edge of the denuded area to the east limit of thrusting. According to this hypothesis, the component platelets of the allochthon rode independently down the forward slope of the advancing waves like surfboards and eventually beached themselves in random order as the crustal waves lost their initial amplitude. This hypothesis is similar to a concept for nappe transport suggested by Wunderlich (1973, p. 283) but considered improbable by Lemoine (1973, p. 211-212). Probably the growing mass of the upper plate contributed to the progressive decrease in amplitude. The Roberts thrust is thus seen simply as the surface on which the upper plate accumulated by piecemeal accretion.

Late Jurassic to Early Cretaceous uplift, plutonism, volcanism, folding, and thrusting constitute the major tectonic events of the Carlin-Pinon Range area. The principal tectonic events of the Late Devonian-Early Mississippian orogeny, although they were of first-order magnitude, took place mainly to the west; and their expression in the Carlin-Pinon Range area located at the east limit of thrusting was relatively subdued. Although the temporal limits of the Mesozoic tectonic events cannot be fixed exactly in the Carlin-Pinon Range area, these events occurred within the time span of similar activity elsewhere in the region (fig. 3). Westward-directed thrusting and overturning of some beds occurred during Mesozoic time in this area in contrast to eastward-directed thrusting that occurred during earlier times of deformation. No known evidence indicates whether this westward-directed movement was more regional or was essentially restricted to the area of this report.

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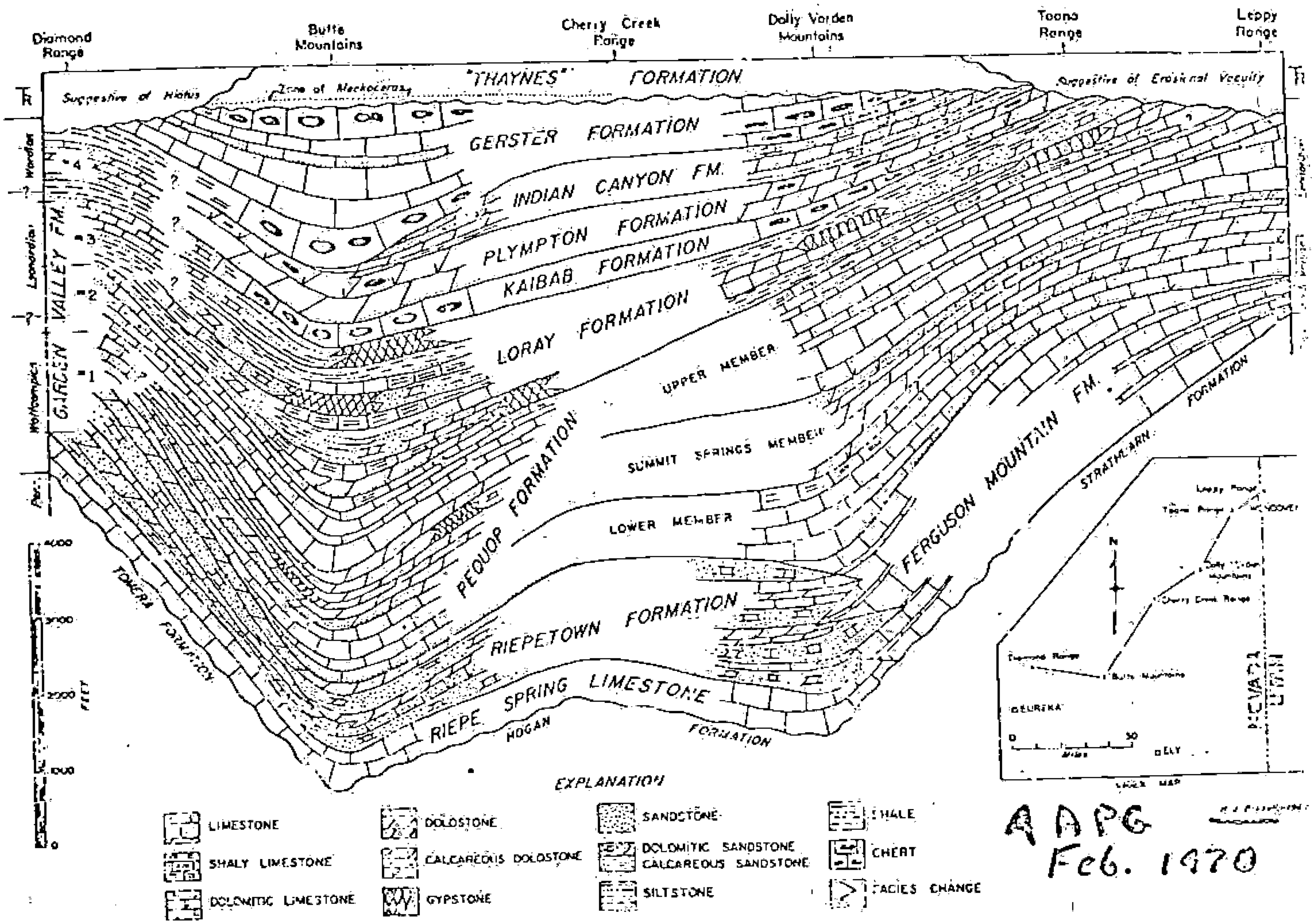

Fig. 5.—SW-NE section of Permian sedimentary rocks from Diamond Range north of Eureka, Nevada, to Leppy Range near Wendover, Utah-Nevada.

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Use of carnotite for dating young ore formations

L.K. Levskiy, A.G. Rublev, and S.V. Aprub

One of the most promising and pertinent goals in geochronology is the determination of the age of ores and the time of the ore formation in general, and of young formations in particular (Ovchinnikov, 1958). Unfortunately, at present ore processes have been dated on the basis of peri-ore altered rocks. Age determination directly from the ore minerals is uncommon and has been complicated by the absence of data on the suitability of various ore minerals for this objective.

We have found it possible to investigate carnotite, which is a uranovanadate mica, $K_2(UO_2)_2 \cdot (VO_4)_2 \cdot 3H_2O$. This mineral is interesting in that its comparatively large amount of potassium (~ 6%) has enabled us to apply the potassium-argon method, which is unusual for a uranium mineral.

It should be noted that the use of uranium minerals in geochronology is of particular interest, since it provides the possibility of determining the age by several independent methods; moreover, the uranium minerals are often involved in the composition of ore paragenetic mineral associations.

Samples were selected from one of the deposits about which there are differing views as to its formation. Therefore, the geochronological data may assist in clarifying the origin, and consequently, the scale and prospects of the mineralization. The deposit consists of several ore bodies and the samples were selected from different bodies. Furthermore, it is important to note that the country rocks surrounding the ores were not subjected, after the formation of carnotite, to temperature effects exceeding 45-50° C, since the carnotite is often found in association with gypsum, and the latter passes into anhydrite at high temperatures.

The age of the deposits has been determined ambiguously by various investigators; in most cases it has been associated with Neogene-Quaternary time.

EXPERIMENTAL METHOD

Age of the carnotites was determined by the K-Ar method. For this purpose samples weighing 0.5-0.7 g were placed in a direct-heat niobium ampoule and fused with subsequent purification of the released gases over CuO and Ca. Fusion and purification were achieved in a high-vacuum, heated, metallic fitting attached to an MS-10 mass-spectrometer (AEJ) modified for work in a static regime. The sensitivity of the instrument in respect of argon is about $2 \cdot 10^{-5}$ A/torr. The background of the "test" experiment (on the basis of argon) was $(3\text{-}5) \cdot 10^{-8}$ cm^3. The precision in measuring the isotope composition is 1-3%. The determination of the amount of radiogenic argon was carried out by the isotope-dilution method. The tracer used was the mono-isotope Ar^{38}.

The amount of potassium was determined by the flame-photometer method on a 'Zeiss' (GDR) instrument. The precision of determination is 1-3%.

The data obtained are presented in Table 1.

In addition to the dating, we carried out dynamic annealing of the carnotites in order to determine the activation energy of release of Ar and He, an objective factor, which indicates the degree of suitability of the mineral for geochronological purposes. In order to study the kinetics of release of radiogenic argon and helium from the samples by the method of dynamic annealing, special experimental equipment was employed, which included a heating device and the analyzing device of a mass-spectrometer MI-1301 (Levskiy, 1963). As a heating device, we used a platinum furnace with exterior heat, which could be heated with varying speeds. For chemical purification of the active gases, we used metallic calcium at t = 500° C, activated charcoal, and a trap of liquid nitrogen. The temperature in the platinum furnaces was measured by means of a platinum-platinorhodium thermocouple, and in addition to the automatic recorder, it was determined on a potentiometer by the compensation method.

The values of activation energy were calculated according to formula [1] (Levskiy, 1964):

$$E = \frac{R\left(\ln\frac{\alpha_1}{\alpha_2} + 2\ln\frac{T_2}{T_1}\right)T_1T_2}{T_2 - T_1}, \qquad [1]$$

Translated from Primeneniye karnotitov dlya datirovaniya molodykh rudnykh obrazovaniy, AN SSSR Izvestiya, ser. geol., 1974, no. 6, p. 22-26. The authors are with the All-Union Scientific-Research Geological Institute (VSEGEI), Ministry of Geology of the USSR, Leningrad.

TABLE 1. Results of age determination of carnotites

Sample	Weight, g	K, %	$\frac{Ar^{40}_{rad}}{Ar^{40}}$	Ar^{40}_{rad}, $\frac{mm^3}{2}$	$\frac{Ar^{40}_{rad}}{K^{40}}$	T, 10^6 m.y.	T_{av}, 10^6 m.y.
M-205	0.702	6.22	0.132	$0.108 \cdot 10^{-3}$	0.0000254	0.45	0.50
	0.672		0.058	$0.072 \cdot 10^{-3}$	0.0000183	0.33	
	0.749		0.122	$0.165 \cdot 10^{-3}$	0.0000390	0.70	
M-20	0.640	6.50	0.161	$0.376 \cdot 10^{-3}$	0.0000848	1.51	1.60
	0.673		0.161	$0.415 \cdot 10^{-3}$	0.0000939	1.68	
M-214	0.667	6.39	0.435	$3.01 \cdot 10^{-3}$	0.000691	12.4	12.4
	0.321		0.531	$3.00 \cdot 10^{-3}$	0.000690	12.4	

TABLE 2. Values of diffusion parameters

Sample	He^4		Ar^{40}	
	E, kcal/mol	C, sec^{-1}	E, kcal/mol	C, sec^{-1}
M-205	17 ± 2	$1.9 \cdot 10^3$	—	—
M-20	21 ± 3	$4.6 \cdot 10^5$	—	—
M-214	17 ± 3	$8.8 \cdot 10^4$	27 ± 2	$1.4 \cdot 10^3$

where α_1 and α_2 are velocities of heating, T_1 and T_2 are temperatures (°K) of maxima on curve of dynamic annealing, and R is the universal gas constant.

The curves obtained for the release of Ar^{40} and He^4 are shown on Figures 1 and 2. The values of the diffusion parameters are given in Table 2. For the "young" samples (M-20 and M-205), we did not succeed, owing to the low content of Ar^{40}, in determining the value of the activiation energy for the release of argon. We believe that the value of E for these samples is not less than for carnotite M-214, since the activation energy in it must be the least owing to the high radiation decay of the crystal lattice. Our assumption is supported by the similar nature of the helium-gram for the carnotites and the closely similar values of E for the principal location of radiogenic helium.

DISCUSSION OF RESULTS

As is seen from Table 1, all the carnotites have given different age values. In this respect, it is necessary to establish whether in fact the carnotites from different ore bodies were formed at different times, or whether the spread of age values is a result of the effect of subsequent processes and the insufficient retention of radiogenic argon. In this connection, we have attempted to calculate the possible losses of argon from the carnotites. In order to simplify the calculations, we shall consider an isothermal approximation for a model of continuous losses. It is assumed that the heat stress may be represented in the form of a П-like impulse, that is, isothermal annealing takes place at a temperature equal to the amplitude of the whole time of existence of the mineral.

The association between the amount of accumulated radiogenic argon and the amount of potassium is given by the well-known formula

$$\frac{Ar^{40}}{K^{40}} = \frac{\lambda_e}{\lambda}\left(e^{\lambda t} - 1\right), \qquad [2]$$

where λ_e is the constant for K-capture, and λ is the dissociation constant for K^{40}.

The accumulation of Ar^{40} in accordance with formula [2] corresponds to the geological situation in which subsequent processes are lacking. During heat stress and associated loss of argon, the formula for accumulation is transformed as follows (Levskiy, 1964):

$$\frac{Ar^{40}}{K^{40}} = \frac{\lambda_e}{\lambda_{Ar} - \lambda}\left[1 - e^{(\lambda - \lambda_{Ar})t}\right]. \qquad [3]$$

Here, λ_{Ar} is the velocity of migration of Ar^{40}, and in accordance with the solution of the equation of reaction of the first order,

$$\lambda_{Ar} = Ce^{-\frac{E}{RT}}, \qquad [4]$$

where C is the frequency factor of diffusion.

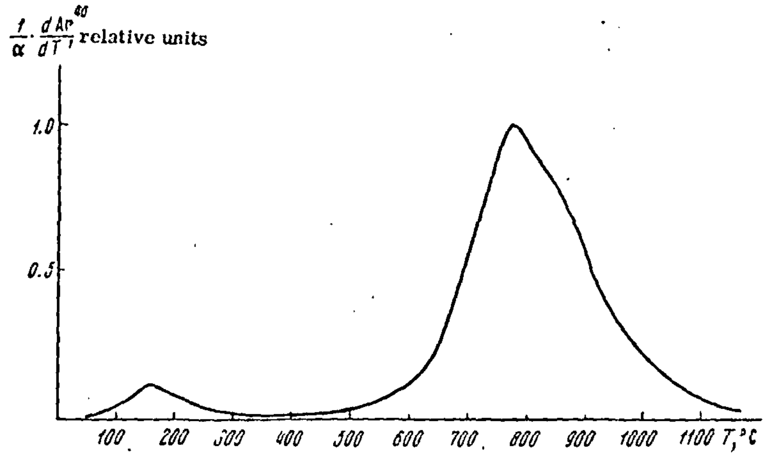

FIGURE 1. Argon-gram of carnotite (Sample M-214).

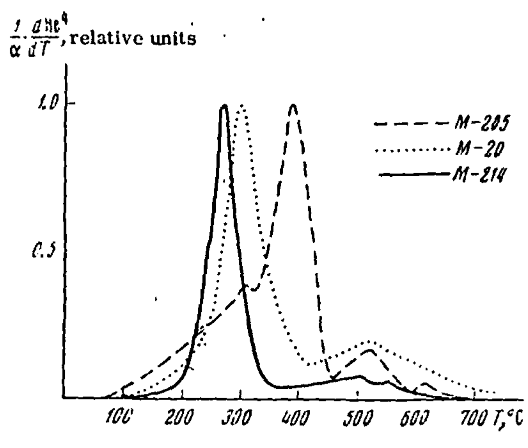

FIGURE 2. Helium-gram of carnotites (Samples M-205, M-20, and M-214).

If we know the temperature of the heat effect and the diffusion parameters, we may, from formula [2], determine the true age and calculate the argon loss.

As indicated above, the activation energy of release of radiogenic argon from carnotites is not less than 27 kcal/mol, and the maximum value of the temperature, estimated from the mineral associations, is 50° C. For Sample M-214 (apparent age 12 m.y.), the age calculated from formula [3] will be 15.2 m.y., that is, the maximum argon loss is about 21%. For the other samples, the possible losses are still less. If the temperature of the carnotites was about 27° C, the losses of radiogenic argon do not exceed 2%.

The temperature effect is not a unique process, capable alone of distorting the true age. The ready solubility of carnotite in a weakly acid medium makes its repeated recrystallization possible under the conditions of the oxidation zone. However, in the present case, this process is unlikely, since it must be assumed that conditions existed simultaneously in practically the same place, which caused solution of the carnotite, and its crystallization.

First of all, in order to assess the meaning of the figures obtained, we must concern ourselves with the method of formation of carnotite. As is known, it is readily formed as a result of conversion of other uranium minerals in the oxidation zone. In this case, carnotite must be regarded as a secondary mineral and its age will be determined by that of the formation of a superimposed zone of oxidation. However, no less likely is the formation of carnotite as a primary mineral directly from the ore-forming solutions. If this view is maintained, and it is by most investigators who work on the present deposit (Smyslova, 1972), the determinations based on carnotites will date the ore-forming process.

Consequently, variation in time of formation of the carnotites is most likely. The difference between the greatest (about 12 m.y.) and least (about 0.5 m.y.) ages of the carnotites gives a minimum length of time for the ore-forming processes for our deposit. In addition, it must be noted that the solution to the problem of primary or secondary carnotite is not involved in this report. It may only be solved by using a complex of methods (geological, geochemical, and isotope). We wished only to demonstrate the basic possibility of dating on the basis of carnotites. In this case, old carnotites, or those that have been subjected to the effect of subsequent processes, must be used with caution to solve geochronological problems, allowing for a comparatively low activation energy of release of radiogenic argon.

Figure 1 shows the argon-graph for carnotite M-214. The first peak ($T_{max} \sim 150°$ C) corresponds to the release of absorbed atmospheric argon. From comparison with the data in Table 1, it is seen that absorbed argon cannot explain those amounts (significantly large) of atmospheric argon which have been observed in the sample. The amount of atmospheric argon is $3 \cdot 10^{-6}$ cm^3, that is, commensurate with the radiogenic argon and exceeding by two orders the background in the test experiment. This means that the basic amount of atmospheric argon occurs in an identical energy position to that of the radiogenic material, since neither is resolved on the argon-graph.

This feature may be explained by the capture of atmospheric argon by the crystal lattice of the mineral during its formation from the ore-forming solutions and waters which were in contact with the atmosphere. Therefore, when dating minerals that have an exogenic origin, the danger arises of capture of large amounts of atmospheric argon, and it is impossible to reduce its amount by traditional methods (by heating or by treatment with acids).

Besides the argon-graph, we have also obtained helium-graphs of the carnotites for comparative analysis. As is seen from Figure 2, the main part of the helium is released at a low temperature with small activation energy. At the values of E indicated (table 2), more than 95% of the accumulated helium is released even at room temperature. However, high-temperature peaks are also clearly recorded on the helium-graphs, corresponding to a higher activation energy, lying within 26-44 kcal/mol. With such energy, helium is more poorly retained than argon, and this enables us to depend on the U-He method of age determination in the case of carnotites, if we can succeed in determining the amount of uranium with which this portion of the helium is associated, or to allow precisely for the loss of helium.

Attention must be drawn to the relationship between the temperature of the principal maximum on the curve $1/\alpha \, dHe^4/dT$ (fig. 2) and the age: in the youngest carnotite, the temperature is greatest, which may be explained by the different radiation decay of the minerals.

CONCLUSIONS

1. The possibility of obtaining radiometric data on carnotite by the K-Ar method has been demonstrated for the first time.

2. Diffusion parameters for the release of Ar and He from carnotites have been determined, on the basis of which we may estimate the effect of the loss of radiogenic gases on the age of the minerals.

3. The activation energy of release of radiogenic argon from carnotite is relatively low, so that accurate data may be obtained when the carnotite was not subjected after the time of its formation to subsequent processes.

4. It has been established that the main portion of the "atmospheric" argon occurs in the crystal lattice of the mineral, owing to its having been formed under exogenic conditions.

The authors wish to thank O. I. Avdeyev for his assistance in the work.

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INTRODUCTIO

The object of the detailed inv was an unusual material collecte of the Southern Basin of the Paci with features of intense manifest canic processes and subsequent l alterations. These processes to degree transformed the ferromai ules and the pelagic sediments, development of a new type of oce mineralization.

The material was collected on the 48th cruise of the scientif vessel "Vityaz'" (1970) in secto in the southwestern portion of th coordinates: lat 22°41' 09" S.; W.). In addition, we also exam collected in 1968 at the time of of the "Vityaz'" from Station 59 the northern part of the sector i (fig. 1).

In the general geologic plan longs to the southern marginal major island ridge of the Cook islands on the Australian subma tectonic respects, it is located sion of one of the long sublatitu zones, known as the Ferdinande (Pushcharovskiy, 1972).

The features of manifestati mal activity in this area have b Skornyakova et al. (1973). The have been investigated by Gorbu

Translated from Sostav i stroyeniy nykh zhelezomargantsevykh konkre novoobrazovaniy gidrookislov marg ayushchikh ikh pelagicheskikh osadk ine dna Tikhogo okeana, AN SSSR geol., 1975, no. 1, p. 91-111. Co Andrushchenko are B.P. Gradusov Shak, R.S. Yanshina, and S.Ye. I are with the Institute of Geology of Petrography, Mineralogy, and Geo USSR Academy of Sciences (IGEM) uchayev Soil Institute; and the Ge of the USSR Academy of Sciences.

EARTH SCIENCE LABORATORY

UNIVERSITY OF UTAH RESEARCH INSTITUTE
Applied Technology Division

SALT LAKE CITY, UTAH

UNITED STATES DEPARTMENT OF THE INTERIOR

GEOLOGICAL SURVEY

Application of Global Tectonic Theory to Metallogenic Studies

by

Philip W. Guild

U.S. Geological Survey
Reston, Virginia 22092

1974

Text of paper presented at the Symposium on Ore Deposits of the Tethys Region in the Context of Global Tectonics organized jointly by the Commission on the Tectonics of Ore Deposits (CTOD) of the International Association on the Genesis of Ore Deposits (IAGOD) and the Subcommission for the Metallogenic Map of the World of the Commission for the Geological Map of the World (CGMW) at Varna, Bulgaria, September, 1974.

Application of Global Tectonic Theory to Metallogenic Studies

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Philip W. Guild

U.S. Geological Survey
Reston, Virginia 22092

The Plate Tectonic, or, as I prefer to call it, the Global Tectonic Theory, has had an enormous impact on geology in the space of only a few short years. It is unnecessary here to go into the details; I am sure you are all familar with the way in which data from seismic, magnetic, oceanographic, tectonic, petrologic, paleontologic, and virtually every other branch of Earth Science have been brought to bear on developing a unified explanation for many of the features of the globe that were formerly considered as independent phenomena. This is not to say that the theory is accepted by all or that all its ramifications have been worked out. Still, the weight of evidence seems to support strongly the concept of large lateral movement of lithospheric plates in post-Paleozoic time, and it seems increasingly likely that plates and at least some plate movement were present in Paleozoic and Proterozoic time as well.

Metallogenic studies obviously must take account of this theory, and, as we know, many efforts have already been made to fit the distribution of ore deposits into the framework of global tectonics. I think it is fair to say that these have been moderately successful in a general way but that a great deal remains to be done before we can pretend that the genesis of ore deposits and their relationship in time and space to other geologic events have been explained satisfactorily. Surely we still lack the predictive capabilities so necessary to guide exploration for the concealed deposits that must furnish the mineral requirements of the future.

There are two major problems: the sources of the elements, and the processes that have concentrated them (Krauskopf, 1967). To what extent does the new theory help us in resolving these problems or in at least restricting the possibilities? And, parenthetically, do the data of metallogeny developed over many years by numerous workers tend to support the theory or raise obstacles to it?

Source
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It is well at this point to recall that de Launay (1906), who coined the term "metallogeny" and published the first metallogenic map nearly three-quarters of a century ago, was well aware of these problems and that he built on the work of his predecessors. Comprehensive schemes have been proposed to relate ore deposits to older theories of tectonic development of the earth. Perhaps the most notable was that of Bilibin, (1955) who quite successfully categorized many mineral deposits in terms of geosynclinal development. Shcheglov (1968) has extended this to include the metallogeny of the median massifs and especially of the regions of "autonomous activization" of old platforms. Billingsley and Locke, (1935, 1941) drew attention to the "crossroads" position of many major districts of the western United States, and others, among them Mayo (1958), Wisser (1959) and Jerome and Cook (1967), noted that intersections of lineaments or of lineaments with geosynclinal axes or other features are favorable loci for mineralization. Kutina (1969, 1972) has been particularly active in advocating the existence of continent and world-wide lineament systems and their influence on mineralization. At first glance some of these theories seem to be in conflict, but to the extent that the facts used to support them are correct, any general theory will have to take all of them into account.

I propose now to look quickly at some aspects of the global distribution of ore deposits in plate-tectonic terms as an introduction to the theme of this symposium--Ore deposits of the Tethys region in the context of global tectonics.

I begin with a simplified map (fig. 1) showing the principal plates of the present deduced chiefly from seismic evidence augmented by the paleomagnetic record in the ocean floor. The plates grow by addition of mantle-derived material at the accreting margins, move laterally as indicated by the arrows, and are consumed by subduction of one plate beneath the other at zones marked by intense seismic activity on the dipping Benioff planes. The picture is relatively simple on the left, where the East and West Pacific plates move away from the East Pacific Rise, and in the Atlantic, but it is exceedingly complex in the area between the Eurasian and African plates of particular interest here.

Several features of importance for our subject should be emphasized. Confrontation of an ocean plate and a continent, as in South America, gives rise to a Cordilleran margin; of two ocean plates, to an island arc. Note that most plates comprise both continent and ocean; trailing continental margins, said to be of Atlantic type, are passive or anorogenic. Transform margins are those, such as much of the west coast of North America at present, where two plates slide past one another without accretion or consumption. Small ocean basins, such as those behind or between the arcs in the northwestern Pacific, are additional features that are still imperfectly understood; some, at least, are expanding through creation of new ocean floor.

The schematic cross section from South America to Asia (fig. 2), shows the East Pacific Rise, lithospheric plates moving away with a thin veneer of basalts and sediments over gabbro and ultramafic rocks, subduction and generation of magma under the Andes and Japan Arc, and the Sea of Japan with a fragment of continental crust in Yamoto Bank. The magmas are largely calc-alkaline; they seem to be generated at depths of roughly 150 km.

The driving force for plate motion is commonly believed to be some sort of convection in the asthenosphere. Upward transfer of energy is probably concentrated in "plumes" that cause hot spots (fig. 3); initial fracturing of continents has been attributed to such plumes rising under sial to form domes that commonly fracture at angles of approximately 120°. Magmatism of alkalic or tholeiitic nature may take place, followed by separation of the fragments along suitably oriented fractures to initiate formation of new ocean crust. One or more of the "arms" ordinarily does not separate or fails after a short time, and many entire uplifts abort without rifting. If the plumes come up under either oceanic or continental crust in motion over the mantle, lines of volcanoes oriented in the direction of plate motion may be formed.

How does this apply to metallogeny? Let us look at the proposition that certain types of ore deposits form in specific positions that can be defined in plate-tectonic terms and evaluate it by what we know of their distribution. I will emphasize the simpler areas, principally of the Western Hemisphere, and discuss only a few ore types.

Deposits may form at or near plate margins (fig. 4), or in plate interiors. The outstanding example of deposits formed at an accreting margin are the Red Sea muds; the volcanogenic massive sulfides on Cyprus and in some other areas are frequently cited as other examples. Podiform chromite deposits of alpine type are probably carried up from the mantle as part of the infusible residue.

Consuming margins are loci of much or all of the calc-alkaline magmatism with which many endogenic "postmagmatic" or "Cordilleran" types of ore are associated. Examples are the porphyry copper-molybdenum deposits. Kuroko and Besshi types of massive sulfides, magnetite-chalcopyrite skarns, some precious and base-metal ores, tin, tungsten, mercury, and antimony.

A plot of the post-Eocene endogenic metallogenic provinces of the world (fig. 5) on the base previously shown confirms that there is indeed a close spatial relationship to the consuming margins defined from geophysical data. The provinces parallel the subduction zones and lie over the downgoing slabs. The picture in South America seems particularly clear, but the deposits of the western Pacific also fit the pattern. I will leave discusssion of the Tethyan region to the speakers to follow, except to point out that at this very small scale the same generalization seems to hold. The pattern in North America north of Mexico is unusual for the great width, some 1500 kilometers, of the province, and there are undoubtedly complicating factors to which I will return later. Note that the northern half of the continent is essentially devoid of post-Eocene mineralization, apparently because the margin has been transform for much of the later Teriary.

I turn now from consideration of several types of endogenic mineralization in a single time frame to a rapid review of the gross distribution of a single class of ores--the massive sulfides--in various geologic periods of the past. It could be termed a kind of time-slicing. Lumping deposits that can have formed at different plate positions, accreting and consuming, may at first glance seem illogical, but ~~convergent plate movement will eventually bring products of mid-ocean rises to consuming margins where they may be either subducted or incorporated into orogenic belts by obduction.~~

Starting with the Tertiary (fig. 6), we have the Kuroko and other deposits of the western Pacific and a few deposits in Tethys. In the Mesozoic (fig.7), the Pacific belt includes deposits in North America, the Caribbean, and northern South America as well as more numerous deposits in Tethys. The pattern in Hercynian time (fig. 8) is much the same, except that the Uralian and Altai belts may mark locations of former plate convergence in the middle to late Paleozoic. The Caledonian picture (fig. 9) is entirely different; nearly all the deposits are restricted to northwestern Europe and eastern North America; they are interpreted as marking the positions of a proto-Atlantic ocean which opened and then closed in late Precambrian and early Paleozoic time. The offset from Europe to North America is of course due to the later breakup of Laurasia that formed the present Atlantic Ocean. Only short belts and isolated deposits or districts are present in older Precambrian rocks (fig. 10).

If we grant that a spatial relationship to convergent margins has been demonstrated for certain types of deposits, and if the general model of magma generation through subduction of oceanic crust is accepted, pre-enrichment of the metals in this crust could explain the gross distribution of some metallogenic provinces. In other words, their metals may have had an initial low concentration in the ocean that has subsequently been enhanced by plate-tectonic phenomena. Sillitoe (1972) has proposed that the porphyry copper deposits derive their metals from this source, pointing out that basic igneous rocks and pelagic clays, manganese nodules, and especially the sediments on the crests and flanks of the East Pacific Rise all contain above-average quantities of copper. Corliss (1971) and Dymond and others (1973) take this one step farther, postulating that sea water circulates to depths of several kilometers in the newely formed basalts, setting up hydrothermal systems that leach them of their metals and form both the low concentrations and perhaps also the massive sulfide deposits (J. B. Corliss, oral commun., 1974). Lateral inhomogeneities, differences in spreading rates, subduction history (as, for example, whether the upper metal-rich layer is scraped off at the trench or carried down), and a myriad geologic factors of magma generation, ascent, differentiation, locus of intrusion, wall-rock geochemistry, and subsequent erosional history all can affect the actual presence or absence of economic ore deposits.

The theory is attractive, and very considerable work is being carried out to test and refine it through deep-sea drilling, major- and minor-element chemistry, isotope research, and so forth. Unfortunately, considerable time must have elapsed since even the youngest deposits known on land were initially generated at an ocean ridge, hence, direct proof is still lacking.

Let me now turn to consideration of deposits not related to subduction that I believe are nevertheless explicable in terms of overall global-tectonic theory. The evidence is admittedly more tenuous and my thesis is even less susceptible to proof than the foregoing. In North America, strata-bound disseminated lead-zinc deposits, with or without copper, barite, fluorite, and, in places, minor nickel and cobalt occur in carbonate platform-cover rocks of Paleozoic age lying on the eroded Canadian Shield (fig. 11). Intrusive igneous rocks are notably absent from the vicinity of the deposits, which obviously are not of the Cordilleran type. Much of the lead is of anomalous isotopic composition, implying two-stage or multistage derivation from crustal sources. Fluid inclusions are generally highly saline; filling temperatures range from less than 100° to about 200° C. In general, the major districts are on or near basement domes.

The genesis of these deposits is widely disputed. Explanations have ranged from syngenetic to "telethermal"; variations on some form of ground-water and (or) connate-water migration and reprecipitation in favorable sedimentary or structural locations seem to fit the facts best. For a few relatively smaller deposits of this general type the evidence for direct leaching of an identified terrane by ground water and syngenetic precipitation in a nearby sedimentary environment is quite convincing.

However, gross distribution of the major districts of the United States suggests deep tectonic control that has localized not only the mineralization but also diverse igneous and structural events. Heyl (1968, 1972) has documented the evidence for the 38th-parallel lineament that extends from the Appalachians to Missouri and perhaps as far as the Rockies - a total of some 2200 km. In addition to eight mineralized districts, 10 alkalic or alkalic-mafic intrusive centers (some of them kimberlites) and eight crypto-explosion structures (Snyder, 1970) occur along this zone. Less well defined lineaments are at 36° and at 42° N. In Canada, the Pine Point district directly overlies the McDonald fault that divides two major provinces of the Shield and has been traced in subsurface many hundreds of kilometers to the southwest across the Phanerozoic Interior Plains.

To digress for a moment, let us consider the admittedly unimportant deposits of the Benue trough in Africa, which occur in Cretaceous sediments and volcanic rocks that filled a rift or aulacogen that opened up for a short time in the Cretaceous when South America and Africa began to separate. They consist principally of lead and zinc with minor copper, barite, and fluorite; most are veins, but some are apparently stratiform in limestone. In any case they seem to have affiliations with the Mississippi Valley deposits under discussion. A plume rising under a triple junction presumably initiated the fracturing and rifting; could it not also have provided the energy to mobilize metals and form the ore deposits?

Returning now to North America, several lines of evidence suggest to me that mantle plumes may have been responsible for the genesis of the Mississippi Valley deposits. Inclusions in the kimberlites prove that the fractures penetrated the sial (Brookins, 1969). Plumes in continental settings are believed to cause emplacement of nepheline syenites, carbonatites and kimberlites, and Macintyre (1973) has proposed that these plumes are periodic and may have reactivated major lineaments and rifts. This seems important, for the region along the 38th parallel has undergone intermittent igneous activity from at least the Cambrian to the Tertiary, and seismic events are continuing at present. Hoffman (1973) has demonstrated that the McDonald fault was the site of an aulacogen in early Proterozoic time; its post-Precambrian movement may also have been intermittent, for the sedimentary history has been quite distinct on either side. If plumes are related to this mineralization, they could provide the relatively low but persistent heat indicated by fluid-inclusion studies (Roedder, 1971). I suggest that most, perhaps all, of the metals were mobioized from sedimentary or upper crustal sources by the energy derived from plumes, but it seems likely that the fluorine may have been a direct contribution from depth. Perhaps the isolated copper and (or) nickel-cobalt concentrations also had localized sources in the lower crust or mantle.

Some similar process may have been responsible for the major lead-zinc deposits of Northwest Africa. Kanes and others (1973), have drawn attention to large fractures paralleling the Atlantic and Mediterranean coasts that are attributed to stresses accompanying the separation of North America from Africa and to movements between Europe and Africa in mid-Mesozoic time. Carbonatites are present. The principal metallogenic epoch was in the middle Mesozoic, although others apparently preceded and followed it. Russell (1973) has speculated that the Irish base-metal deposits, which are aligned on north-south zones, formed on geofractures which focused heat rising from degassing of the mantle. The metals were probably leached from the sediments by convective circulation in crustal waters and deposited where sufficient sulfur was encountered. He relates the postulated geofractures to stressed accompanying formation of the Rockall trough.

If the Mississippi Valley deposits are perhaps related to hot spots, and hence to the plumes that may drive the plates, it seems evident that the diamond deposits of the kimberlites and the carbonatite-related niobium, vanadium, phosphate, and rare-earth deposits are at least equally dependent on the global-tectonic regimen for their genesis. Additionally, the linear mafic and ultramafic intrusions such as the Great Dyke and Muskox intrusions must, by their very size, be derived from the mantle and related to continental fracturing that did not proceed to the rifting stage. The chromite, iron-titanium-vanadium, and platinum-nickel-copper deposits of the Bushveld Complex are obviously of direct mantle origin; is it possible that the substantial quantities of tin and fluorite, and minor amounts of lead, zinc, tungsten, molybdenum, and rare earths associated with the acid differentiates and Bushveld granite (Willemse, 1969) at the top of the complex are also derived from depth, or were they concentrated from epicrustal materials?

Let me turn back to North America, and especially to the anomalously wide zone of mineralization in the western United States mentioned earlier. On this sketch map (fig. 12) we have the Canadian Shield, the Appalachian orogen, the Interior Plains, and the Cordillera. I will emphasize the last. Very simply, the Cordillera is a complex, long-lived feature that began in the Precambrian and has continued to the present. The greatest concentration of magmatism and orogeny took place in the Mesozoic and early Tertiary, but various episodes both preceded this and have continued to the present.

On this slide (fig. 13) we have merely the outlines of these major geologic provinces with a few selected features that may bear on my subject. The circles represent districts of the Mississippi Valley type already discussed; many lie along the 38th parallel, and in a gross way all are peripheral to the exposed margin of the Shield. To skip to the western margin of the continent, the numerous porphyry copper deposits in Canada, shown by the squares, lie in a fairly narrow belt near the coast. To the south, in the United States and Mexico, only a few porphyry deposits are known until we go quite a way eastward. Metal zonation inward from the western edge of both North and South America is a well-known feature that has been documented in great detail by Wilson and Laznicka (1971). Their copper-lead line is indicated on the sketch. In classic terminology, it approximately follows the eugeosynclinal-miogeosynclinal boundary of Mesozoic time. On the west, the ratio of copper to lead ranges from 20 to as much as 600 times, whereas east of the line, lead exceeds copper by a factor of 2 to 100 or more. As Wilson and Laznicka point out, however, three isolated major districts, Butte, Bingham, and Ely, lie within the miogeosynclinal domain. The numerous large porphyry deposits in the Sonoran province are also in a miogeosynclinal or even a shelf environment; their magnitude causes a sharp indentation in a statistical copper-lead line that seems to return again to its near-coast position to the south.

Two features distinguish the deposits of this interior zone from those to the west. (1) The districts are not distributed more or less parallel to the structural grain, but rather lie transverse to it along a number of poorly to fairly well defined linear patterns that are generally northeast or northwest trending. (2) Age of mineralization even for nearby districts in a belt can vary widely, as can also the nature of the ore. For example, in the Colorado Mineral Belt, best documented (Tweto and Sims, 1963), of all those indicated on the sketch, a zone of Precambrian shearing with relatively minor intrusions of upper Precambrian alkalic-mafic rocks and associated deposits of thorium, rare earths, niobium, and titaniferous magnetite was, in "Laramide" and Oligocene time, the locus of important base- and precious-metals, molybdenum, tungsten, gold-silver-tellurium, fluorite, and uranium mineralization. Copper is very subordinate. The Coeur d'Alene district of silver-lead-zinc-copper-antimony ores lies on the Osborn fault zone, locus of six periods of mineralization ranging from Precambrian to Tertiary (Fryklund, 1964). This zone forms part of the Lewis and Clark line of Billingsley and Locke (1941), which extends at least to Butte and perhaps considerably farther. Two or perhaps three linear zones can be drawn through the porphyry deposits of the Sonoran province, and many others have been proposed (e.g. Mayo, 1958).

Various persons (Lipman and others, 1971; Moores, 1970) have tried to explain the orogeny, magmatism, and presumably the mineralization across this wide belt in terms of a gently dipping Benioff zone (fig. 14A) or by multiple zones successively overridden and cut off by the westward-advancing continent (fig. 14B). Gilluly (1971), however, expressed great doubt that subduction can be responsible for magmatism to distances greater than 700 kilometers from the trench, or to approximately the average position of the copper-lead line.

An alternative explanation seems preferable to me. I suggest that we are seeing the results of intermittent reactivization of a continental segment not caused directly by interaction with an oceanic plate but rather through ascent of energy from the mantle, perhaps as a giant eddy or upwelling behind the oceanic lithosphere of a descending slab (e.g., Nelson and Temple, 1972). Figure 14C illustrates this schematically. Tidal forces tend to hold back the Earth's crust so that the athenosphere moves slightly faster; the latter tends to rise in the lee of the descending slab, slowly heating and melting the base of the crust. The late Cenozoic Great Basin with its thin crust, shallow seismicity, high heat flow, and bi-modal volcanism can be considered as a late stage of this process (Scholz and others, 1971).

Billingsley and Locke (1941) pointed out 35 years ago that the great ore districts occur in structurally complex regions typified by persistent deep-seated crustal breaks and that heat, not magmatism per se, is the least common denominator of mineralization. Shcheglov (1968) cited the Rocky Mountains as one of his examples of regions of "autonomous activization." I suggest that the word "autonomous" is unnecessary and misleading -- the process is explicable in global tectonic terms -- and that "reactivization" is preferable.

As I envision it, the slow increase in heat applied from below by this broad upwelling would seek out flaws in the crust and gradually mobilize and channel ore elements into them. Some of the elements could be derived from descending oceanic crust as Sillitoe postulates, especially near the edge of the subducted slab, but most would be derived from the upper mantle and especially from the crust itself. Lead isotope data indicate derivation of that metal from various terranes -- some leads give anomalously old model ages as if they had been isolated for a long time from radioactive uranium and thorium, perhaps deep in shield areas where these elements decrease sharply with depth. Other leads suggest, by their mixed compostitons of old and radiogenic components, derivation from two or more sources (Stacey and others, 1968; Zartman and Stacey, 1971). If the lead is derived from sial, I see no reason why other metals may not also have a similar provenance. Periodicity -- in Nevada, five distinct metallogenic epochs, superimposed almost inextricably in space, have been recognized from the Jurassic to the late Tertiary, (Roberts and others, 1971) -- could be due to times of more rapid convergence of the plates, and hence of the intensity of the athenosphere eddies, without the need to call on oceanic crust to provide the metals. The metals provinces have been shown by Noble (1974) and Lowell (1974) to be persistent and apparently unrelated to any simple plate-tectonic regimen, yet the entire geologic pattern of the Cordilleran margin of North America seems to demand a fairly consistent unified explanation. Inheritance of ancient metal-distribution patterns, combined with a global-tectonic environment conducive to creating the processes that could mobilize and concentrate disperse elements into ore deposits, may be the answer to the undoubted existence of metallogenic provinces in this region and perhaps in others as well.

In summary, I suggest that there are four major global-tectonic environments in which endogenic mineralization can occur. In attempting to identify the deposits that formed in each of these, I believe that we can gain a better understanding of the sources of the ore elements and the processes responsible for their concentration.

1) The ensialic mantle plume or hot spot, which can, but does not necessarily, lead eventually to fragmentation of continents. Some elements are introduced directly from the mantle, as in the case of those that occur in the great mafic lopoliths and perhaps those in their differentiated tops. Ores associated with kimberlites, carbonatites, sialic-mafic, and anorogenic alkaline intrusive rocks are other examples. Where sedimentary covers, and especially carbonates, overlie basement, the effect of the heating is chiefly to mobilize elements such as lead and zinc already present in the crust, although probable the fluorine of the fluorite deposits, perhaps some of the barium, and possibly some other metals are also direct contributions from depth.

2) The accreting margin or zone of plate divergence, in which mantle material rises in an ensimatic environment. It seems obvious that in the mid-ocean rises the metals must come from the mantle, inasmuch as sialic sources are absent. However, even here the process may consist of two stages with deeply circulating sea water responsible for leaching and concentrating elements dispersed in the newly formed ocean crust.

3) Zones of plate convergence, where subduction and the attendant magmatism may lead directly to creation of both volcanogenic stratabound deposits and the numerous epigenetic types of deposits found along orogenic belts. The ore metals may well be derived from the oceanic plate. Evidence for this is particularly strong in the young island arcs lacking any significant substratum of sialic basement; in the active continental margins and in older arcs, I suspect that part of the metals may be concentrated from preexisting reservoirs in the overriding plate.

4) What may be termed an ensialic back-arc region where a subducting lithospheric plate causes upwelling of the asthenosphere along the base of the crust and provides the energy to mobilize ore elements from the upper mantle and continental crust. Presumably much of the ore metal is derived from the crust, but direct contributions from the mantle are not precluded. Along the interface near the edge of the descending plate, some metals could be of ocean-plate derivation. The best known example is in western North America, which is, in many ways, unusual from a metallogenic point of view, but it seems likely that similar situations may have arisen elsewhere in the past. The nearest analogies at present are the ensimatic back-arc basins of the western Pacific, where absence of sial probably precludes much if any ore genesis, and the water cover would in any case prevent the discovery of deposits.

What of possible applications of metallogenic data to confirmation or modification of the general theory itself? One may be that a very preliminary examination of the evidence suggests to me that deposits of the first -- mantle plume -- environment tend to be distributed either parallel or at right angles to major plate margins, whereas those of the fourth (ensialic back-arc) environment are in belts oriented at roughly 45° to them. If born out by detailed examination of all regions of the world -- and this obviously means completion of the world metallogenic map -- we could provide a useful tool for extending the global tectonic picture back in time. Obviously to do so we must first refine the crude scheme I have presented and document its validity or, conversely, discard it and devise a better one. And, of course, we must sharpen our ability to distinguish ore-deposit types to avoid the pitfall of lumping genetically unrelated deposits into provinces or belts and drawing erroneous conclusions. This problem of equating apples, oranges, and even a potato or two has been the downfall of more than one elaborate scheme, and could certainly discredit mine!

I have not gone into the subject of the exogenic deposits for lack of time, but many, such as the evaporites, detrital accumulations on trailing continental margins, and weathering crusts are consequent on vertical and horizontal movements that can also be related to the global tectonic picture. Hopefully, we can look forward to the day when metallogeny can explain the reason for the existence and localization of all ore deposits. Some progress has been made, but much remains to be done!

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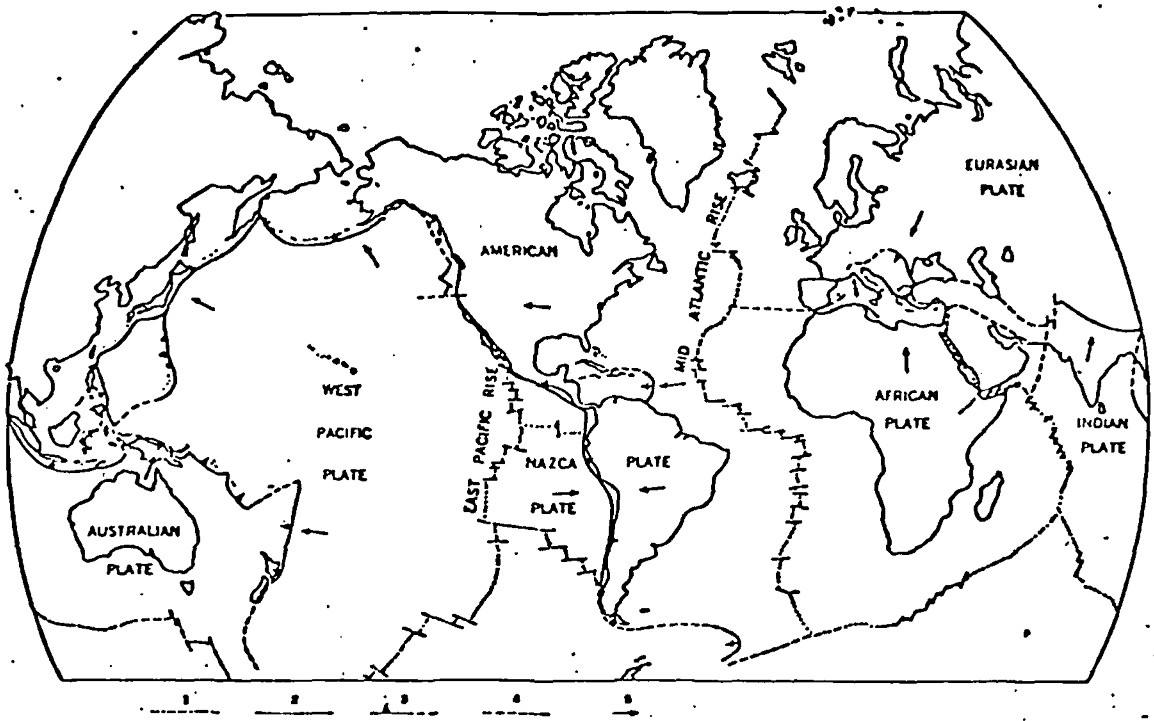


Fig. 1. Major lithospheric plates. 1) Accreting plate margin; 2) transform plate margin; 3) consuming plate margin with dip direction of downgoing plate; 4) margin of uncertain nature and (or) location; 5) relative plate motion.

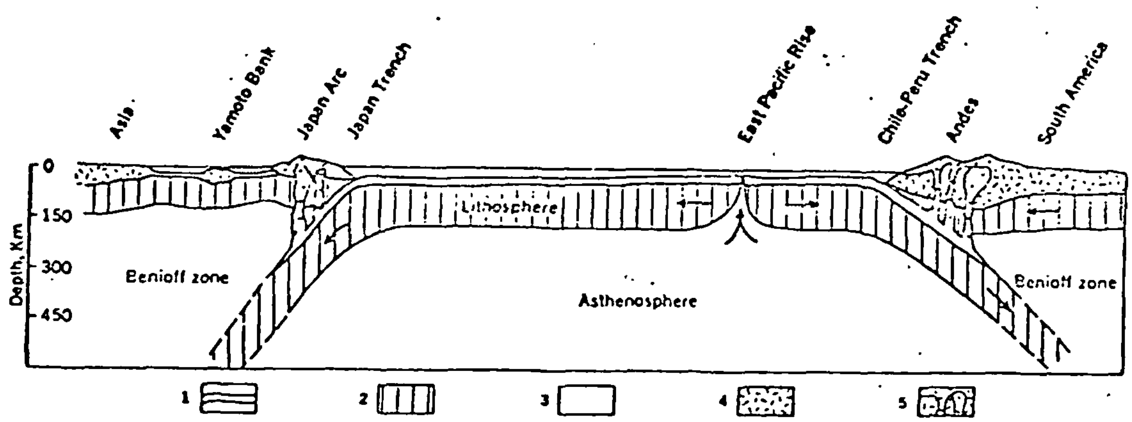
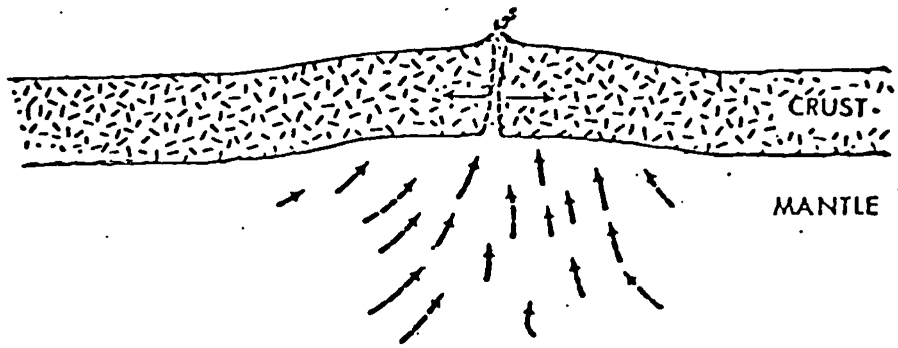


Fig. 2. Schematic section from South America to Asia illustrating major features of the plate tectonic theory. Not to scale. After Dewey and Bird (1970). 1) Oceanic crust; 2) upper mantle; 3) lower mantle; 4) continental crust; 5) calc-alkaline magmas, intrusive (+) and extrusive (v) products.



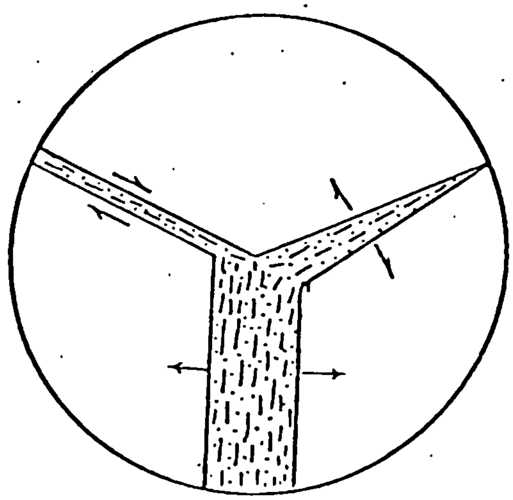


Fig. 3. Schematic section and plan of ensimatic hot spot. A. Mantle plume domes the crust, causes tension, and may produce alkaline or alkalic-mafic volcanism. B. Tangential tension commonly results in three fractures at about 120°; depending on orientation, these may react in different ways. Here the north-south fracture has opened; tholeiitic magma has initiated accretion of oceanic crust and separation of plates. The upper left fracture is essentially transform. The upper right opens for a while by spreading but eventually fails and is preserved as an aulacogen. Modified from Burke and Dewey (1973).

(Fig. 4) DEPOSITS FORMED AT OR NEAR PLATE MARGINS

Accreting (diverging):

Red Sea muds

Volcanogenic massive sulfides of
Cyprus type

Podiform Chromite

Consuming (converging):

"Cordilleran" types of Sawkins (1972):

Porphyry copper

Massive sulfide ores of
Kuroko and Besshi types

Fe (Cu) skarn ores

Au-Ag

Ag-Pb-Zn

W, Sn, Hg, Sb

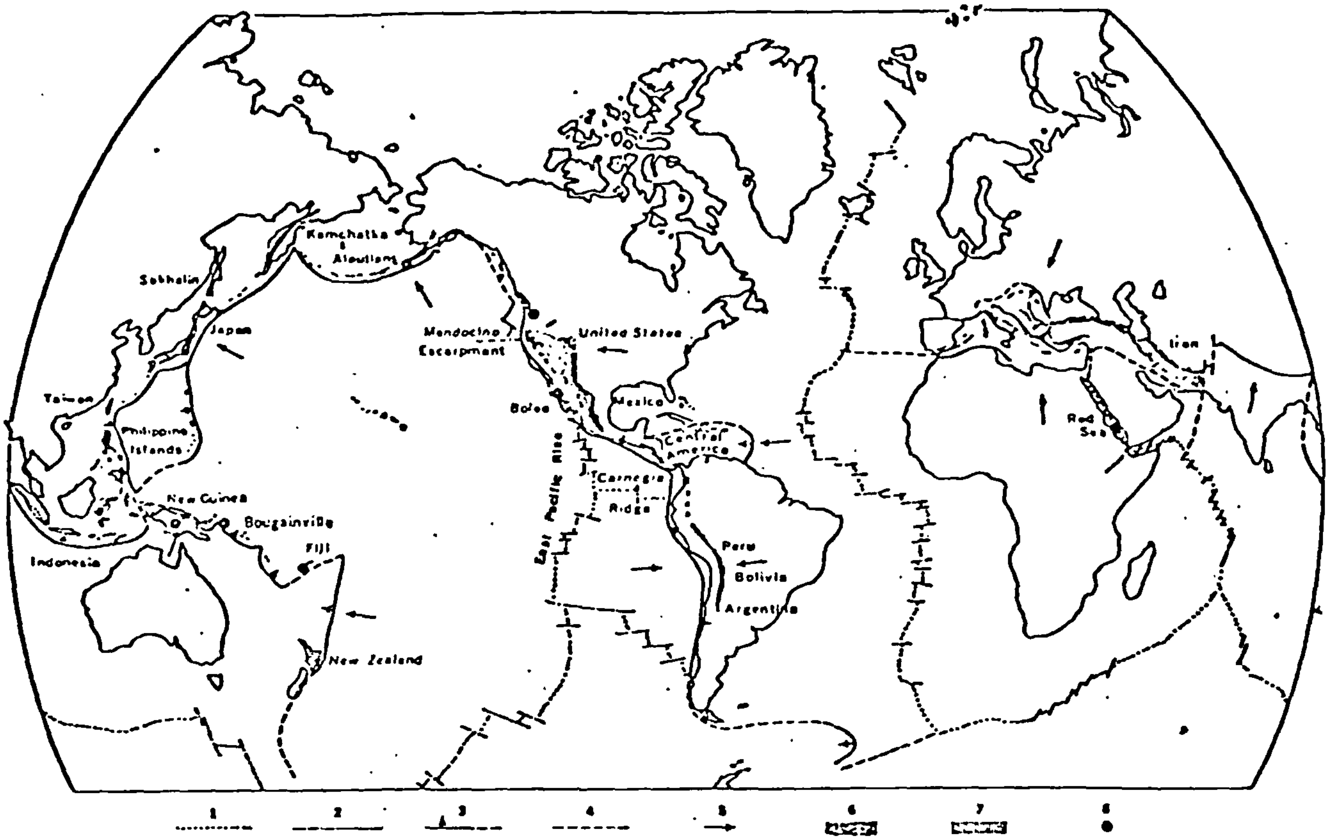


FIGURE 5. — Principal post-Eocene endogenic metallogenic provinces of the world in relation to the major lithospheric plates. Key to ornament: (1) accreting plate margin; (2) transform plate margin; (3) consuming plate margin with dip direction of downgoing plate; (4) margin of uncertain nature and (or) location; (5) relative plate motion; (6) area of mineralization of post-Eocene age; (7) minor or suspected post-Eocene mineralization; (8) major or noteworthy isolated ore deposit of post-Eocene age.

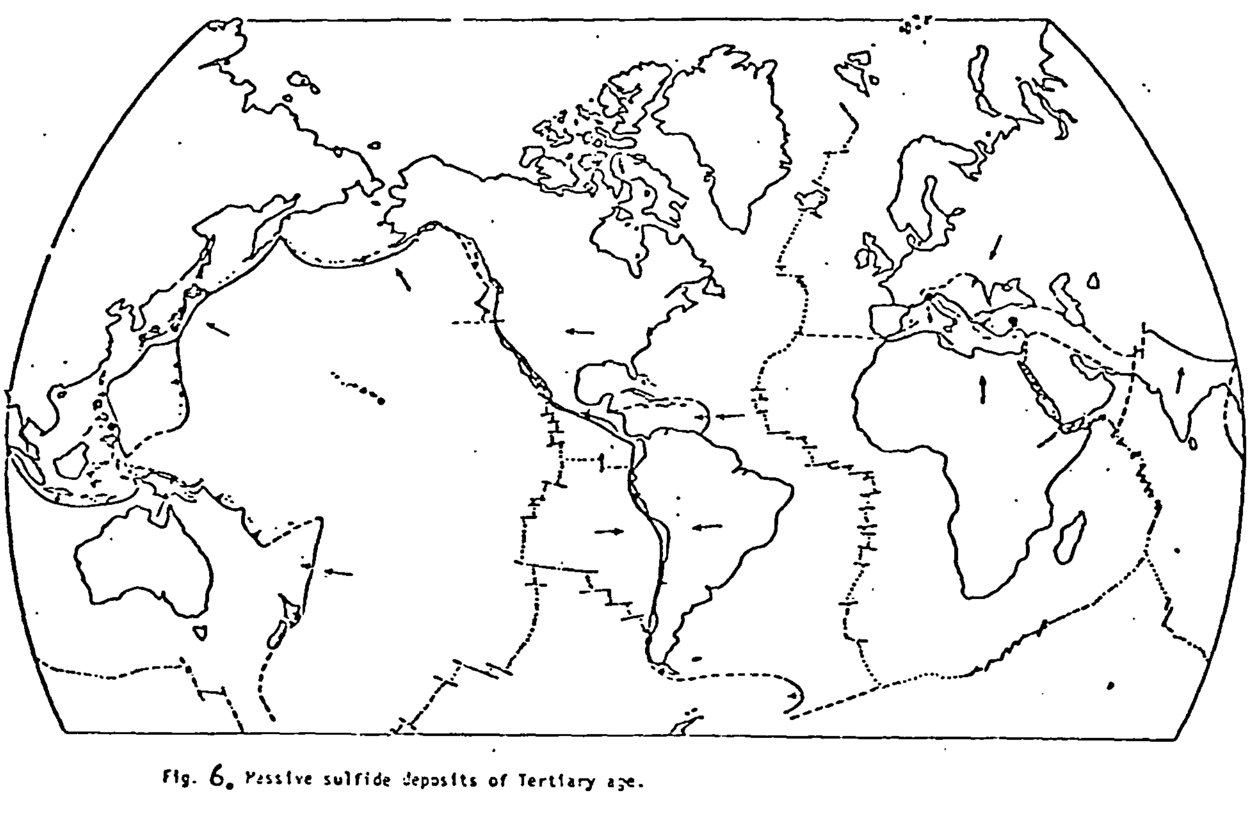


Fig. 6. Massive sulfide deposits of Tertiary age.

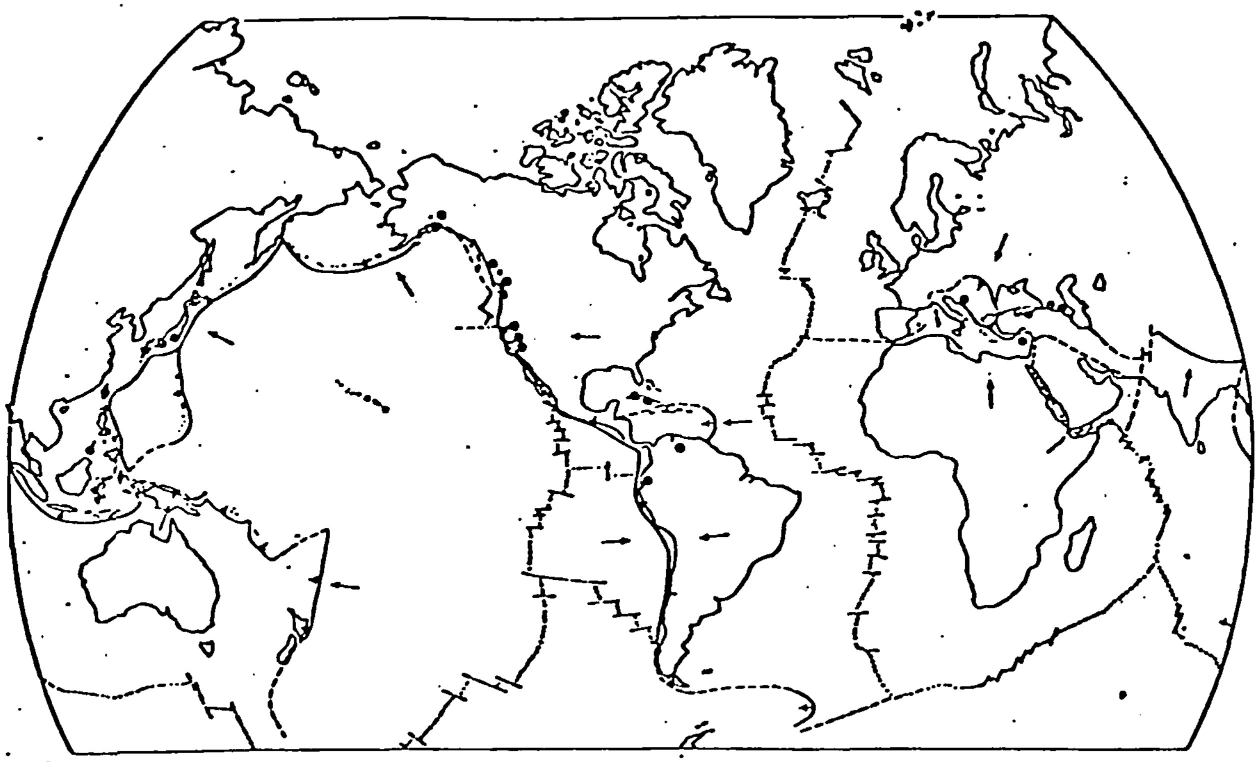


Fig. 7. Massive sulfide deposits of Mesozoic age.

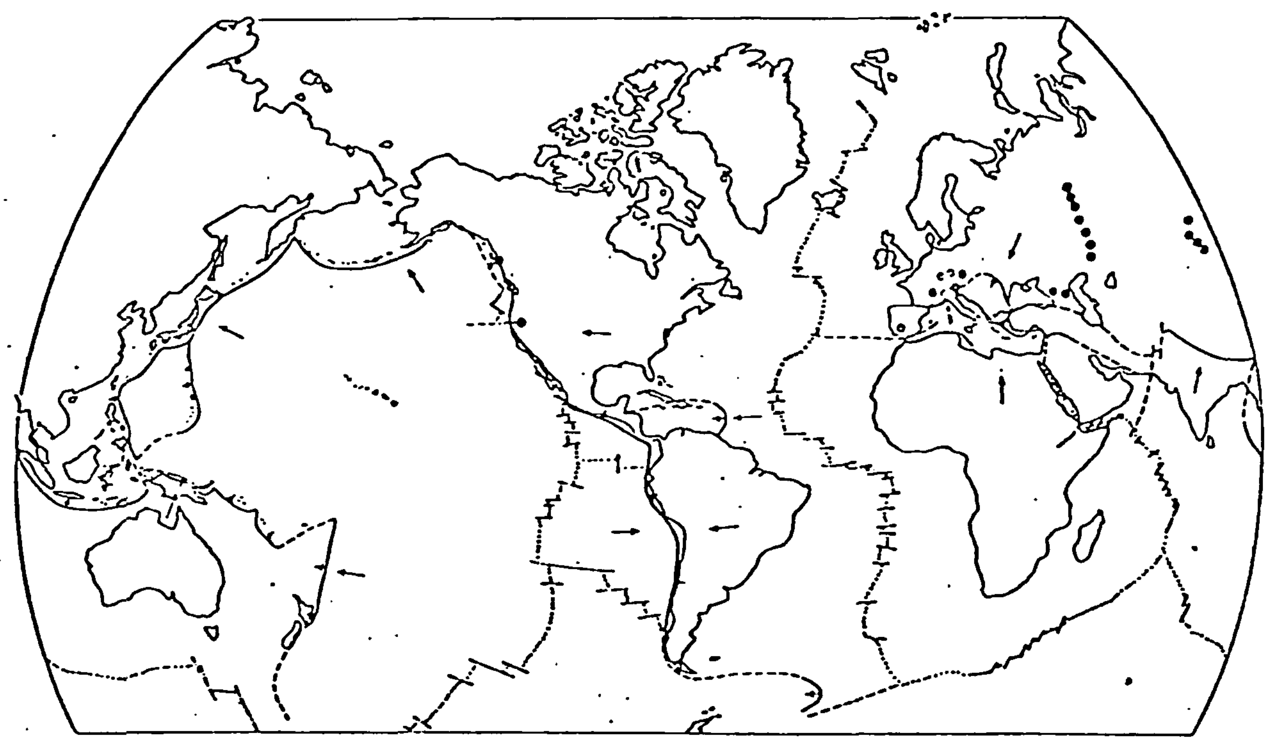


Fig. 8. Massive sulfide deposits of Late Paleozoic (Variscan or Hercynian) age.

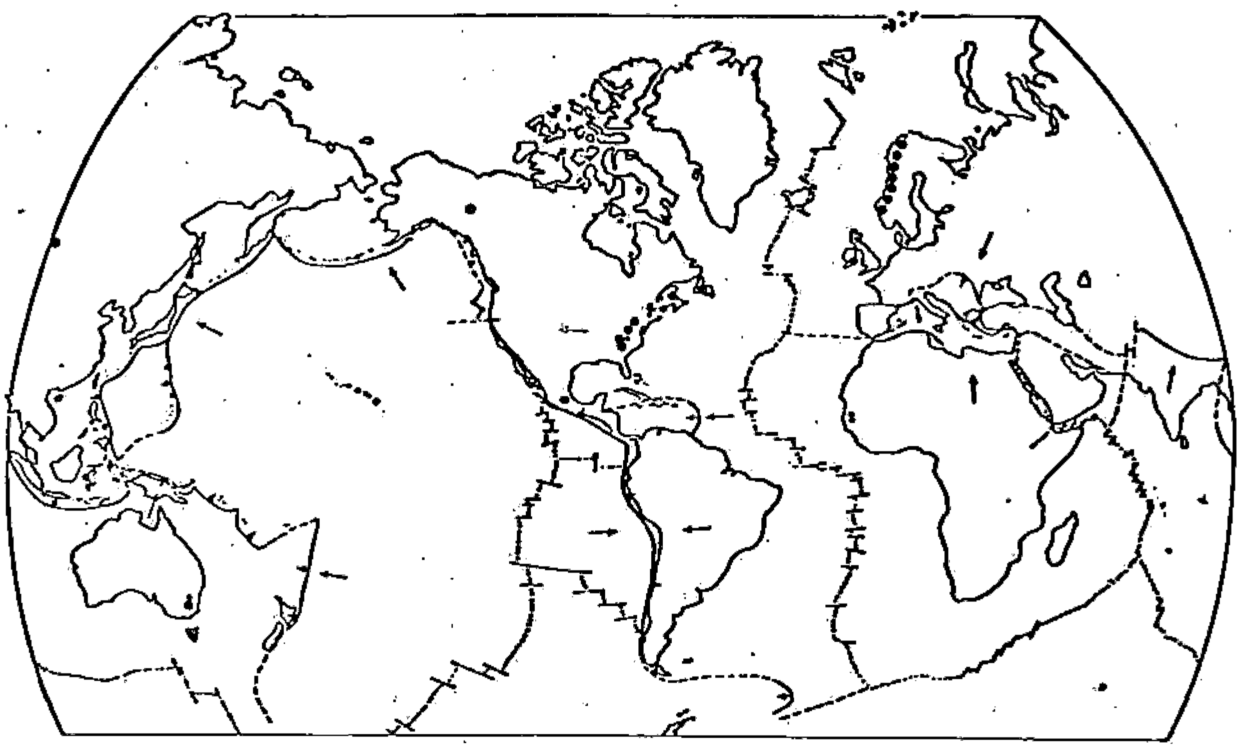


Fig. 9. Massive sulfide deposits of Late Precambrian and Early Paleozoic (Caledonian) age.

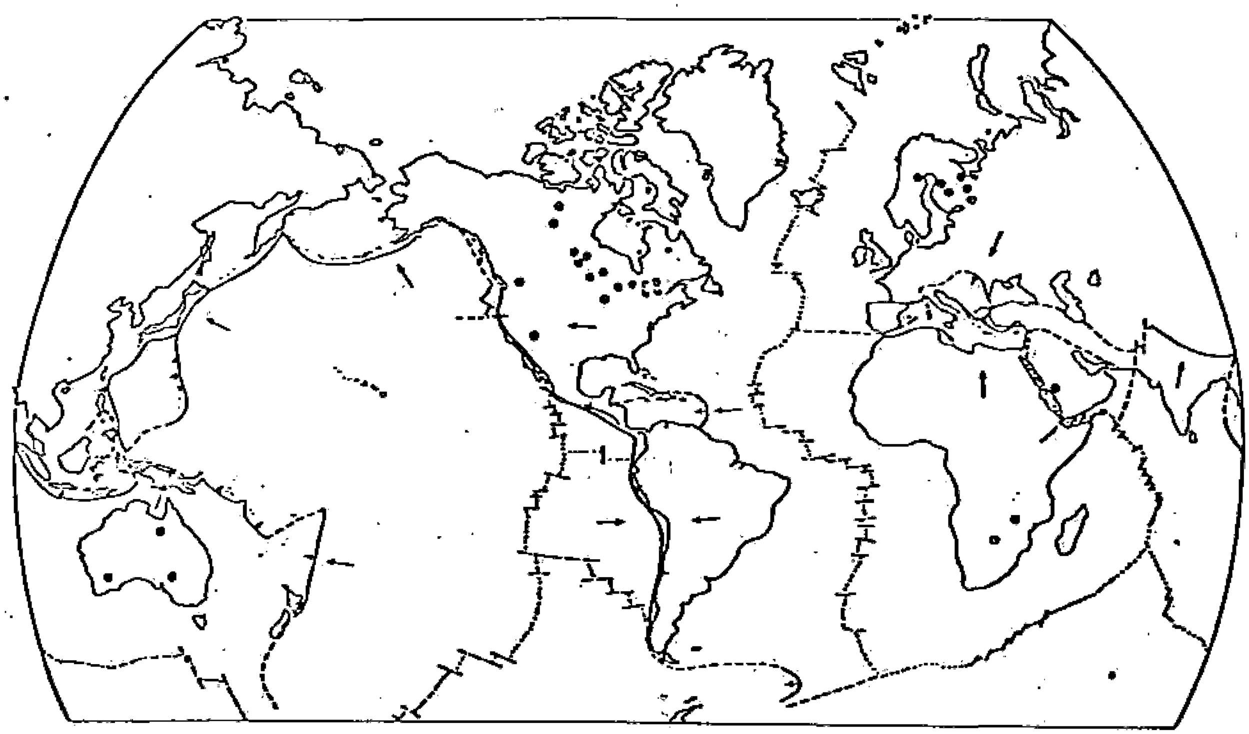


Fig.10. Massive sulfide deposits of Early Precambrian age.

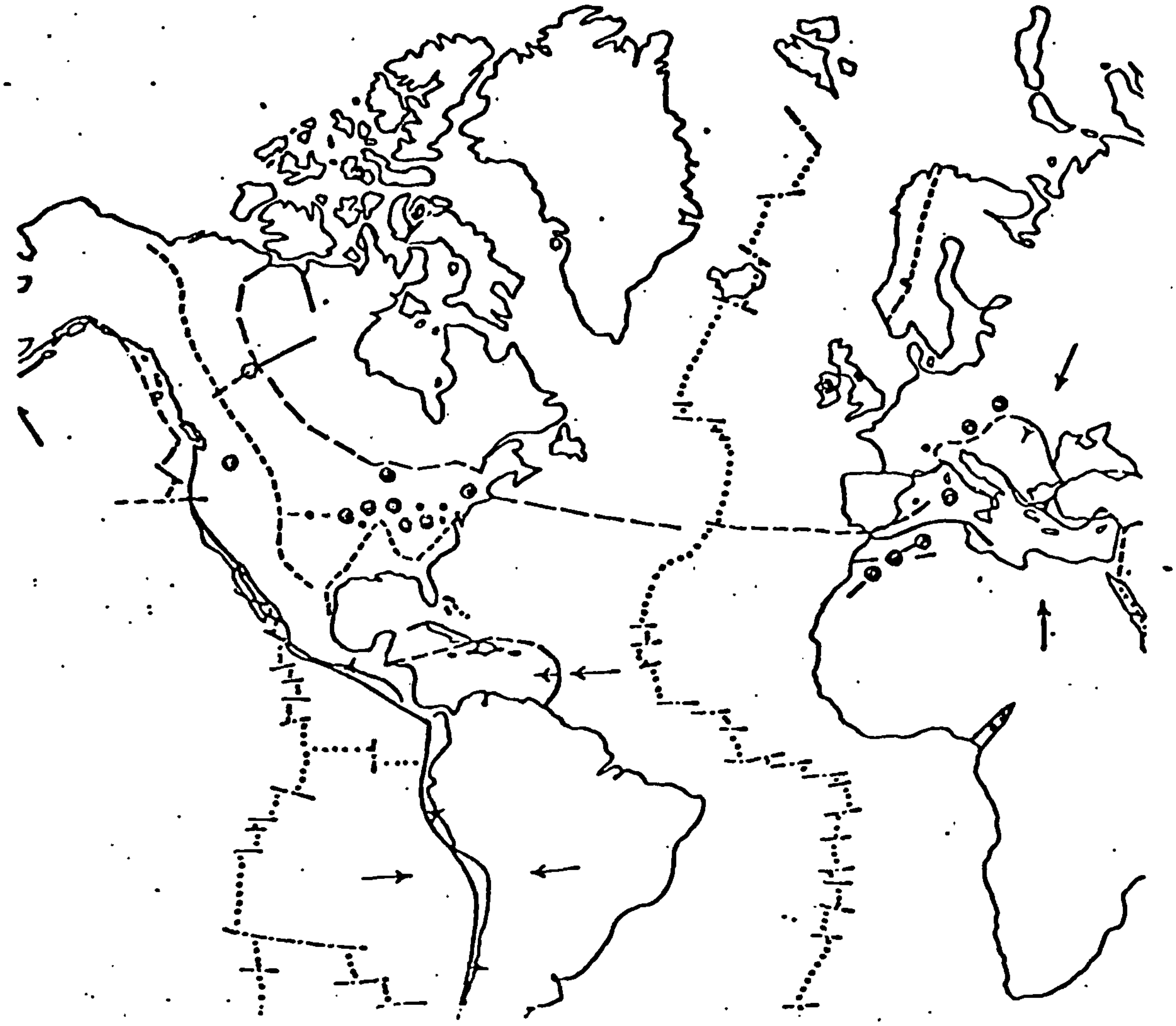


Fig . 11. "Mississippi-Valley" and related(?) districts of the world.

Fig. 12. Colored slide only - no copy furnished.

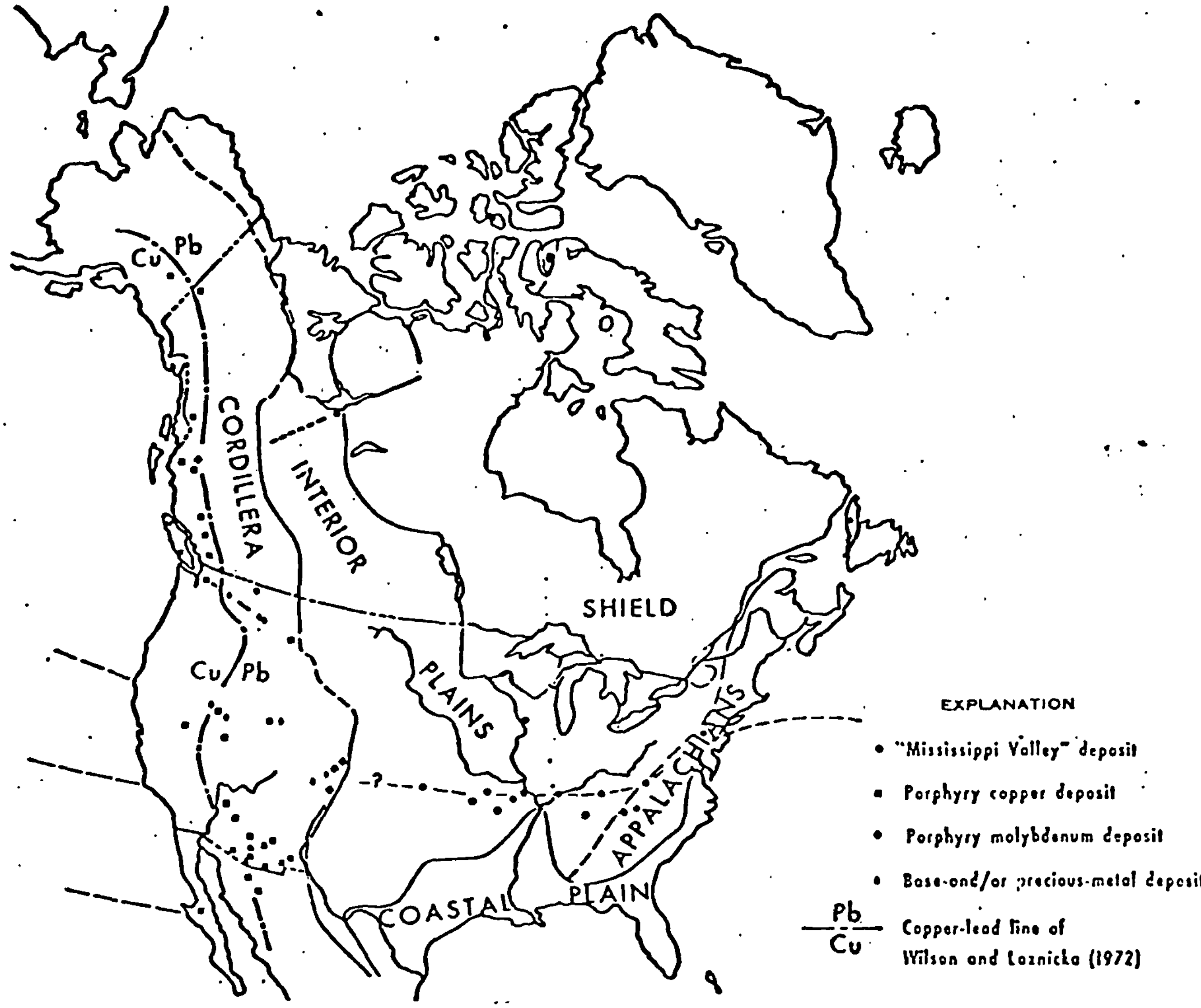


Fig. 13. Some ore deposits of North America.

Guild

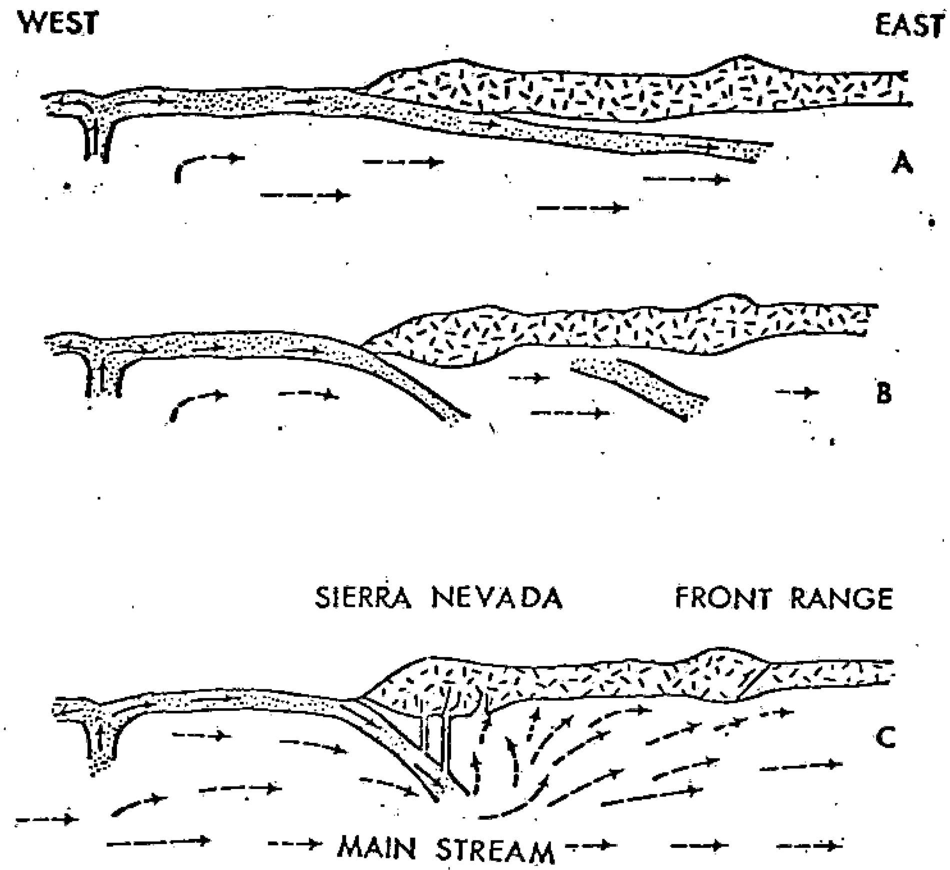


Fig. 14. Alternative plate-tectonic models for North America in "Laramide" time.

Basic theories

Sir Edward Bullard
Professor of Geophysics,
University of Cambridge (United Kingdom)

Introduction

In this chapter a brief account is given of the structure of the earth and of the processes going on within, particularly of the generation and transport of heat. In the last few years there has been a major revolution in geology in which new ideas about the history of the earth and the nature of the motions in it have been developed and widely accepted. Here, the views of the successful revolutionaries are adopted: those of the remnant defending the traditional position cannot be explained at every point. Briefly the traditional position is that there have been no major horizontal motions of large blocks of the earth's crust and that the whole outer part of the earth to a depth of 3,000 km is essentially a solid undergoing only vertical motion; these views have been persuasively and ably expounded by Belousov (1970) and Jeffreys (1970). Here a different view is taken.

Crust, mantle and core

At the surface of the earth a great variety of rock is found; sometimes sediments, such as clays, sandstones or limestones; sometimes ancient 'shields' composed largely of granite; and sometimes lavas which have poured out from volcanoes. The sediments are of great importance for geology—they contain all the oil and a major part of the mineral resources of the earth; they also contain the fossils on the study of which our knowledge of the history of the earth so greatly depends; for the purpose of this article, however, they are not of great significance.

Beneath the sediments of the continents there lies a 'basement' composed predominantly of granite. Some granites have been formed by the freezing of molten material and some by the metamorphism of sediments by heat and by liquids and gases rising from below. Granites are the result of complex thermal, mechanical and chemical processes and have a varied chemical and mineralogical composition. In general it may be said that they represent material with a low melting point and a considerable proportion of free silica in the form of crystalline quartz. In one way or another this material has been separated from more basic rocks (i.e. material containing less quartz) deeper in the earth. The more basic material often appears near the surface as basalt dykes cutting the granites; (a dyke is a vertical sheet of igneous rock that has forced its way upward and solidified).

It is likely that the proportion of basalt increases with depth and in some places there may be a layer of almost pure basalt underlying the granite. The study of the elastic waves from earthquakes shows that at a depth of about 35 km beneath the continents there is a boundary known as the 'Mohorovičić discontinuity' or 'Moho', after its discoverer Andrija Mohorovičić. At this discontinuity the velocity of seismic waves suddenly increases. For compressional waves the increase is from about 7 to 8.1 km/s. The discontinuity represents the lower boundary of the granites and basalts which constitutes the 'crust' of the earth. On the continents the mountain ranges are composed of masses of sediments which have been squeezed, heated and distorted and have been intruded by molten rocks from below. The forms of the mountains as we see them are the result of the carving of such masses by flowing water and moving ice.

In the oceans everything is different. The ocean floor is a new geological world. All the rocks are basalts; there are no granites either beneath the sea floor or on islands (except for a few 'continental fragments', such as the Seychelles, and a few patches of granite formed by the separation of minerals from large masses of molten basalt, as on Ascension Island). Beneath the basalts there is a Mohorovičić discontinuity just as there is under the continents, but it lies only about 5 km beneath the ocean floor, that is 10 km below the sea surface. The structure of the continental and oceanic crust is illustrated in Figure 1. The mountains of the oceans are not carved from folded rocks but are volcanoes; when they emerge above the sea surface they are often capped by coral, but beneath the coral there is always a volcanic core of basalt.

Beneath the crust lies the 'mantle'. For forces lasting for a few seconds, such as those concerned in the propagation of the waves from earthquakes, its material acts as a solid. It also behaves as a solid in the natural vibrations of the earth with periods of up to an hour, for tidal periods of 12 hours and for the periods of around a year involved in wobbles of the earth about its axis of rotation. For the much longer periods involved in geological processes it may be expected to 'creep' as do all solids at temperatures in excess of a few hundred degrees. The most direct evidence for its behaviour for long continued forces is the rise of previously glaciated areas, such as Scandinavia, after the removal of the ice. This rise takes a time of the order of 10,000 years and is compatible with the material of the upper mantle behaving like a fluid with a viscosity of about 10^{21} g/cm.s. There has been much discussion of the probable rheology of the upper mantle and particularly as to whether it possesses a finite 'strength' below which it does not flow. The view taken here is that for periods of up to a day it is a very 'good' solid which literally 'rings like a bell', and that for periods greater than 10,000 years it is a 'Newtonian fluid' which will move under indefinitely small stress differences; (for a discussion of the physics, see McKenzie, 1968 and for a contrary view see Jeffreys, 1970). Its behaviour for periods between a day and 10,000 years is doubtless complicated: at the lower end it is presumably a solid with some type of imperfect elasticity, and at the higher end a liquid whose effective viscosity varies with the stress.

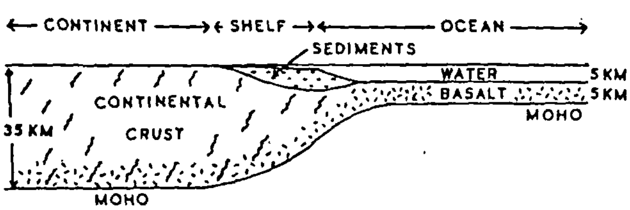


FIG. 1. Schematic section through earth's crust at an inactive coastline, such as that of eastern North America. Under the continent the proportion of basalt probably increases with depth, but it does not necessarily form a separate layer. A great thickness of sediments usually underlies the continental shelf.

The material of the mantle is not directly accessible to observation and our knowledge of it depends partly on deductions from the velocities of seismic waves and partly on the study of rocks at the surface that may be supposed to be derived more or less directly from it. Owing to the shallowness of the mantle in the oceans the study of rocks from the sea floor is of particular importance. The consensus of opinion is that at the Moho there is a marked change in composition and that the material below it is of ultrabasic composition, a peridotite consisting predominantly of the mineral olivine—iron magnesium orthosilicate, $(Mg, Fe)_2SiO_4$. The composition is probably quite similar to that of stoney meteorites.

With increasing depth in the mantle the pressure and temperature both increase and at some point changes in crystal structure may be expected. Seismology suggests a gradual change at depths in the neighbourhood of 400 km which probably corresponds to the changes towards more compact crystal structures that are observed at high pressures in the laboratory. It is possible that the viscosity increases greatly in the transition, but there is little direct evidence as to the amount of the change.

The mantle extends to a depth of about 2,900 km where there is a change to the much denser and liquid core. This is probably composed largely of molten iron. Within this liquid core is an 'inner core' of radius about 1,350 km which may be composed of iron that has been solidified by pressure.

The main facts of the large scale structure of the earth are summarised in Figure 2. For the purposes of this article only the crust and the upper part of the mantle are of importance. What lies below is too remote to affect the surface except on a very long time scale.

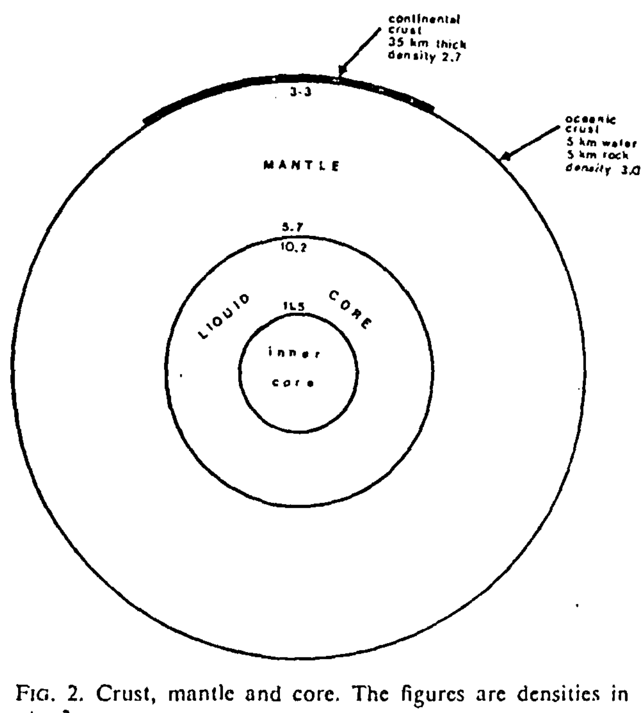


FIG. 2. Crust, mantle and core. The figures are densities in g/cm^3.

Underground temperatures and heat flow

It has been known since the 17th century that the temperature in deep mines exceeds that at the surface of the earth. Such a temperature gradient implies a flow of heat outwards from the interior to the surface and raises many problems concerning the amount of heat involved, its variation from place to place and its origin. If the heat is

TABLE 1. Some typical measurements of heat flow

Place	Depth m	Greatest temperature °C	Mean gradient °C/km	Mean conductivity 10^{-3} cal/s cm °C	Heat flow 10^{-6} cal/cm^{2}s
Gerhardminnebron (South Africa)	3,022	46.4	9.5	13.5	1.28
Tocketts (Yorkshire, England)	906	35.7	27.7	4.26	1.18
Adams Tunnel (Colorado, U.S.A.)	940	24.8	24.1	8.00	1.93
Lat. 48¼° N., Long. 17° W. (E. basin of N. Atlantic)	4,670[1]	0.116[2]	25.4	2.28	0.58
Lat. 46½° N., Long. 27¼° W. (Central Valley of Mid-Atlantic Ridge)	4,109[1]	0.87[2]	315	2.07	6.52

1. Depth of water.
2. Measured temperature difference over depth range of 4.6 m for 1st and 2.8 m for 2nd station.

carried by thermal conduction, the amount emerging per unit area is equal to the product of the temperature gradient and the thermal conductivity. To indicate the orders of magnitude of the quantities in non-volcanic areas a few typical examples are given in Table 1. Table 2 gives typical conductivities of some types of rock: considerable variations are found depending on the composition and water content, particularly for sedimentary rocks.

TABLE 2. Thermal conductivity of rocks, 10^{-3} cal/cm s °C

Granite	6-9	
Dolerite	7-8	
Gneiss	5-9	normal to foliation
Gneiss	6-11	parallel to foliation
Quartzite	7-19	
Limestone	4-7	
Dolomite	9-14	
Sandstone	4-11	
Shale	3-6	
Rock Salt	13-17	
Wet ocean sediments	1.7-2.4	

For more extensive tables see Clark (1966).

In a place where there are no shallow sources the upward flow of heat should be the same at all depths accessible to the drill: the temperature gradient will therefore vary with the conductivity, being less in well conducting sections than in poorly conducting ones. It is not easy to find bores holes or mines in which reliable measurements can be made. Time must be allowed for the heat produced by drilling to dissipate itself; gas, oil, or water must not flow into or out of the hole; and cooling by mine ventilation must be avoided. Moreover in many places the emerging heat is not all transported by conduction: some may be carried by circulating water. A striking example is shown in Figure 3.

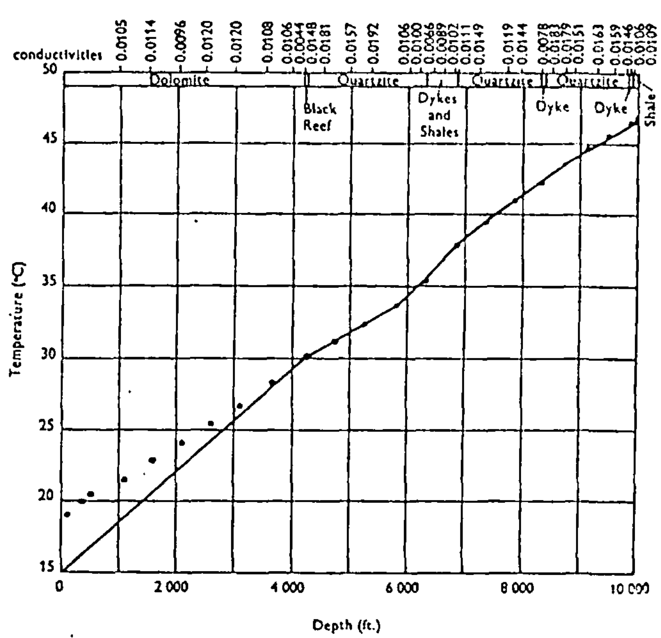


FIG. 3. The variation of temperature with depth in the Gerhardminnebron bore in the Transvaal. The dots show the observed temperatures and the full line the variation expected for a constant heat flow fitting the observations in the lower part of the hole. In the dolomite that occupies the upper 4,000 ft, the observed gradient is much less than expected: this is due to the free circulation of water in the fissures which carries part of the heat. (From: *Proc. R. Soc.*, ser. A, vol. 173, p. 489, 1939.)

At sea there is less possibility of the circulation of water in the sediment, and meaningful measurements can be made over a much shorter range of depths. Probes, 2 to 5 m in length, are forced into the bottom and by this means a large number of measurements can be obtained relatively quickly.

There are at present a few hundred places on land and a few thousand at sea where both temperature gradient and conductivity have been satisfactorily measured. The distribution of the values has been studied by Lee and Uyeda (Lee, 1965); the results are shown in Figure 4. The majority of the values give a more or less Gaussian distribution with a maximum at about 1.1 μcal/cm²s and with most of the measurements between 0 and 2.5 μcal/cm²s. On the high heat flow side of the peak there is a long tail extending beyond 8 μcal/cm²s. The distributions for the individual oceans and continents, taken separately, are very similar. The mean for the whole earth is 1.4 μcal/cm²s.

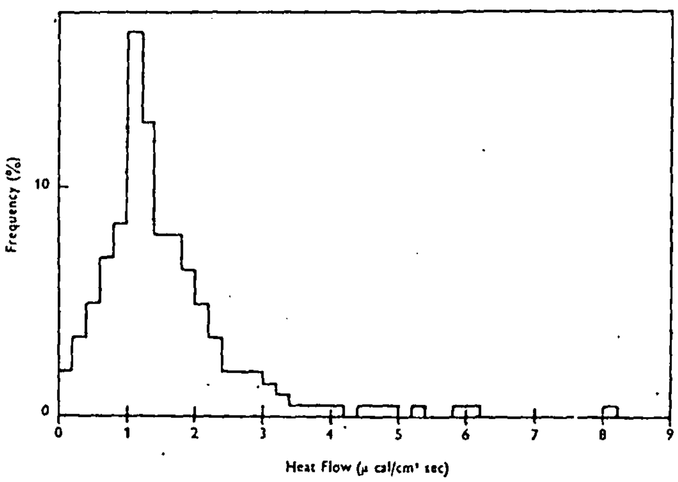


FIG. 4. Histogram showing the distribution of the measurements of heat flow for the whole world (excluding geothermal areas). (From: Lee, 1965.)

When the observed values of the heat flow are examined in more detail it is found that different geological regimes have systematically different heat flows. The continental shields give low values averaging about 0.9 μcal/cm²s; the recent mountain ranges tend to give high values around 2; the values on the crests of the ocean ridges are very high, often between 3 and 8; high values are also found in the inland seas behind island arcs, such as the Japan Sea. Very low values, around 0.6, are frequent in the oceans a few hundred kilometres on either side of the ocean ridges. Very high values are, naturally, found in volcanic and hydrothermal areas of the continents.

The source of the heat

A flow of 1.2 μcal/cm²s is not a large flux of heat by the standards of the kitchen or the factory. In a geological setting, however, it is impressive owing to the long period over which it is maintained. The earth is 4,500 million years old and of this the last 3,500 million years are represented by rocks near the present surface. In 1,000 million years a heat flow of 1.2 μcal/cm²s implies a total heat flow of 4×10^{10} cal/cm². If this were produced by the combustion of coal it would require the consumption of about 4 tons under each square centimetre of the earth's surface—that is the complete burning of a layer 20 km thick. Clearly the heat cannot be produced by burning coal or, indeed, by any chemical reaction near the surface, and clearly also some very powerful source of heat is needed to produce the observed flow over geological time-spans.

Some of the heat may be the original heat with which the earth was endowed when it was formed. Since the initial temperature is unknown there is some doubt about how large a part this may be. Lord Kelvin showed that if heat is brought to the surface by conduction in a solid earth, then, even if it was initially at the melting point, after 100 million years the original heat would be only a small fraction of that now observed. If heat is brought up by motions in the mantle, Kelvin's argument would need modification; it seems unlikely, however, that original heat contributes any large part of the observed flow. It is possible to suggest a number of other sources of heat within the earth, the outstanding one being the radioactivity of rocks.

All rocks contain small amounts of radioactive elements of which the only ones giving important amounts of heat are uranium, thorium and potassium and their products. Uranium has two long lived isotopes; U^{238} decays to lead (Pb^{206}) through a long series of intermediaries, and U^{235} decays similarly to Pb^{207}. Thorium has a single long lived isotope, Th^{232}, which decays, again through many intermediate stages, to Pb^{208}. The rare isotope of potassium, K^{40}, decays by two routes—one to Ca^{40} and the other to A^{40}. The half lives and heat productions are summarised in Table 3. The amounts of heat generated in different types of rocks are given in Table 4.

TABLE 3. Heat production by long-lived radioactive isotopes and their products

Isotope	Half life 10^9 yr	Proportion of isotope %	Heat generation[1] cal/g yr
U^{238}	4.50	99.27	0.70 } 0.73
U^{235}	0.71	0.72	0.03 }
Th^{232}	13.9	100	0.20
K^{40}	1.31	0.012	27×10^{-6}

1. Per gram of the chemical element including all isotopes.

From the figures in Table 4 it is clear that there is no difficulty in accounting for the observed heat flow. In fact, since a thickness of 14 km of granite would produce a flow of 1 μcal/cm²s it is necessary to assume that the rocks below the crust are much less radioactive than crustal rocks; otherwise the heat flow would be much greater than is observed. Such an assumption is quite reasonable since the more basic rocks are much less radioactive than the granites.

TABLE 4. Typical heat production in rocks

Rock type	Concentration U ppm	Th ppm	K %	Heat production (μ cal/g yr) U	Th	K	Total
Granite	4.7	20	3.4	3.4	4.0	0.9	8.3
Basalt	0.6	2.7	0.8	0.44	0.54	0.23	1.21
Peridotite	0.016	0.004?	0.0012	0.012	0.001	0.0003	0.013

There is some direct evidence that the heat flow through the continental shield areas is, in large part, due to radioactivity in the crust. It is found that the heat flow is greater in areas of high radioactivity than it is in areas where the rocks are less radioactive (Roy *et al.*, 1968). Some results of this kind are shown in Figure 5; by extrapolating the observed values back to zero radioactivity, the heat flow for a non-radioactive crust can be estimated—i.e. the amount of heat coming from the mantle. It seems that on the continents this varies from one region to another in the range one third to two thirds. In the oceans there is only about 5 km of rather feebly radioactive basalt above the Moho and most of the heat must come from the mantle.

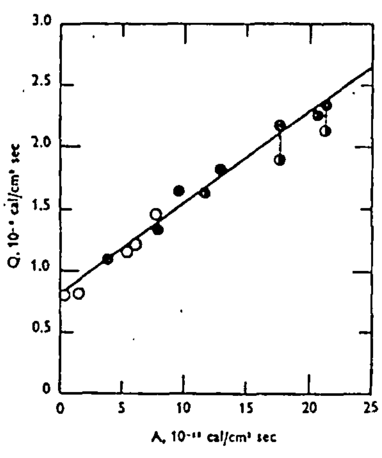


FIG. 5. Dependence of heat flow, 'Q', in Eastern and Central U.S.A. on the radioactivity, 'A', of the granitic basement rocks. (From: Roy *et al.*, 1968.)

Knowing the heat flow at the surface it is possible to make a rough estimate of temperatures all through the crust and in the upper part of the mantle, the main uncertainty being the distribution of radioactivity in depth. If, in a continental area, the heat flow is 1.2 μcal/cm^{2}s, and if half of this comes from a crust 35 km thick with a conductivity of 0.005 cal/cm s °C then the temperature at the Moho would be 630 °C above that at the surface. Obviously there is great uncertainty, but something in the range 500 to 700 °C seems reasonable. If the material below the Moho were static and the heat were transported by conduction, the gradient just below the Moho would be 12 °C/km and a temperature of about 1,400 °C would be reached at a depth of 100 km. The melting point increases with depth, but it seems likely that the temperature increases more rapidly and that the material of the mantle is near its melting point at depths of about 100 km. At greater depths the increase of melting point may exceed the rise in temperature and the material is probably in less danger of melting. The approach to melting at depths around 100 km is perhaps the cause of a slight decrease which is found in the velocity of elastic waves in this region.

Volcanoes, ridges and island arcs

The picture of an earth with a slightly radioactive crust and a still less radioactive mantle gives temperatures greatly below the melting point all through the crust. It is therefore an earth normally without molten rock near the surface and therefore without volcanoes or igneous activity in the crust. In a sense this is a good first approximation. Active volcanoes are exceptional features that occur in special places; they are not usual features of the landscape.

There are, essentially, two series of volcanoes: those around the island arcs and those on the mid-ocean ridges. To understand why this should be so is one of the central problems of geology. It is easy to see why volcanoes do not occur everywhere, but why should they occur on these two very different kinds of lines? Why also should the lavas of the two types be different? To understand these matters it is necessary to digress a little to explain the nature of mid-ocean ridges and island arcs and the current theories of their origin.

The mid-ocean ridges are shown in Figure 6; they are a great series of submarine mountain ranges which run around the world and are by far the longest chain of mountains on earth. The mid-Atlantic ridge is typical; some cross sections are shown in Figure 7. The ridge is completely different from a continental mountain range. Mountains such as the Alps are carved from great piles of compressed and distorted rocks, largely sediments; the mid-ocean

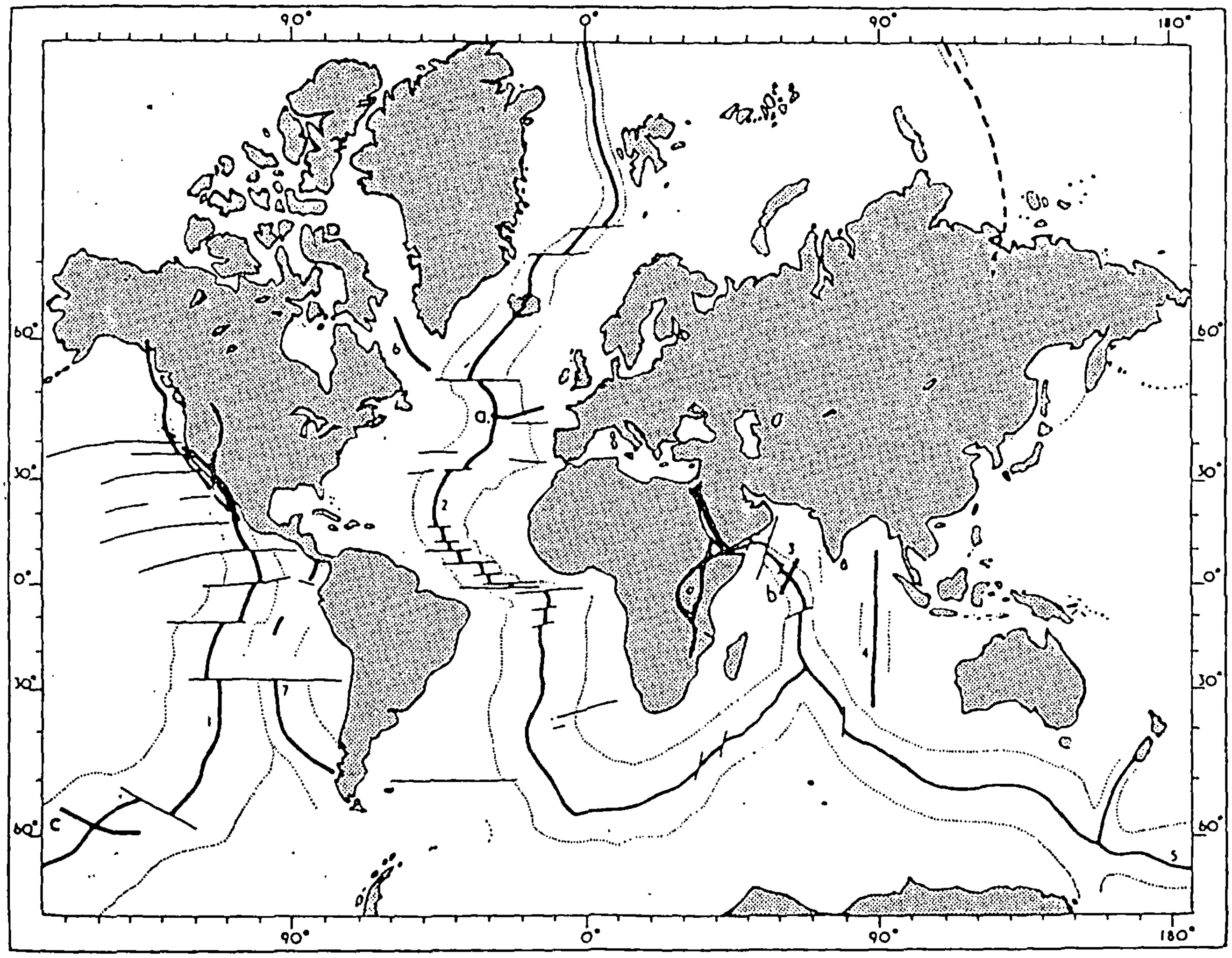
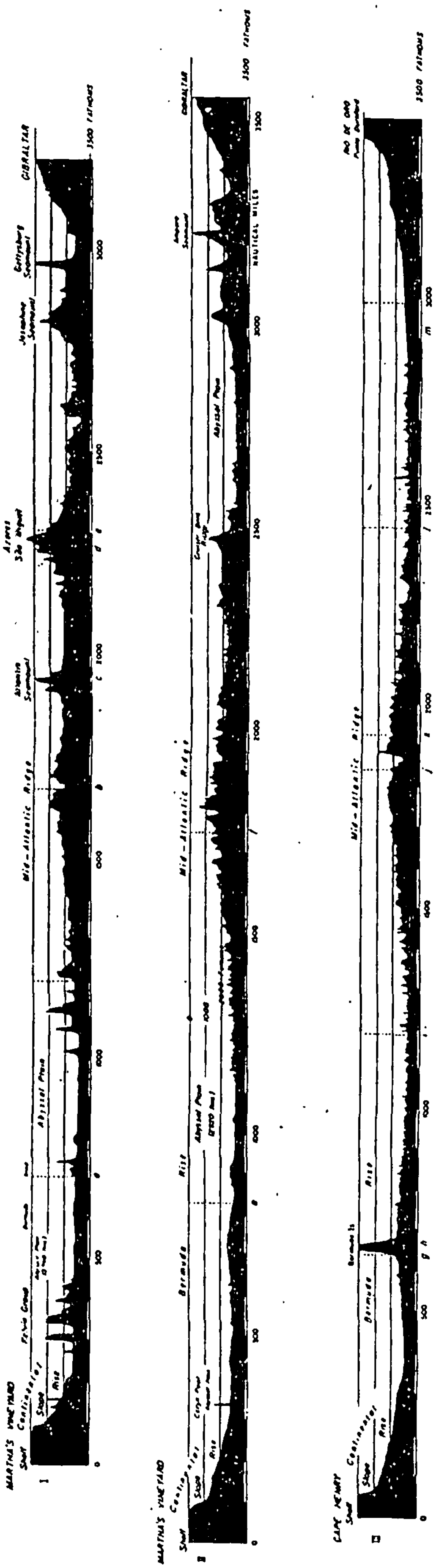


FIG. 6. The Mid-Ocean ridges. The heavy line represents the crest of the ridge and the central valley: the thin lines across the ridge are transverse faults. (From: D. H. Matthews, *International Dictionary of Geophysics*, p. 981.)

ridges, on the other hand, are composed entirely of submarine volcanoes. Along the crest of the ridge runs a steep sided, crack-like valley which is the seat of numerous earthquakes and also of volcanic eruptions. A study of the earthquakes has shown that the valley is an opening crack which opens a little more at each earthquake. As the crack opens, lava flows out and forms new volcanoes. It is not possible to discuss all the evidence here but there is now no doubt that the floor of an ocean behaves as two rigid plates that move apart by continual splitting along the axis of the ridge; reviews by Bullard (1969) and Menard (1969) provide more detail. In many places the ridge consists of sections offset along transverse faults, as may be seen from Figure 6. The sections of these transverse faults that lie between the two ridge crests have earthquakes; those further out, beyond the crests, do not. In the earthquakes the motion is parallel to the fault as would be expected if two plates are separating along the ridge axis (Figure 8).

Outpourings of lava may occur at any point along the central valley of the mid-ocean ridge. Most of these will be beneath the surface of the sea and will pass unnoticed, but some will form volcanic islands. Such islands are seen at intervals along the whole length of the ridge; sometimes they are on the ridge crest but more often they are on the transverse faults. In the Atlantic there are Jan Mayen in the far north; Iceland, one of the few places where the central valley can be seen on land; the Azores where the volcanoes are spread out east-west along a transverse fracture rather then along the ridge crest; Ascension Island; St. Helena (well clear of the crest); Tristan da Cunha and Gough



◀ FIG. 7. Three sections across the North Atlantic Ocean, showing the mid-ocean ridge and its central valley. The vertical scale is exaggerated 40:1. The upper section passes through a transverse ridge from the Azores to Gibraltar: consequently the mid-ocean ridge appears less prominent than in the other sections. (From: D. C. Heezen and H. W. Menard, *The Sea*, vol. 3, p. 274.)

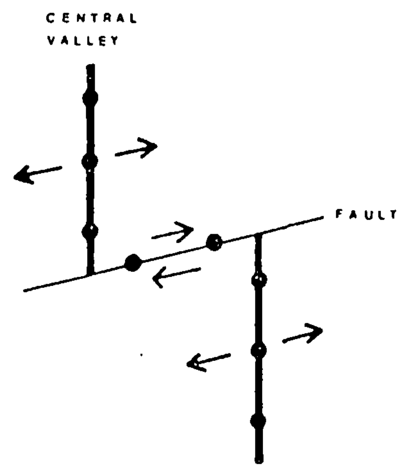


FIG. 8. Separation of plates along the central valley of a ridge. The motion must be parallel to the transverse fault. The earthquakes (●) are confined to the central valley and the part of the fault between the two sections of valley.

Island. From the South Atlantic the ridge sweeps round to the South of Africa into the Indian Ocean. Here Marion and Prince Edward Islands are recently extinct volcanic cones. Further north Amsterdam and St. Paul Islands also lie on the ridge. The section of the ridge running north into the Gulf of Aden has no islands; it is continued in the Red Sea and is probably also connected with the eastern branch of the African Rift valley which shows at least one active volcano and many that are recently extinct. In the centre of the Indian Ocean the ridge forks: one branch goes around South Africa and one north as already described; the third goes south of Australia and New Zealand and enters the eastern Pacific. In the Pacific the ridge runs north and enters the Gulf of California. This section has no central valley and no active volcanoes; it has, however, a number of islands bearing recently extinct volcanoes, for example, those of Easter Island. The San Andreas fault runs from the northern end of the Gulf of California to Cape Mendecino where it runs out to sea. This fault is probably a transverse fault joining two sections of ridge, the southern piece in the Gulf of California and the more northerly piece running from near Cape Mendecino roughly parallel to the coast as far as the north end of Vancouver Island.

The course of the ridge has been described in some detail to show what part of the earth's active volcanoes lie on it; of areas that are actually being used for the production of steam or hot water we have Iceland and the Imperial valley of California. There are also a number of places where geothermal power could probably be developed if it were required; for example the Kenya Rift Valley, the Azores and perhaps the area of Abyssinia and Somaliland to the west of the Gulf of Aden. Unfortunately most of the ridge crest is beneath the sea and therefore inaccessible for geothermal development; the high heat flows measured in the central valley show that temperatures of several hundred degrees must be reached at depths of a few kilometres.

It has been said above that the ocean floors behave as rigid plates spreading out from the ridge axes and being continually renewed there by intrusion and extrusion of molten rock from below. The rate at which the plates spread differs from place to place, but is usually in the range of 1 to 10 cm/yr on each side of the axis. Such rates are, geologically speaking, very fast; spreading at 5 cm/yr on each side of a ridge would give a relative motion of 10,000 km in 100 million years. Since the earth is 4,500 million years old it is clear that plates must be destroyed as well as created. There is not much choice for the site of the destruction and all those who believe in moving plates believe that they are destroyed at the ocean trenches.

The Tonga trench is typical and has been particularly closely studied. The Tonga group of islands is a north-south chain of extinct volcanoes with a deep ocean trench running parallel to the islands and a little to the east of them. A plate is moving westward from the section of the ridge running north-south in the eastern Pacific. Spreading is fast, and in about 100 million years a piece of lava formed on the crest of the ridge would reach the trench. Here the moving plate dives down at about 45° and passes under the islands. The reality of this description is established by the earthquakes that occur within the plate as it moves and fractures. The earthquakes behind the Tonga trench occur at all depths down to 650 km. Earthquakes at such great depths occur only in places where there is reason to suppose that there are sinking plates; most are behind island arcs, such as those of Indonesia, Japan and the Aleutians, but they also occur beneath the Andes where there is an 'ocean trench' and a descending plate, but no island arc. The greatest depth of any earthquake is about 750 km; by the time the plate reaches this depth it will be near its melting point and unable to store the elastic energy needed to produce an earthquake. It is probable that the material of the plate finally becomes mixed again with the mantle material from which is was derived. The sediments on its surface may be partially scraped off and piled up just behind the trench and may contribute to the piles of distorted rocks found in the mountain ranges that often lie along the edges of continents. These mountains frequently contain volcanoes producing andesitic lavas, as in the Andes; and even where there is no great mountain range the volcanoes are often present, as in Indonesia. It is likely that the andesitic lavas are derived from the first products of melting of the sinking plates. Such melting would be expected at depths of about 150 km: with a plate dipping at 45° the lava would emerge at about this distance behind the trenches, assuming that it works its way vertically upwards from its source to the surface.

The areas of downgoing plates contain a large proportion of the geothermal areas of the world. Many of these are around the Pacific and include Indonesia, the Philippines, Japan, the Aleutians and the Andes. The New Zealand geothermal area is a southward extension of the Tonga trench and is probably associated with a sinking plate. Other areas where similar phenomena occur are the Caribbean, the Aegean and the Scotia Arc which runs between the southern tip of South America and Graham Land. There are some important geothermal areas whose positions in this scheme are not clear. The Mediterranean is an area in which very complicated events are in progress. Africa is moving northwards and colliding with southern Europe and the Middle East. In the collision the plates have become fragmented and directions of motion have changed. It is probable that the main, north-going African plate is going down and being consumed beneath the Cretan island arc, but it is not clear what is happening in Italy; it is possible that there is a small downgoing plate moving in from the west. Another area of doubt is in the mountain states of the U.S.A. The continent has moved westwards away from Europe and has probably overrun its own westward boundary trench and also part of a mid-ocean ridge. It is also possible that California is a continental fragment that had its origin somewhere to the west and has collided with the rest of the U.S.A. In view of these complexities, which are typical of continental geology, it is difficult to say what is the origin of the geothermal activity in north-east California and Nevada. It is clear that it is connected with the great outpourings of lava but it is not clear what is the origin and history of these.

There are many coastlines of the world where there are no earthquakes, no volcanoes, no downgoing plate and no geothermal areas. These coastlines, such as the eastern seaboard of the United States, the west coast of Europe and both the east and west coasts of Africa, are places where the ocean floor and the continent are parts of the same plate. The major plates are shown in Figure 9; they are delineated by the earthquakes of the mid-ocean ridges, which are associated with their generation, and those of the trenches which signal their destruction. There are in addition smaller plates in areas such as the Mediterranean which are imperfectly understood and cannot be shown in a small diagram.

In spite of the doubts about the meaning of geothermal phenomena in some areas, the suggestion that there are two main environments for igneous activity is a useful unifying concept in considering the phenomena. For any particular example it is useful to ask: 'Is this a place where the crust is splitting and new material is being added to a plate in the form of basaltic lavas and intrusions, or is it called 'thermal convection'; once it starts, most of the heat is transported thereby and conduction is unimportant except in a layer near the surface. As has already been explained, it is likely that the mantle of the earth behaves as a liquid for long continued forces—that is, it has no long-term strength and no power to prevent convection once the adiabatic gradient is exceeded. The adiabatic gradient is '$g \alpha T/C$' where 'α' is the coefficient of expansion of the material, 'T' is the absolute temperature, 'g' the acceleration of gravity and 'C' the specific heat. All these quantities, except the coefficient of expansion, are quite accurately known; the coefficient of expansion of rock in the mantle will be less than it is at atmospheric pressure and, from extrapolation of experiments at high pressure and from theory, may be estimated at about $10^{-6}/°C$. With this value the adiabatic gradient is about 1 °C/km, which is certainly less than the temperature gradient in the mantle if it were static and the heat were brought up entirely by conduction. It is therefore not unreasonable to expect convective motions in the mantle. The form that these motions will take is difficult to predict: it depends strongly on the rate of change of viscosity with depth and with temperature and on the thickness of the layer taking part in the motion. The occurrence of long, narrow, linear features such as ridges and trenches suggests that the motion may be in long rolls rather than in isolated plumes or jets. A realistic theory would be quite complicated since it must allow for the thermal effects of the motions of the plates, for the effects of the cold sinking plates on the temperature distribution and for the differences in the distribution of radioactivity between oceans and continents. The heat engine may move the plates and may be the driving force of geological change, but we must also not ignore the effect of the moving plates on the heat engine. So far only rather simple models have been treated and much remains to be done before we have a realistic physical and mathematical theory of large scale geological processes.

The transfer of heat to the surface

As molten rock approaches the surface of the earth a very complicated series of processes will occur. The pressure will be reduced, gases and liquids held in solution will be released, the temperature will fall and crystallisation will begin. A lava is a complicated mixture of oxides and silicates and is capable of producing many different solid substances depending on the pressure, the temperature and the amount of water and other volatiles present. In general the first substances to crystallise will be those of highest melting point: these will be basic rocks without free quartz. The crystals will be denser than the liquid and will tend to settle to the bottom of the molten material; how far they are able to do so will depend on the size of the crystals, the viscosity of the melt and the geometry of the cavity in which it is contained. The composition of both the melt and of the separating crystals may also be affected by the eating away and solution of material from the walls of the cavity. The details of all these processes form a large part of the subject matter of the science of petrology and cannot be discussed in any detail here. In general there will be a tendency for anything that will not form high melting compounds, or whose atoms will not easily fit as impurities into the common silicate minerals, to be left over as gases and liquids in the late stages of igneous activity. In particular, hot water and steam will be released and will carry away a great variety of other materials by 'steam distillation'. In this way many relatively rare elements may be concentrated. Deposits of elemental sulphur may be formed and its gaseous compounds, such as H_2S and SO_2 may emerge; even extremely rare elements with volatile compounds, such as mercury and silver, may occur as ores in workable concentrations.

In fact, a large part of the water and steam emerging in geothermal areas is not derived from deep seated molten rock but from the circulation of ground water. Lavas are full of joints and cracks and it is easy for rainwater to penetrate deeply and reach the hot lava; here it is itself heated and caused to rise, taking with it dissolved materials. When it reaches the surface it may emerge as steam or hot water or as a mixture of the two.

Some idea of the proportion of 'new' water—that is, water coming from the interior of the earth and not derived from rainwater—may be obtained from the study of the ratio of the isotopes of hydrogen in the water. Hydrogen has two stable isotopes; ordinary hydrogen, H, with atomic mass 1, and heavy hydrogen, D, (deuterium) with mass 2. The light isotope is much the more abundant. The ratio of the numbers of the two kinds of atoms in sea water is a little variable but is typically $H/D = 6{,}400$. Most of the deuterium will be combined in water molecules HDO; (since deuterium is so rare, there will be very few D_2O molecules). Owing to their greater mass the HDO molecules have a lower vapour pressure than H_2O molecules and, when water is evaporated from the sea, a higher proportion of the H_2O molecules will go into the vapour. It would therefore be expected that the water that falls as rain, which is derived from water vapour evaporated from the sea, would contain less deuterium than does sea water. This is found to be so: the ratio H/D in rain water, river water or tap water being typically 6,800. The ratio for water derived from geothermal steam is close to that for rainwater, and this suggests that most of the steam is derived from rainwater circulating through cracks in the rocks. The uppermost stage of a geothermal area is therefore a heat exchanger in which the heat of the rocks is transferred to circulating water. A review of these complicated matters will be found in Tongiorgi (1963).

Conclusion

Geothermal areas are of interest as sources of heat and power and they present important and challenging engineering problems which form an important part of the

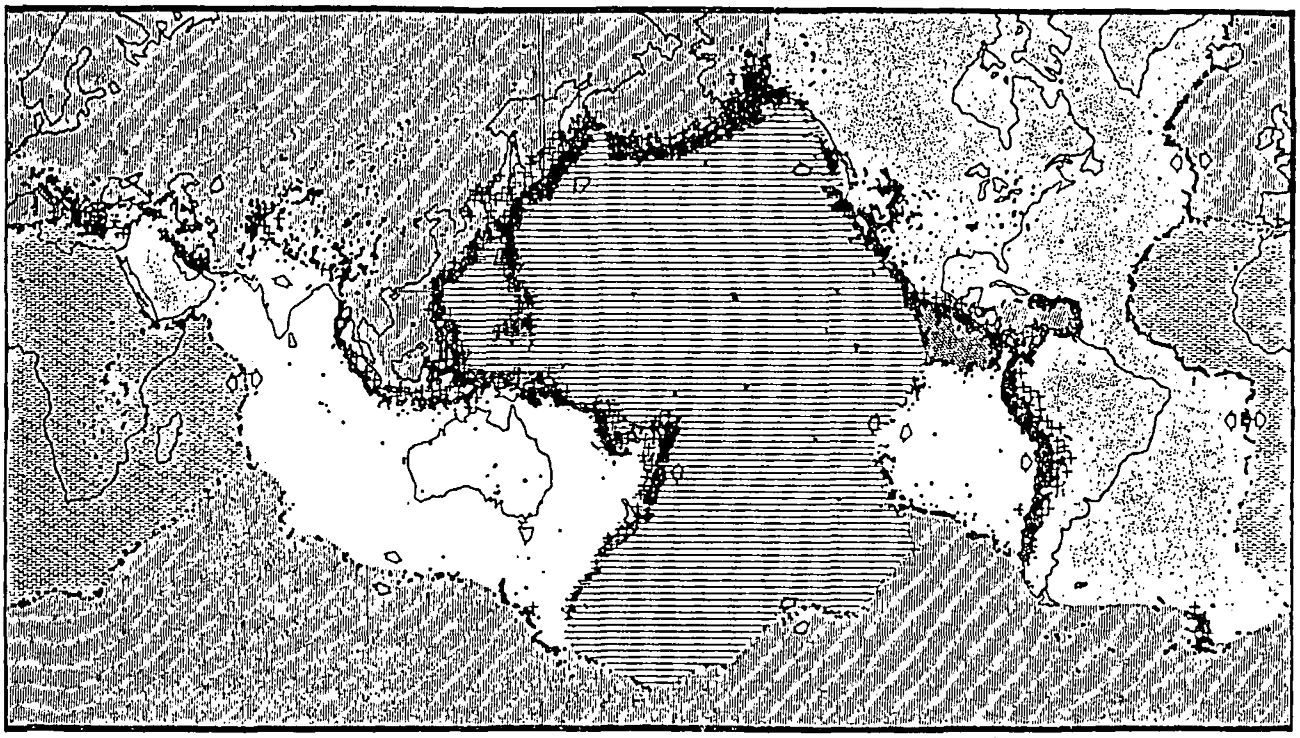


Fig. 9. The moving plates. The arrows show the direction of motion. The plates are bounded by earthquake zones: shallow earthquakes are shown by dots, and those deeper than 150 km by crosses. Six large plates are shown—Europe/Asia, Africa, Eastern Indian Ocean, Antarctic, North and South America, and the Pacific. Some, but not all, of the small and medium sized plates are also shown—South-east Pacific, Arabia, Caribbean.

a place where a plate is being destroyed and melted and andesitic lavas are working their way to the surface?'.

During the last year the ideas of 'plate tectonics' have received important confirmation from drilling in the ocean floor by the U.S. project J.O.I.D.E.S. (Joint Oceanographic Institutions for Deep Earth Sampling). If our ideas are correct, the youngest rocks should appear at the ridge crests and in the central valley; some way on either side we should expect recent sediments at the sea floor with older sediments beneath. The age should increase with depth till sediments are reached that were deposited when the rocks at the drilling site were on the ridge axis. Below these sediments there should be lavas of the same age as these sediments. These predictions have been triumphantly verified in the 130 holes so far drilled in both the Atlantic and the Pacific oceans. The rates of spreading of the plates determined quite independently of the drilling give very good estimates of the ages of the oldest sediments found in the bore holes.

The mechanism of plate motion

Although we have now acquired a rather definite idea of what is happening at the surface of the earth, we know very little about the causes. We know how the plates are being formed, are moving and are being destroyed, but what is driving them? It seems likely that the earth is, in some sense, a heat engine. Heat is continually being generated by the decay of radioactive elements and this is likely to be the ultimate source of the energy we see displayed in the motions at the surface.

If heat is generated in a layer of liquid the temperature will rise and heat will flow out by thermal conduction. This involves no motion of the material; but, if the rate of heat generation and thus the temperature gradient exceeds a limit known as the 'adiabatic gradient' the liquid becomes unstable and motion starts. Hot liquid rises towards the surface where it cools and sinks again. Such a motion is theme of this Review; but it is well to remember that the power extracted from a geothermal area is a manifestation of the workings of the great internal heat engine which drives the processes of geological change. It is not by chance that the geothermal areas lie where they do, but only in the last few years have we begun to understand the hidden workings of the engine and the reasons for what we see at the surface.

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Deep Mantle Convection Plumes and Plate Motions[1]

W. JASON MORGAN[2]
Princeton, New Jersey 08540

Abstract Evidence shows that volcanic island chains and aseismic ridges are formed by plate motion over fixed-mantle "hot-spots" (Iceland, Hawaii, Galápagos, etc.) and new arguments link these hot-spots with the driving mechanism of continental drift. It is assumed that the hot-spots are surface expressions of deep mantle plumes roughly 150 km in diameter, rising 2 m/year, and extending to the lowest part of the mantle. The rising material spreads out in the asthenosphere, producing stresses on the plate bottoms. Order-of-magnitude estimates show these stresses are sufficiently large to influence plate motion significantly. The total upward flow in the plumes is estimated at 500 cu km/year, which would require the entire mantle to overturn once each 2 billion years.

Introduction

We may account for the main features of the Hawaiian Islands (the long linear chain, the uniform progression of ages toward the northwest, the transition from the tholeiitic main stage to more alkalic later stages of volcano growth) by assuming that the Pacific plate is moving northwestward over a fixed-mantle "hot-spot." Likewise the Greenland-Iceland and Iceland-Faeroe ridges emanating from Iceland, and the Rio Grande and Walvis ridges emanating from Tristan da Cunha and Gough Island, may be interpreted as the result of plates moving away from fixed hot-spots located on the crest of a spreading mid-ocean rise. Wilson (1963a, b; 1965a, b) advanced this hypothesis for the origin of island chains and aseismic ridges, and in a sequence of papers developed how these features may be used to determine the present motion of each plate, how the aseismic ridges are important guides in reconstructing pre-drift continental configurations, and how aseismic ridges and transform faults interrelate. Morgan (in press) has presented three additional lines of evidence supporting the concept of rigid plates moving over fixed-mantle hot-spots; this evidence is summarized in Figures 1–3.

In Figure 1, I quantify observations of the parallelism of the Pacific island chains and the continuity of the Hawaiian Islands and Emperor Seamount chains. Four sites of present volcanism are noted: (1) the Juan de Fuca Rise near Cobb Seamount, (2) Hawaii, (3) MacDonald Seamount (for a report on its discovery see Johnson, 1970), and (4) the Pacific-Nazca rise near Easter Island. The four heavy lines in Figure 1 were generated by rotating the Pacific plate backward in time over these four fixed hot-spots, first 34° about a pole at 67°N, 73°W (0–40 m.y.), then 45° about a pole at 23°N, 110°W (40–100 m.y.). The close agreement of these predicted lines with the trends of the Gulf of Alaska seamount chains, the Hawaiian-Emperor chain, the Austral-Gilbert-Marshall chain, and the Taumotu-Line chain substantiates this hypothesis. Even more exact agreement can be obtained by removing the constraint of fixed hot-spots and allowing the hot-spots to migrate at about ½ cm/year—a small fraction of the roughly 7-cm/year motion of the Pacific plate.

Figure 2 shows paleomagnetic pole positions determined from seamounts in the Pacific. The letter shown with each circle of confidence identifies the seamount group used by Francheteau *et al.* (1970); the number shows the age of the pole in millions of years. The heavy line is the Pacific polar-wandering curve predicted by the motion of the Pacific plate shown in Figure 1. (It is implicitly assumed that, at least during the past 100 m.y., the geomagnetic pole has not wandered relative to the hot-spots fixed in the lower mantle.) The paleomagnetic evidence thus verifies the plate motion predicted by the hot-spot trajectories. This test should be repeated for each of the other major plate units.

Figure 3 shows the present plate motion over the fixed hot-spots. This figure was constructed by finding the relative plate motions deduced from fault strikes and spreading rates on the

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[2] Department of Geological and Geophysical Sciences, Princeton University.

Kenneth Deffeyes first made several of the order-of-magnitude estimates shown herein. I thank him and many others at Princeton for their contributions to this problem. This work was partly supported by the National Science Foundation and the Office of Naval Research.

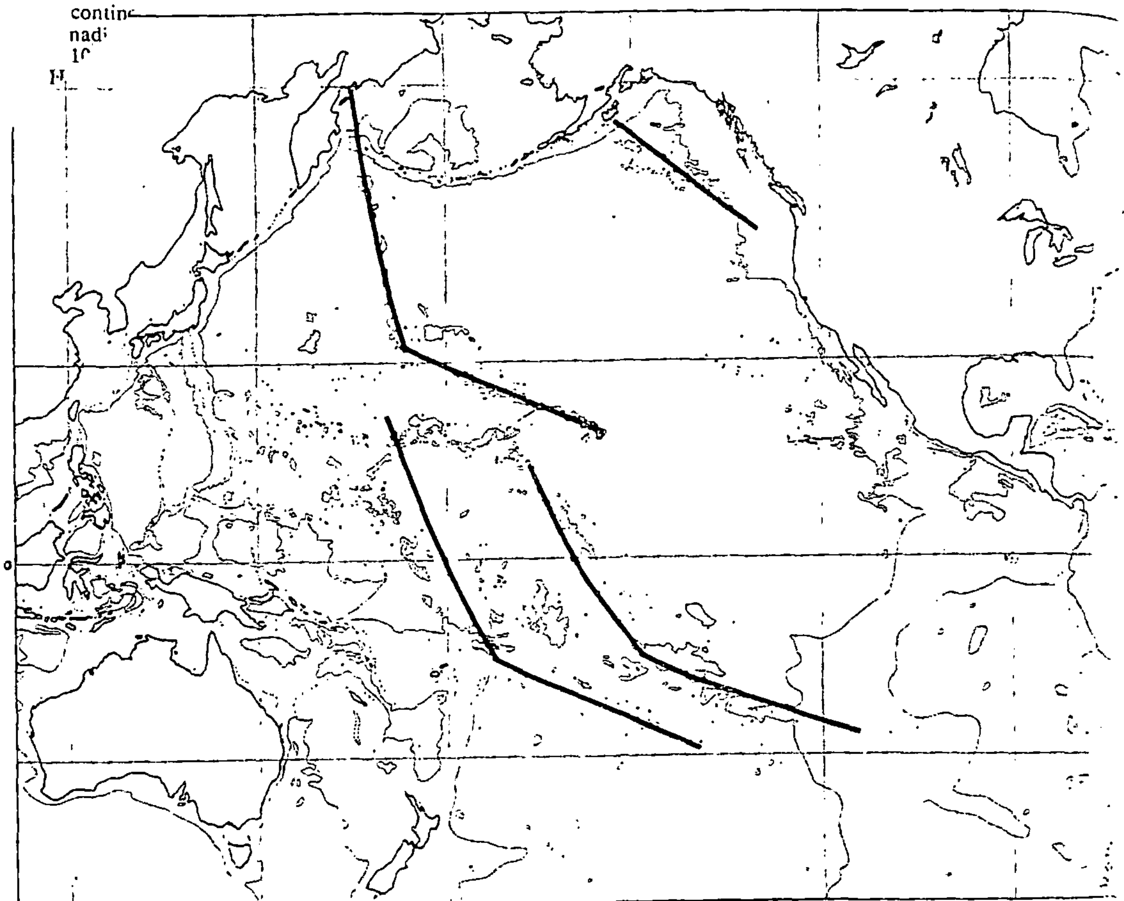

FIG. 1—Hot-spot trajectories constructed by rotating Pacific plate over four fixed hot-spots.

rise boundaries, and then adding a constant rotation to make the Pacific plate rotate properly over its hot-spots. If the synthesis tabulated by Morgan (in press) is correct, and if all the hot-spots are fixed in the mantle, then the velocity vectors shown in Figure 3 should predict accurately the trends of the island chains/aseismic ridges away from hot-spots.

Morgan (1971; in press) proposed that these hot-spots are surface manifestations of lower mantle convection which provides the motive force for continental drift. Assume that about 20 deep mantle plumes bring heat and relatively primordial material up to the asthenosphere, producing horizontal currents in the asthenosphere which flow radially away from each plume. The points of upwelling have unique petrologic and kinematic properties, but I assume there are no corresponding unique points of downwelling—the return flow is uniformly distributed throughout the mantle. The deep convection thus has a thunderhead character, whereas the shallow convection, constrained by the rigid plates at the top surface, has a roll or two-dimensional character.

Some of the consequences of the interaction of rigid plates with localized upwellings and an interpretation of the observed petrologic differences of oceanic island type basalt and oceanic ridge type basalt were presented by Morgan (1971; in press). In this paper I shall amplify the arguments supporting the claim that the hot-spots provide the driving force for continental drift. These arguments fall into three categories: (1) the observation that most hot-spots are near rise crests and evidence that hot-spots become active before continents split apart; (2) an interpretation of the gravity and topography around each hot-spot, showing that the mantle plumes generate moderately large stresses; and (3) estimates comparing the magnitude of stresses generated by plumes to the magnitude of rise and trench stresses.

Location of Hot-Spots

The primary criterion for selection of the hot-spots in Figure 3 was recent volcanic islands not associated with andesitic trench-type activity. Several volcanic islands were removed from this list based on the assumption that some volcanic activity is delayed by a magma-storage mechanism somewhere in the lithosphere. For example, the oldest rocks on Heard Island on the Kerguelen Ridge are 40 m.y. old, but there has been some activity in historic times. I assume that the entire Kerguelen Ridge is a hot-spot feature and that the recent activity on Heard Island is a delayed action of hot-spot material placed there in the lithosphere 40 m.y. ago; thus Heard is eliminated from the present hot-spot list. A more puzzling case is the Cameroon Trend, which lines up with the ridge that heads northeast from St. Helena. I have tentatively assumed that the Cameroons are a delayed action of the St. Helena hot-spot, and that the Cretaceous volcanic rocks northeast of Mt. Cameroon are the expression of the St. Helena hot-spot prior to the breakup of South America and Africa.

Four oceanic centers of volcanism are not near mid-ocean rises: Hawaii and MacDonald in the Pacific plate, the Canary Islands in the African plate, and the Comores Islands in the Somalian plate. In addition, Yellowstone (and the Snake River basalts), Tibesti (in céntral Sahara), and Mount Kenya have characteristics suggestive of continental hot-spots. The Yellowstone, Kenya, and Comores hot-spots are near present-day breakups in the western

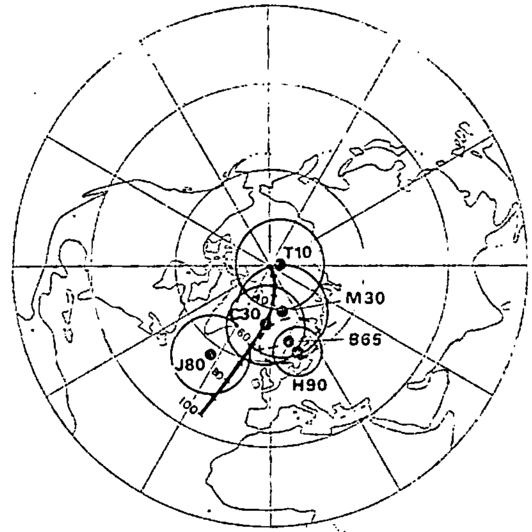

Fig. 2—Pacific paleomagnetic pole positions (adapted from Francheteau *et al.*, 1970) and polar-wander curve predicted by motion of Pacific plate shown in Figure 1.

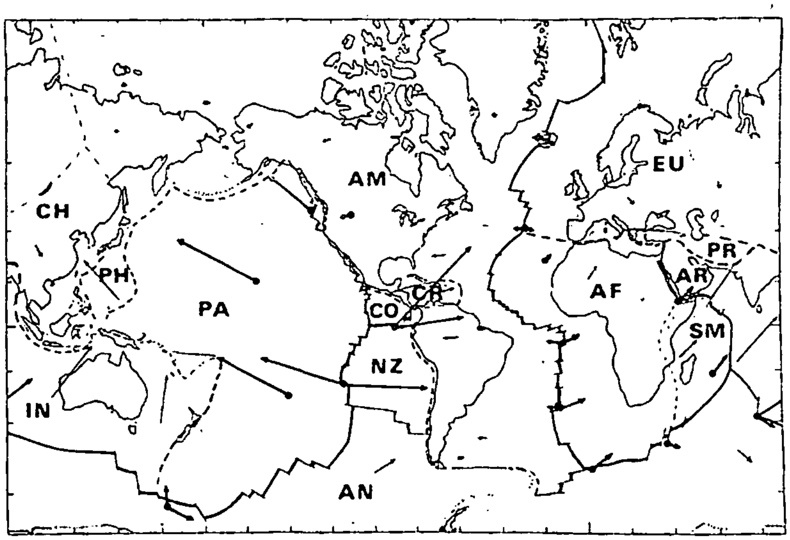

Fig. 3—Present motions of plates over hot-spots. Relative plate motions were determined from fault strikes and spreading rates on rise boundaries; with an appropriate constant rotation added, absolute motions of each plate over mantle were determined. Lengths of arrows are proportional to plate speed.

United States and East Africa, therefore only four of the locations listed are far from present spreading axes. In contrast, 14 hot-spots are near rise axes; four are in the Pacific (Juan de Fuca, Galápagos, Easter, and Balleny) and two lesser hot-spots suggested by seamount chains are near the mouth of the Gulf of California and the Eltanin fracture zone. There are four in the Indian Ocean (Amsterdam, Reunion, Afar, and Prince Edward), and six in the Atlantic (Bouvet, Tristan de Cunha-Gough, St. Helena, Ascension, Azores, and Iceland). One may argue that the lack of identifiable hot-spots on continents results from continental complexities camouflaging their presence. However, the reverse argument is no less valid; the hot-spots are mostly in the open ocean because they have pushed the crust away.

More dramatic than the location of the present hot-spots near the present rises is the evidence that the same hot-spots became active *before* the rises were formed. This evidence is best displayed in the lands bordering the Atlantic. The Jurassic volcanics in Patagonia may be regarded as the early expression of the Bouvet plume. (The even earlier Cape Volcanics in South Africa may be an expression of this plume. This interpretation depends on how Gondwanaland moved over this plume.) The flood basalts in the Parana basin and the ring dike complex of Southwest Africa may be due to the Tristan da Cunha plume. The White Mountain Magma Series in New Hampshire can be associated with the same hot-spot that produced the New England Seamount chain (probably the Azores plume). The Skaergaard and the Scottish Tertiary volcanic province are associated with the Iceland plume. I claim this line of plumes produced currents in the asthenosphere which led to the continental breakup creating the Atlantic. Likewise the Deccan Traps (Reunion plume) were symptomatic of the forthcoming Indian Ocean rifting, and, if my premise is accepted as proved, the Snake River basalts (Yellowstone plume) foretell a breakup of North America.

Gravity and Topographic Highs

Figure 4 shows a worldwide gravity map computed for spherical harmonics up to order 16 (Kaula, 1970). Isolated gravity highs are apparent over Iceland, Hawaii, and most of the other hot-spots (Galápagos is a notable exception). Such gravity highs are symptomatic of rising currents in the mantle—the less dense material in the rising plume produces a broad negative gravity anomaly; but the satellite passes closer to the excess mass in the elevated surface pushed up by this current, and the net gravity anomaly in the area over the rising current is positive. The mid-ocean rises are exceptionally shallow near the hot-spots; note particularly the 10^6 sq km areas surrounding the Iceland, Juan de Fuca, and Galápagos plumes. This regional high topography is another manifestation of the rising plume, and I shall now use the magnitude of the high topography and gravity to estimate roughly the size of the rising current.

The formulas following are adapted from derivations shown by Morgan (1965). A spherical ball of mass deficiency M, located a distance D

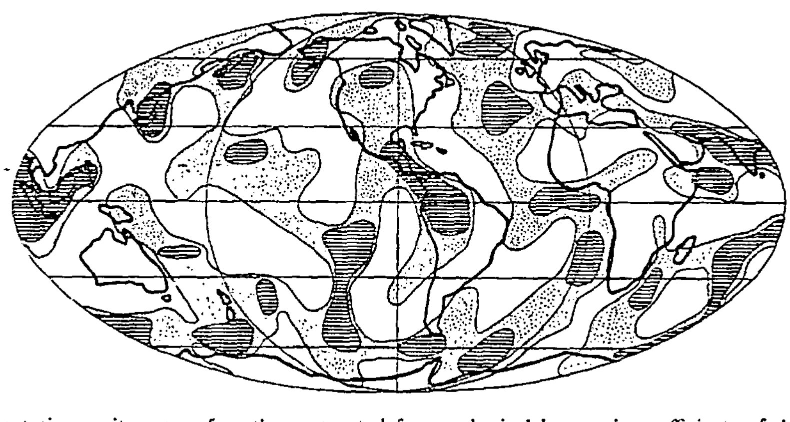

Fig. 4—Isostatic gravity map of earth constructed from spherical harmonic coefficients of degree 6 through 16. Shaded areas are regions of positive anomalies; heavier shaded areas are regions where anomalies are greater than +10 mgal. Note correlations of gravity highs with Iceland, Hawaii, and most other hot-spots (adapted from Kaula, 1970).

below the surface, is rising in a fluid of uniform viscosity. The Navier-Stokes equations were solved for a uniform half-space with rigid-plate boundary conditions for the top surface. The formulas give the normal stress σ_n, the shear stress σ_s, and the gravity anomaly δg at the rigid-plate boundary. The height the surface is pushed up is related to the normal stress by $\sigma_n = \rho g h$, and δg is the gravity anomaly produced by adding the effect of the raised surface to the gravity effect of the ball M. In these formulas, g is the gravitational field strength, G is the Newtonian Gravitational Constant, and r is the horizontal distance from the point directly above the ball.

$$\delta g = GMD\frac{(2D^2 - r^2)}{(D^2 + r^2)^{5/2}} \tag{1}$$

$$\sigma_n = \frac{Mg}{2\pi}\frac{3D^3}{(D^2 + r^2)^{5/2}} \tag{2}$$

$$\sigma_s = \frac{Mg}{2\pi}\frac{2r}{(D^2 + r^2)^{3/2}} \tag{3}$$

From Figure 4, we estimate that δg over a plume is typically +20 mgal and that the gravity anomaly falls off with distance, so that it is zero at 1,000 km away. Using equation (1) with these values, I find $D = 700$ km and $M = 0.8 \times 10^{21}$ g. If from topographic maps we estimate 1 km extra height as typical of the region near a plume, then equation (2) yields a value of $M = 3 \times 10^{21}$ g. These estimates are based on a uniform viscosity model; what effects would a more complicated viscosity pattern have? The uniform viscosity case yields the result that the total mass excess of the elevated surface equals the mass deficiency of the rising ball; *i.e.*, the stresses produced by the rising ball are balanced by the stresses produced by the increased surface load. If plumes deep in the earth are rising in pipes surrounded by a very viscous mantle (much more viscous than the asthenosphere), then much of the stress of the rising material will be distributed throughout the more rigid mantle and will not produce a local elevated surface. Thus, a larger mass deficiency may be present than that estimated from the uniform viscosity formulas. We therefore estimate $M = 10^{21} - 10^{22}$ g as a typical mass deficiency of a single plume.

Using plume dimensions to be discussed later, the density of a plume can be calculated from its total mass deficiency ($M = 3 \times 10^{21}$ g). Assuming a cylindrical shape 150 km in diameter and assuming that only the top 1,000 km of the cylinder contributes to the M estimated by the surface gravity and topography, we find $\delta\rho = -0.2$ g/cu cm, or about a 5 percent density deficiency. This density change could be produced by a migration of "400–600"-km phase change boundary. This magnitude density deficiency is ideal; if the mass and dimensions yielded a density difference 10 times larger or 10 times smaller, the result would be respectively unreasonable or uninteresting.

Formula (3) may be used to estimate the shear stresses acting on the plate bottom. On the assumption that $M = 3 \times 10^{21}$ g and $D = 700$ km, then at $r = 500$ km the shear stress is 80 bars. How the shear stress falls off with distance away from the plume is very sensitive to the exact viscosity pattern; however I shall use 100 bars as a rough estimate of the shear stress on a plate near a plume.

Stresses at Rises and Trenches

If the stresses produced by plume currents were clearly larger than the push of a rise or the pull of a trench, then the problem of finding the stresses acting on the plates would be greatly simplified. I have constructed the following mathematical model with three sources of stress: (1) stresses on plate bottoms falling off as $1/r$ away from each hot-spot, (2) a drag stress on the bottom of each plate proportional to the plate's velocity over the lower mantle and (3) stresses generated by plate-to-plate interactions of rises, trenches, and faults. The last category would have a moderately complicated set of equations predicting the stress generated by a specified closing rate at a trench or slip rate at a fault, *etc.*, but it would have Newton's Third Law of action and reaction as a simplifying feature. With this model the torques on plates with the present boundary locations can be determined, and the direction and rate of motion of each plate predicted. The parameters specifying plume size and plate-to-plate interactions then could be accurately found by adjusting them until the present observed plate motions were predicted. If the plate-to-plate interactions were smaller than the stresses produced by plumes, very elementary assumptions could be made about rises and trenches, as small errors in this specification will not be important. If the push of rises and pull of trenches are stronger than the plume-generated stresses (as appears to be the case), the equations relating stress and strain rate at boundaries must be known accurately. Thus I consider the evidence relating to the magnitude of stresses at rises and trenches, and in particular reexamine the argument that the symmetry of rises shows that the rises exert no push on the plates.

A simple calculation places an upper limit on the amount of stress that can be generated by a spreading rise. Equate the work done pushing the plates apart (the total force on a plate times the rate of displacement) with the gravitational energy available in the light material rising into the spreading area (the buoyant force times the rate of upward movement). This calculation is performed most easily in a triangular geometry with a wedge-shaped unit rising and pushing two lithospheric plates apart. The calculation is independent of the shape of the triangle and the spreading rate; it depends only on the thickness of the lithosphere and the density deficiency of the material entering at the bottom compared to the average density of the lithosphere. If it is assumed that $L = 70$ km and $\delta\rho/\rho = 3$ percent, the horizontal compressive stress in the lithosphere would be 300 bars. Any of the gravitational energy of the rising wedge that is dissipated in viscous flow will not be available to do the work required to push the plates apart: thus 300 bars is an upper limit. A 1 percent density deficiency may more accurately describe the material entering the region below the rise; in this case the horizontal compressive stress averaged over the thickness of the lithosphere would be less than 100 bars.

Focal mechanism studies of earthquakes along rise axes have shown that the lithosphere at a rise is under tension in the direction of spreading. Wyss (1970b) has concluded that this tension has a magnitude of roughly 200 bars; this result is based on studies of the seismic moments of rise earthquakes showing an "apparent stress release" of 20 bars, combined with an estimate of 10 percent for the seismic efficiency of stress release. This tension of 200 bars can be interpreted several ways. In one interpretation, we assume that some distant forces are pulling the plates apart and that the asthenosphere is rising passively to fill the void that would be created by plate separation. The lithosphere is very thin beneath the rise crest, and this "necking" of the lithosphere acts as a stress concentrator. That is, the tensional stress may be 200 bars in a region 5 km thick at the rise crest, but only 50 bars spread out over the entire 70 km thick lithosphere at some distance away from the rise. The amount of stress concentration in the thin lithosphere depends on how the stress load is distributed between the cool, strong lithosphere and the hotter, weaker asthenosphere flowing into the broad "gap" between the plates. In an alternate interpretation, the rising asthenosphere is pushing the plates apart, causing horizontal compression in the plates except in the small thin section of lithosphere at the rise which resists the separation. The average compression in the plates thus will be reduced by a factor which depends on the effective viscosity and thickness of the lithosphere at the rise crest. The dissipation in this thin section of lithosphere may form the major part of the viscous dissipation mentioned in the preceding paragraph. Therefore, the stresses at rises may be compressive or tensile, but in any event have a magnitude less than a few hundred bars.

Why are the mid-ocean rises "mid-ocean", and why is the seafloor magnetic pattern symmetrical about the rise crest? It would be easy to imagine that a rise creates new sea floor on one side only, analogous to the one-sided consumption of crust in a trench system, and yet new sea floor is created in equal amounts on the two sides of a rise. As a consequence, rise crests cannot be fixed with respect to the mantle; they must migrate over the mantle to maintain their position midway between continents. An example of such rise migration is seen for the rise boundaries that enclose Africa on three sides. As the Mid-Atlantic Rise spreads symmetrically, there is ever more sea floor between the rise crest and the African coastline. With a similar increase in the distance from Africa to the crest of the Mid-Indian Rise, the distance from the crest of the Mid-Atlantic Rise to the crest of the Mid-Indian Rise must be increasing. Thus both rises cannot be fixed with respect to the mantle—one or both must be migrating over the mantle.

The Mid-Atlantic Rise apparently is fixed to the mantle, as all the Atlantic plumes are near the present crest. Thus it is the Mid-Indian Rise that is migrating east at a rate faster than the African plate is moving northeastward. The Mid-Indian Rise has migrated over the Reunion plume—this plume was once in the Indian plate (Deccan Traps, Laccadive-Maldive island chain), but is now on the African side of the rise. Similarly, the growth of the Afar Triangle on the southwest may be regarded as the plume staying fixed while the Red Sea and Aden rifts migrate northeastward.

It has been argued that the symmetrical spreading and the migrating rise crest indicate

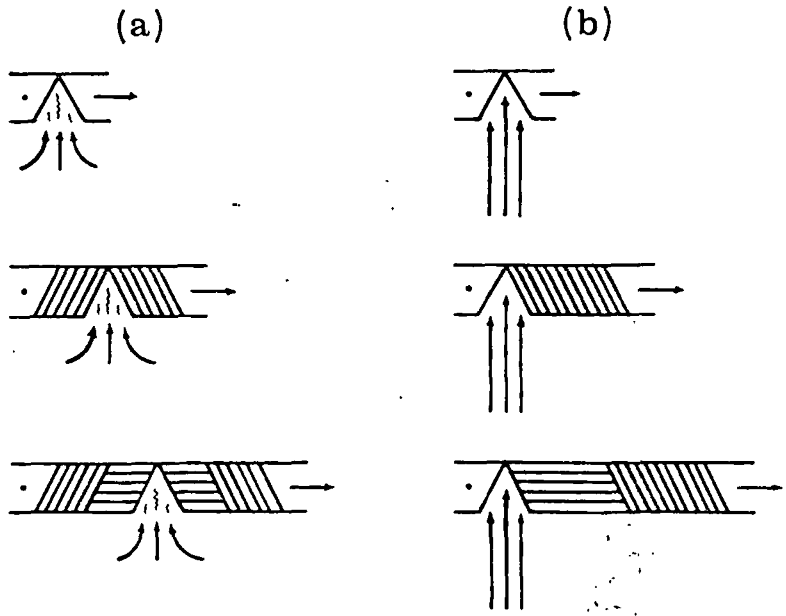

FIG. 5—In these symmetrical and asymmetric models of sea-floor spreading, left lithospheric plate is constrained to be fixed and right plate to move away at constant velocity. (a) If asthenosphere near rise fills gap made as plates move apart, symmetrical sea-floor spreading results. (b) If location of rising current is influenced strongly by conditions near bottom of asthenosphere, one-sided sea-floor spreading results.

that the rises do not drive the plates. If the two sides of a rise are moving away from a fixed rise crest with equal velocities, there is little problem; in contrast in Figure 5 I postulate a case in which the left plate has zero velocity over the mantle and the right plate is moving away with constant velocity. I do not inquire why the plates have these velocities—there may be a trench nearby on the right and another rise off left, or whatever is needed to produce the motions depicted in Figure 5. As the plates move apart at the rise crest, material from the asthenosphere rises to fill the void that would otherwise develop. The exact center of the most recently injected "dike" is hotter than any other part of the lithosphere, and because strength is extremely temperature dependent, this is where the plates will tear apart and another dike be inserted. Thus if the temperature pattern about the rise crest is symmetrical, the temperature dependence of strength will assure a symmetrical pattern of sea-floor spreading.

Any arguments about symmetrical or asymmetric spreading must thus concentrate on those factors which will make a symmetrical or asymmetric temperature pattern within a rise. It was concluded that the important factor is to have a soft asthenosphere below the plates, and that material flowing into the "gap" should be drawn from very shallow depths to avoid any kind of coupling with conditions at the bottom of the asthenosphere. It appears that passive pulling of asthenosphere into the gap or active driving of the asthenosphere upward into the gap is not related to the question of symmetry. The tensional or compressive nature of rises is related to whether or not there is a density inversion between the lithosphere and asthenosphere; the symmetrical spreading is related to the laws of heat conduction and the temperature dependence of strength. Thus it is thought that earlier conclusions of the writer (Morgan, 1971) and of Elsasser (1969) relating symmetry to passiveness are in error. A two-dimensional numerical rise model with viscosity and density varying with temperature could aid in answering this question.

A related problem concerns the existence of fracture zones. Objectors to sea-floor spreading have used the pattern of transform faults as an argument against spreading; namely, it is inconceivable that convection currents beneath the surface could have the numerous offsets of the surface pattern. The notion of a crustal plate removes the objection, as the deeper flow may have a smoother, more fluid pattern and only the "rigid" lithosphere need be broken into the irregular pattern observed at the surface. How-

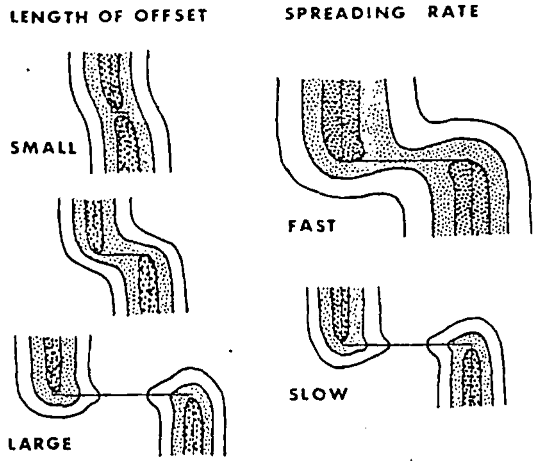

FIG. 6—Stability of fracture zone is influenced by length of offset and spreading rate. Contours are, schematically, thickness of lithosphere or, alternatively, depth to particular isotherm. If offset is very small, obliquely spreading rise may develop and transform fault would no longer exist. Length of offset of minimum stable transform fault would be dependent on spreading rate as with high spreading rates broader region of "thin lithosphere" would enclose rise (from Vogt *et al.*, 1969).

ever, the causative factors behind the maintenance of many short transform faults and the tendency for the rise to form segments perpendicular to the transform faults remain unknown.

Vogt *et al.* (1969) have noted that the minimum observed offset of transform faults is less in the Atlantic than the Pacific. They have devised the model shown in Figure 6 to account for this difference. The contours in this figure schematically represent either isotherms at a given depth (*e.g.*, 10 km), with the hottest temperature at the rise crest, or alternatively they represent the depth to a given isotherm (*e.g.*, 1,000°C). We may say arbitrarily that all material cooler than 1,000°C is lithosphere and any hotter material is asthenosphere; in this case the contours in Figure 6 show the thickness of the lithosphere. These contours are patterned after heat flow models of rises (see McKenzie, 1967) in which it was shown that the distance between the isotherm contours is directly proportional to the spreading rate of the rise. The model in Figure 6 also assumes that heat is generated on the transform faults by frictional heating. Vogt *et al.* (1969) used this diagram to illustrate that, if the offset of a transform fault is "'sufficiently small" (where "sufficiently small" is directly related to spreading rate), then the regions of thin lithosphere may merge and an oblique spreading ridge replaces the two short perpendicular segments and the transform fault. Figure 6 also indicates that an offset greater than a certain minimum length will be stable and will not be replaced by an obliquely spreading rise.

The flow pattern chosen by nature must minimize the total dissipation rate. Therefore, the friction per unit length along a transform fault must be markedly less than the tensile resistance of a rise segment, otherwise an oblique rise would be observed more commonly. Further, I conclude that there is more tensile resistance for an oblique rise than a perpendicular rise, or else the dissipation rate of "perpendicular rise and fracture zone" would not be less than "oblique rise." The larger resistance of the oblique rise must be due to increased heat loss through the longer sides resulting in a higher effective viscosity for the oblique rise. Several models to relate these factors were constructed by the writer to attempt to relate the tensile stress to the dissipation rate. However, it was realized that a constant compressive stress produced by buoyant material would have no effect on the model. The buoyant terms are determined by the rate of the upflow, which is the same in the oblique and normal cases; thus the solution of this problem could indicate much about the mechanisms at rise crests but cannot be used to place a limit on the overall tensile or compressive nature of the rises.

I conclude that we do not really know whether rises are characterized by compressive stresses aiding the plate motion or by tensile stresses acting as a brake to plate motion. Inasmuch as tensile stresses at rises would be most favorable to the plume driven model, it seems conservative to expect all rises to generate compressive stresses of about 100–300 bars.

On the question of the tensile (actively pulling down) or compressive (resisting being pulled down) character of trenches, the earthquake mechanism solutions of Isacks *et al.* (1969) show that earthquakes at 100–200 km depth indicate tensile stress along the sinking slab; the slab above is being pulled down by a density excess at or below this depth. Deeper earthquakes, from 300–700 km, show compressive stresses along the slab; the deeper part is resisting being driven deeper into the mantle by stresses generated above. From earthquake seismic moments (Wyss, 1970a), from surface gravity anomalies (Morgan, 1965), and from calculations based on assumed temperature profiles in the sinking slab (McKenzie, 1969), the stress produced by the sinking slab has been estimated at from a few hundred bars to a thousand bars. The motion of the sinking slab is resisted by the material on the slab boundaries, particularly on the boundary of the highest seismic activity, where the two lithospheric plates rub together. Wyss (1970a) found that 200-bar stresses are indicated by the shallow earthquakes at trenches, the same magnitude of stress as found at rises. There are many questions about the magnitude of the stresses at trenches. How much of the pull of the sinking slab is cancelled by the shallow friction of the two lithospheric plates? Do deep trenches pull with more tension than intermediate depth trenches, or does the deeper slab push into a very resistant media which tend to reduce the pull of the slab? Do areas of continental underthrusting, as in the Zagros and Himalayas, exert any pull at all, or are all stresses in these regions compressive, resisting the closing motions?

I add one new point to the discussions of trenches. Figure 3 (and more precisely if not more accurately, Table 1 of Morgan, in press) shows the absolute rate of each plate over the mantle, as determined by relative spreading rates plus the trajectories of hot-spots. The Nazca plate is moving eastward toward the Peru-Chile trench at about 7 cm/year, whereas the South American plate is moving westward toward this trench at about 1.5 cm/year. When we examine the velocities of plates at other trench systems, we find the following to be a general rule: both plates move toward the trench with the underthrusting plate moving about four times faster than the overriding plate. (This claim is made with great reservations, because, to establish it a much more accurate determination of absolute plate motions is needed.) However, note the special situation of the Philippine Sea. It is moving westward toward Asia and thus is not moving toward the Marianas trench. Also, the Indian plate does not have a component of motion toward the Tonga trench. It thus appears that the areas of anomalous spreading behind trenches (Karig, 1970) can be identified if the absolute motion of the plates is known. Moberly (in press) and Elsasser (1971) have given theoretical explanations of this phenomenon; they suppose the lithospheric slabs are not exclusively sliding obliquely into the mantle but also have a vertical sinking component. The sliding and sinking cause the trench axis to migrate "seaward," and the overthrusting plate must migrate toward the trench or a gap will open behind the trench.

This motion of both plates toward a trench appears to clinch the argument in favor of tensile stresses at trenches. In the following discussion I shall assume this horizontal tensile stress has a magnitude of a few hundred bars. I shall neglect the stresses produced by plates rubbing together at great faults, partly because such stresses must surely be reactive and not drive plates, and partly because the model of Vogt *et al.* (1969) on transform faults shows the dissipative stresses on faults must be much less than on rises.

Estimates of Plume Magnitudes

Estimates of stresses, heat flow, and lead isotope data are now used to estimate the sizes of plumes. A self-consistent set of relations is found if each plume is 150 km in diameter, with an average upward velocity of 2 m/year; 20 such plumes would bring up a total volume of 500 cu km/year. Other values used in this model are a lithosphere 70 km thick and an asthenosphere 200 km thick, with an average viscosity of 3×10^{21} poise. No exactitude should be placed on any of these numbers, as it is the overall effects of the parameters and not the precise value of any one which is significant in the ensuing discussion. My purpose in the use of this model is to show that plumes can provide the stresses needed to move the surface plates and have important implications in the interpretation of heat flow and age of the mantle, and at the same time do not violate any of the "known" values of the earth.

Using equation (3) I found that the shear stress 500 km away from a plume was about 100 bars. I obtain this same magnitude of stress in the plume and thin asthenosphere model specified above. The average velocity $\bar{v}_a$ of the asthenosphere at a distance R from the plume is related to the upward velocity of the plume v_p by $2\pi R D \bar{v}_a = \pi d^2 v_p/4$, where D and d are the thickness of the asthenosphere and diameter of the plume respectively. Using the model values given above, I calculate that the average asthenosphere velocity 500 km from the plume is 5 cm/year. The asthenosphere flow is channeled between the rigid upper plate and the lower mantle (assumed to be slightly more viscous), and the velocity profile in this channel takes the well-known parabolic shape. (If the upper plate is moving, a linear velocity profile will be superposed on this pattern.) The stress at the top boundary (σ) is related to the average velocity in the asthenosphere ($\bar{v}_a$) by $\sigma = 6\eta\bar{v}_a/D$, giving $\sigma = 150$ bars at 500 km radius.

Integration of the effect of this stress on the bottom of a plate determines the total force one plume can exert on a plate. I assume that a rise axis passes right over a plume and integrate the component of force directed away from the rise. I choose as limits of my integration 75 km (the radial velocity is zero directly above a plume and increases to a maximum value a plume radius away) and 1,500 km (roughly half the distance between plumes). This integration yields $F = 1 \times 10^{24}$ dynes for the total force exerted on a plate.

I then compare this magnitude with the stress created by a plate moving over the asthenosphere. With the properties of the asthenosphere given above, a plate moving 3 cm/year over the mantle creates a shear stress of 15 bars. If we assume the plate is a square about 5,000 km on a side, then the total drag on a plate is $F = 4 \times 10^{24}$ dynes. Another comparison is with the stress generated by plate-to-plate interactions. I assume a stress of 100 bars on a plate 70 km thick along a boundary 10,000 km long; the total force acting on the plate is $F = 7 \times 10^{24}$ dynes. The plume generated stress is smaller than the other two, but it could be increased by changing the values of the viscosity and thickness of the asthenosphere, or by increasing the flow up a pipe. My conclusion is that, given the uncertainty in these factors, all three mechanisms (plume, drag, and plate-to-plate) should be considered in a model of plate motion.

The key estimate is the volume of flow up the pipes. There is a clear-cut way to obtain a lower limit on the rate if the plumes are driving the plates. From roughly 40,000 km of rise axis with an average (half) spreading rate of 3 cm/year, I can determine how much new crust is generated each year. An accurate summing of the spreading rates along each rise yields 2.5 sq km/year for this rate (Deffeyes, 1970, p. 214). If this is multiplied by the thickness of the lithosphere, it is apparent that lithosphere is being generated at a rate of 170 cu km/year, and of course destroyed at an equal rate at the trenches. Suppose I had concluded that the total volume brought up by the plumes was only 10 cu km/year, or some other equally small number. Then asthenosphere currents of total flux 10 cu km/year would be spreading horizontally away from several points on or near rise crests, whereas 170 cu km/year would be flowing toward the rises as a counterflow to the lithosphere motion. The net stresses on plate bottoms would be such as to close up the plates. Thus the total volume emanating from the plumes must be several times larger than 170 cu km/year. The value of 500 cu km/year specified in the model is three times this; one may like a larger multiple, but this value does satisfy the lower bound and fits the lead isotope criteria to be discussed subsequently. In this light, even small plume flow aids the shallow convection in a way not noted in the preceding paragraph, where I found that a plate spreading at 3 cm/year created a viscous drag of 15 bars. If I include in this calculation the assumption that the 200-km-thick asthenosphere must have a net flow to counter the mass transport of the lithosphere, then the stress on the bottom of the plate is 50 bars instead of 15 bars.

These numbers have interesting consequences for the interpretation of heat flow data. Suppose 500 cu km/year (ϕ) brought up by the plumes is on the average 300°C (ΔT) hotter than the nonplume mantle at the same depth. Then using $C_p = 0.25$ cal/g°C, I find that the total upward heat available from the plumes is $Q = \rho\, C_p\, \Delta T \phi = 1.5 \times 10^{20}$ ergs/sec. This number is half the total heat flow of the earth. Here is a mechanism for concentrating all the deep mantle's heat production into predominantly oceanic regions. The correct interpretation of this may show why the oceanic and continental heat flow averages are so nearly equal. There is another surprise in these numbers: the return flow of the plumes, involving the slow sinking of the entire mantle, is at a rate of 0.1 cm/year and downward convection at 0.1 cm/year can dominate over conduction or radiation mechanisms of heat transport. For example, if I assume a temperature (T) of 1,500°K at the base of the asthenosphere, then the downward flux of heat by convection is $q = \rho\, C_p\, T\, v = 3.7$ μcal/sq cm/sec. Thus, heat could leave the lower mantle only at the plumes. How much heat recycles compared with the amount of heat lost at the upper surface is a measure of the efficiency of the heat engine—thus the earth could be regarded as being a moderately efficient heat engine.

The isotopic composition of lead from Tristan da Cunha and St. Helena has been discussed by Oversby and Gast (1970) together with earlier results from Ascension and Gough. They showed that the lead data cannot be interpreted with a simple one-stage growth model, i.e., a mixing of lead and uranium isotopes 4.5 b.y. ago, when the mantle was formed, with no separation or mixing since (except perhaps in the last few million years as the rocks were brought to the surface). Instead, they found a two-stage growth history with lead events at 4.5 b.y. and 1.8 b.y. (plus possible changes in the last few million years). That is, they interpreted their data to show that there was an homogenization and then separation of the lead and uranium isotopes 1.8 b.y. ago in the material that now makes up these islands, and that no further mixing or separation occurred until very recently.

I incorporate this observation into the plume model as follows—I assume that the rocks which now make up these islands were last near the earth's surface 1.8 b.y. ago; *i.e.*, that 1.8 b.y. is the cycle time required for a particle to sink slowly in the mantle and then to rise in a plume back to the asthenosphere (or in this case, for part of the plume to reach the surface). The rate of upwelling of 500 cu km/year fits these data: the total volume of the mantle, 1×10^{12} cu km, divided by this rate gives 2 b.y. If the lead isotope data are interpreted in this manner, it makes two restrictions on the plume model. First, this is evidence that the entire mantle is involved in the overturn—that the plumes extend all the way down to the core-mantle boundary. Second, the rate of upwelling may be estimated most accurately by knowing the period of the mantle overturn—a very straightforward estimate if lead isotope data from other plumes far from the South Atlantic also show the 1.8 b.y. lead event.

In conclusion, the mid-ocean position of most of the plumes and the land evidence of plume activity prior to continental breakup suggest that the plumes produce the stresses which drive the plates apart. An order of magnitude estimate shows that stresses produced by plume currents are comparable to other stresses. The model implies that the entire mantle overturns once each 2 b.y., a conclusion which would require a new interpretation of heat flow and chemical evolution problems.

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DAVID ELLIOTT *Department of Earth and Planetary Sciences, The Johns Hopkins University, Baltimore, Maryland 21218*

Deformation Paths in Structural Geology

ABSTRACT

The evolution of folds and the origin of mineral orientations and schistosity are analytically related to the deformation paths or histories in deforming rocks. The several different finite and incremental methods of representing deformation paths are connected by matrix algebra, and graphical presentation in natural strain space is found to be particularly clear. New methods permit calculation of various components of deformation paths from boudinage, pressure shadows, and inclusion trails in naturally deformed metamorphic rocks. In coaxially accumulating deformation paths the principal strains remain parallel to the same material lines in the rock and have been referred to inappropriately as irrotational. Straight inclusion trails in metamorphic minerals may be produced by synkinematic grain growth during a coaxially accumulating path and do not necessarily indicate postkinematic grain growth. Synkinematic inclusion trails can be produced by growth of the grain through its pressure shadow rather than matrix schistosity, and the quartz protected from recrystallization by being included in snowball garnets may sometimes form this way.

INTRODUCTION

A given finite deformation can be reached by an infinite number of different paths only one of which actually occurred. A deformation history or path indicates the states through which a deformation progresses en route to its final value.

Familiar fluids such as water, air, tar, and plasticine preserve none of the history of the deformations they have been through. Metamorphic rocks are different, as their texture may be utterly dominated by the effects of deformation history.

Understanding the history of deformation plays a crucial role in structural geology, yet the concept was not discussed in any definite way until Flinn (1962). To work out the development of a particular fold, for example, one has to know either the deformation history for a sufficient number of points, or the history of velocity boundary conditions and there is simply no way to get velocity boundary conditions *a priori*.

This paper outlines new ways of defining deformation history in naturally deformed rocks and critically reviews methods in current use. In order to accomplish this in a unified way, it is necessary first to extend and review the mathematical formulation of deformation paths and then to relate this in turn to folding, boudinage, pressure shadows, inclusion trails in minerals, and the formation of schistosity or cleavage. This paper, then, is an attempt to interpret deformation paths directly from observations in the field and in thin section; two other different approaches can also be used to determine deformation history. One of these approaches is to assume all the material properties and boundary conditions as given, and then produce the folds by performing computer experiments (for example, Chapple, 1969; Dieterich, 1969). The results may then be compared to the natural structures. Another possible way is to seek special situations, such as ductile shear zones where the deformation is restricted to simple shear paths by the undeformed rock on both sides (Ramsay and Graham, 1970). All of these different methods have important common ground and are complementary.

DEFORMATION PATHS AND COORDINATE FRAMES

The next six sections of this paper deal with the mathematical framework behind the concept of deformation paths. Unfortunately, not much of this material is available in the geological literature, and it is only partly available and widely scattered in the mechanics literature. Analytical relations are established between the various ways of defining and

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describing deformation history and several new terms are introduced. Much of the treatment is two-dimensional for ease of presentation, but extension into three dimensions is straightforward. First we must examine the matrix algebra in three different types of coordinate frames.

Fixed External Frame

Externally fixed coordinate axes can be visualized as east-west and north-south grid lines scratched on a sheet of glass. We are interested in what is happening in a particular mass of rock, so as the rock deforms beneath the glass the coordinate origin is kept fixed over the same material point but the coordinate frame maintains the same east-west and north-south grid lines and scale. Let the mass of rock be small enough that the strains remain homogeneous. A homogeneous but otherwise perfectly general rotational strain will be called deformation, and can always be represented by a matrix **D**.

The deformation path is the history of **D** and can be analyzed as two separate components: the pure strain history and the rotation history. This is because any deformation matrix can be factored by left polar decomposition (Elliott, 1970a):

$$\mathbf{D} = \mathbf{T}\mathbf{R}\,. \tag{1}$$

The simplest description of the symmetric left stretch matrix **T** is when it is in diagonal form **Td** (all diagonal forms are distinguished by d).

$$\mathbf{T} = \boldsymbol{\theta}\,\mathbf{Td}\,\boldsymbol{\theta}^{-1}$$
$$\mathbf{Td} = \begin{bmatrix} T_I & 0 \\ 0 & T_{II} \end{bmatrix} \tag{2}$$

T_I and T_{II} are the long and short semiaxes of the strain ellipse.

The orthogonal matrix θ in equation (2) is an orientation matrix defining the angle θ of the long axis of the strain ellipse to the externally fixed x coordinate (Fig. 1).

From (1) and (2):

$$\mathbf{D} = \boldsymbol{\theta}\,\mathbf{Td}\,\boldsymbol{\theta}^{-1}\,\mathbf{R}\,. \tag{3}$$

When a rock deforms, an original embedded sphere changes progressively through a sequence of continuously varying finite strain ellipses, and each instant is characterized by a unique ellipse. We can see that the left stretch

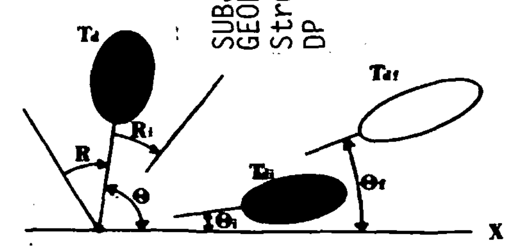

Figure 1. Original unit circle rotates R and then strains Td at an angle θ to the X axis, then rotates an increment Ri and is strained an increment Tdi at an angle θi. The total final strain ellipse Tdf is oriented θf. This is a geometric interpretation of deformation in a fixed external frame of equation (6b).

matrix **T** is concerned with this finite strain ellipse. The pure strain history is the history of the magnitudes of the axes of the finite strain ellipses T_I, T_{II}, and also the history of the orientation θ of the finite strain ellipse.

The orthogonal rotation matrix **R** is concerned with the rotation through an angle R of a material line in the undeformed rock into parallelism with the long axis of the strain ellipse in the deformed rock. The history of the angle R is the rotation history.

Let us specify at the outset that the deformation path must be smooth and continuous during one period of deformation. No sudden reversals in principal direction or jumps in amount of finite strain occur. Different deformation periods are probably separated by changes in physical conditions, so each period should be described by its own unique deformation path.

Deformation history is frequently composed of information on successive portions, or increments, of the total deformation. After a general initial deformation **D** allow a subsequent deformation increment **Di** (incremental matrices are distinguished by i). In general, this increment **Di** can be any size.

The total final deformation **Df** (final or total deformation matrices indicated by f) from the original geometry into the current shape is

$$\mathbf{Df} = \mathbf{Di}\,\mathbf{D}\,. \tag{4}$$

Note that the deformation **Di** premultiplies the earlier deformation **D**, following the standard convention that matrices representing later events accumulate toward the left (Truesdell and Toupin, 1960, p. 246; Elliott, 1970a).

Any given finite deformation can be regarded as a product of incremental deformations successively compounded according to equation (4). Each new increment modifies the total deformation. Assuming that the rock was undeformed at the beginning, then a sequence of superimposed incremental deformations (indicated by subscripts) is

$$\mathbf{Df} = \mathbf{Di_n} \cdots \mathbf{Di_3}\, \mathbf{Di_2}\, \mathbf{Di_1}\,, \tag{5}$$

and $\mathbf{Df}$ is the total finite deformation at time t.

Equation (5) is the finite deformation accumulation equation for a general path and represents the entire history of deformation increments. One must multiply all the increments and there is no shortcut except for special deformation paths. In summary, deformation history is completely known if one has either the history of the total finite deformation $\mathbf{Df}$ or the sequence of incremental deformations $\mathbf{Di}$.

Equation (4) can be expanded by left polar decomposition, from (1) and (2):

$$\mathbf{Df} = \mathbf{Tf}\,\mathbf{Rf} = \mathbf{Ti}\,\mathbf{Ri}\,\mathbf{T}\,\mathbf{R}\,; \tag{6a}$$

$$\theta\mathbf{f}\,\mathbf{Tdf}\,\theta\mathbf{f}^{-1}\,\mathbf{Rf} = \theta\mathbf{i}\,\mathbf{Tdi}\,\theta\mathbf{i}^{-1}\,\mathbf{Ri}\,\theta\,\mathbf{Td}\,\theta^{-1}\,\mathbf{R}\,. \tag{6b}$$

Equation (5) can be expanded the same way:

$$\mathbf{Df} = \theta\mathbf{i_n}\,\mathbf{Tdi_n}\,\theta\mathbf{i_n}^{-1}\,\mathbf{Ri_n} \cdots \theta\mathbf{i_2}\,\mathbf{Tdi_2}\,\theta\mathbf{i_2}^{-1}\,\mathbf{Ri_2}\,\theta\mathbf{i_1}\,\mathbf{Tdi_1}\,\theta\mathbf{i_1}^{-1}\,\mathbf{Ri_1}\,. \tag{7}$$

With time the incremental strain ellipse is continuously changing its axial lengths, orientation, and rotation. There is a geometric interpretation of this incremental deformation history (Fig. 1).

Clearly, the incremental deformation history can be determined if the finite deformation history is available, and vice versa. Now

$$\mathbf{Di} = \mathbf{Ti}\,\mathbf{Ri} = \mathbf{Df}\,\mathbf{D}^{-1}\,,$$

Multiplying each side of the equation by its transpose:

$$(\mathbf{Ti})^2 = \mathbf{Df}\,\mathbf{D}^{-1}\,(\mathbf{D}^{-1})^T\,\mathbf{Df}^T\,.$$

Having $\mathbf{Ti}$ from this equation we can now calculate $\mathbf{Ri}$:

$$\mathbf{Ri} = \mathbf{Ti}^{-1}\,\mathbf{Df}\,\mathbf{D}^{-1}\,.$$

Rather surprisingly, to calculate the pure strain $\mathbf{Ti}$ (or the rotation $\mathbf{Ri}$) we must not only have the finite pure strains $\mathbf{T}$ and $\mathbf{Tf}$, but also the rotations $\mathbf{R}$ and $\mathbf{Rf}$.

Frame of Incremental Principal Directions

There is another way of interpreting incremental deformation history. Adjacent orthogonal matrices can always be compounded into one orthogonal matrix, so defining

$$\mathbf{R_1}' = \theta_1^{-1}\,\mathbf{Ri_1};\ \mathbf{R_2}' = \theta_2^{-1}\,\mathbf{Ri_2}\,\theta_2;\ \mathbf{R_3}' = \theta_3^{-1}\,\mathbf{Ri_3}\,\theta_2, \text{ etc.}$$

Substituting into (7), the finite deformation accumulation equation (5) now becomes

$$\mathbf{Df} = \mathbf{Tdi}\,\mathbf{R_n}' \cdots \mathbf{Tdi_3}\,\mathbf{R_3}'\,\mathbf{Tdi_2}\,\mathbf{R_2}'\,\mathbf{Tdi_1}\,\mathbf{R_1}'\,. \tag{8}$$

A perfectly general finite deformation path may be interpreted as a sequence of incremental rotations interspersed with incremental strains whose strain ellipses are all oriented parallel to the coordinate frame (Fig. 2).

Rigidly Moving Frame

There is often no particularly useful way of choosing a coordinate frame which maintains constant spatial orientation throughout the deformation, such as the east-west x coordinate used up to now. In many cases the equations show more of the physics with a rigidly moving coordinate frame undergoing all the solid body motions of the rock. Choose the rigidly moving x coordinate as the one which is eventually oriented in the east-west direction in the final deformation state, and defining

$$\phi\mathbf{f} = \theta\mathbf{f};\ \phi\mathbf{i} = \theta\mathbf{i};\ \phi = \mathbf{Ri}\,\theta;\ \mathbf{Rm} = \mathbf{Rf}\,\mathbf{R}^{-1}\,\mathbf{Ri}^{-1};$$

substitute these equations into equation (6b) and after rearranging

$$\phi\mathbf{f}\,\mathbf{Tdf}\,\phi\mathbf{f}^{-1}\,\mathbf{Rm} = \phi\mathbf{i}\,\mathbf{Tdi}\,\phi\mathbf{i}^{-1}\,\phi\,\mathbf{Td}\,\phi^{-1}\,,$$

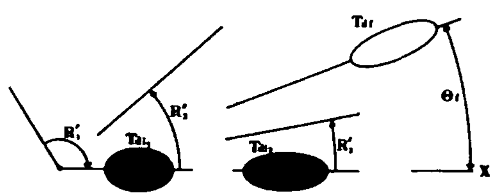

Figure 2. Deformation in a frame of incremental principal directions. The progress of the deformation is geometrically interpreted by reading the sequence of matrices toward the left in Df = Tdf R'f = R₃' Tdi₁ R₂' Tdi₁ R₁' (Equation 8).

the left and right sides of this equation define

$$\mathbf{Dfm} = \mathbf{Tim}\,\mathbf{Tm}\,;$$

the deformation accumulation equation now becomes

$$\mathbf{Dfm} = \mathbf{Tim_n} \cdots \mathbf{Tim_3}\,\mathbf{Tim_2}\,\mathbf{Tim_1}\,.$$

The horizontal sheet of glass is now keeping the grid origin over the same material point and turning from one direction to another with the solid body rotations of the rock (Fig. 3). Deformation matrices calculated in such a frame clearly cannot "see" incremental rotations. In a rigidly moving coordinate frame all the solid body motions are removed and greater attention is focused on changes of shape and size.

The calculation of the pure strain increment is a little simpler than before

$$\mathbf{Tim} = \mathbf{Tfm}\,\mathbf{Rfm}\,\mathbf{Tm}^{-1}\,.$$

Yet another type of coordinate frame is one in which the x axis is kept always parallel to the same material line throughout the deformation history. A special case where the chosen material line is also a simple shear direction has received considerable attention in structural geology and is not considered further here (*see* Ramsay and Graham, 1970).

There is an infinite choice of coordinate frames which could be used to describe deformation history and from a mathematical point of view it does not matter which is used, but some coordinate frames lend themselves to a simpler physical interpretation of the processes.

NATURAL TIME

Even if the increments making up a deformation path are known, it is still not possible to plot the path unless the order in which successive increments occur is known. History implies time ordering and some sort of time unit must be considered.

The time unit could be true time but it is difficult to imagine geological information which could provide a record of true time during deformation under metamorphic conditions. Order-of-magnitude estimates of real-time deformation rates, such as the often quoted $\dot{e} \simeq 10^{-12}$ to 10^{-16} per sec, are interesting but cannot be defined around individual folds. In natural folds deformation rates must be based on suitable continuously increasing parameters other than true time.

Suppose some scalar parameter t ("natural" time) increases continuously during deformation. The total deformation at a particular instant has a specific value of t, and increasing values of t unambiguously order the progress of the deformation. The continuously increasing arc length of growing quartz fibers is used later in this article as an appropriate time-like parameter t, and if

$$t = \frac{\text{current arc length}}{\text{typical arc length}}\,, \tag{9}$$

then the natural time t is used in a dimensionless sense. The growing radius of a metamorphic mineral, a steadily increasing strain axial ratio near a boundary of a region being studied, or the increasing dip of a bed could also be used as dimensionless time-like parameters. Indeed, any monotonically increasing quantity which can be correlated with the progress of the deformation provides a built-in clock. The parameter t takes on a broader meaning than is usual in geology, but this broader interpretation is widespread in mechanics.

DIFFERENTIAL DEFORMATION AND RATES

The initiation or development of a structure depends upon what is happening at an instant. The concern is with the rate of change of finite deformation. The derivative with respect to natural time t of each of the elements of the finite deformation matrix in an externally fixed oriented frame defines the deformation rate matrix $\dot{\mathbf{D}}$ at that instant. For example:

$$\dot{D}_{11} = \lim_{\Delta t \to 0} \frac{\Delta D_{11}}{\Delta t} = \frac{d}{dt} D_{11}\,,$$

or as a matrix

$$\dot{\mathbf{D}} = \frac{d}{dt}\mathbf{D}\,.$$

Consequently, the finite deformation up to a natural time t can be determined by integrating in a common externally fixed frame each of the elements of the deformation rate matrix $\dot{\mathbf{D}}$,

$$\mathbf{Df} = \int_0^t \dot{\mathbf{D}}\,dt\,.$$

This integral only applies to a region of homogeneous deformation.

The deformation rate matrix $\dot{\mathbf{D}}$ is most suitably factored by Euler-Cauchy-Stokes decomposition (Truesdell and Toupin, 1960, p. 362).

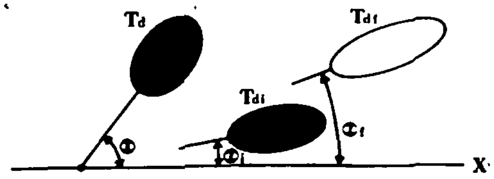

Figure 3. Deformation in a rigidly moving frame. Progress of the deformation is geometrically interpreted by reading the sequence of matrices toward the left in $Dfe = \phi f\, Tdf\, \phi^{-1} = \phi i\, Tdi\, \phi i^{-1}\, \phi\, Td\, \phi^{-1}$.

$$\dot{D} = \dot{e} + \dot{W} \text{ where } 2\dot{e} = \dot{D} + \dot{D}^T,$$

and

$$2\dot{W} = \dot{D} - \dot{D}^T.$$

The strain rate matrix $\dot{e}$ is symmetric, and any principal value, say $\dot{e}_I$, is defined as

$$\dot{e}_I = \frac{1}{\ell}\frac{d\ell}{dt} = \frac{d}{dt}\ln\left(\frac{\ell}{\ell o}\right),$$

when ln is the natural logarithm, ℓ and ℓo are original and final lengths.

The strain-rate matrix is represented by a strain-rate ellipse whose longest semi-axis $(1 + \dot{e}_I)$ has an orientation angle θr to the externally fixed coordinate frame.

The spin $\dot{W}$ is the rate of rotation of material lines parallel to the axes of the strain-rate ellipse at a particular instant t. The spin matrix is always skew-symmetric.

If a rigidly moving coordinate frame is used, then similarly

$$\dot{D}m = \dot{e}m = \frac{d}{dt}Dfm.$$

The strain rate ellipse is oriented at an angle ϕr to this coordinate frame.

Paths of constant deformation rate $\dot{D}$ when the parameter t is true time are particularly important for experimental determination of material properties because stress strain-rate relations are then largely free of the effects of previous history or memory. Such paths are known in mechanics as constant stretch history, viscometric, or substantially stagnant flows. The constant strain-rate test widely used in experimental structural geology can only qualify insofar as one accepts that

$$\frac{d\ell}{\ell o} \simeq \frac{d\ell}{\ell}$$

is sufficiently accurate for the strain considered.

A deformation increment can have any size, as it is simply the total deformation which has occurred over some interval. If the increment Di is relatively small compared with the total final deformation **Df**, then a good approximation to the true deformation increment **Di** is the differential deformation increment **dD**:

$$\mathbf{Df} = \mathbf{dD} + \mathbf{D}$$

where **D** and **Df** are now the finite deformations at the beginning and end of a short interval. This equation for a differential increment is simpler than for a finite increment and is widely used. We can now analyze the deformation **Df** into what has happened (**D**) and what is now happening (**dD**). The matrix algebra of **dD** is similar to $\dot{D}$, and

$$\mathbf{dD} = \mathbf{e} + \mathbf{W}$$

where the Cauchy strain matrix **e** has as principal values the ratio of change in length $\Delta\ell$ over the current length ℓ during the differential deformation; for example

$$e_I = \frac{\Delta\ell}{\ell}.$$

"Infinitesimal" deformation is often used in the sense of a differential deformation (for example, Flinn, 1962, 1965; Ramsay, 1967).

The small masses of material making up a structure each have deformation rates or differential deformations related to the velocity field or small displacement field at that instant; an aspect thoroughly reviewed by Hsu (1966) in mechanics and Howard (1968) and Ramsay and Graham (1970) for structural geology.

NATURAL STRAIN SPACE

A particularly useful measure for longitudinal strain of a material line is natural strain

$$H = \ln\left(\frac{\ell}{\ell o}\right).$$

The Hencky strain matrix **H** with respect to the deformed state has the same principal directions as the left-stretch matrix (equation 3), and its principal values, H_I, H_{II}, and H_{III} in three dimensions, are defined by the last equation. This matrix will only be used in a diagonal form in this paper:

$$Td = e^H \text{ or } \ln Td = H.$$

The rate of change of the natural strain H is precisely the longitudinal strain rate $\dot{e}$:

$$\frac{dH}{dt} = \frac{d}{dt}\ln\left(\frac{\ell}{\ell o}\right) = \dot{e}.$$

For small strains the Cauchy strain is approximately the same as natural strain:

$$H = \ln\frac{l}{lo} \simeq \frac{l - lo}{lo} = e.$$

The history of the magnitude of strain is most clearly interpreted on a suitable graphical plot. In three dimensions plot the finite strain at an instant by the three principal natural strains H_I, H_{II}, H_{III} as cartesian components of a strain vector **h** (Fig. 4). This representation will be called natural strain space.

Volume change is measured by mean natural strain **hm** which is the projection of the strain vector onto the space diagonal. The shape of the strain ellipsoid measures the deviatoric component of the strain matrix and is the projection **O'P** of the strain vector onto a plane normal to the space diagonal and called the deviatoric plane (Fig. 4). All finite strain ellipsoids with given axial ratios have the same shape and lie on the same straight line in natural strain space.

The strain vector **h** always has its tail at the origin of the natural strain space and its tip at the current value of the finite strain. As t advances, the deformation progresses, and the tip of the strain vector traces out a curved line in three dimensions as a graph of the strain history (Fig. 5).

Principal stresses and strain rates can be plotted in the same way as finite strain, and these types of representation are all examples of Haigh-Westergaad Space (for stress references *see* Hill, 1950, p. 17). Haigh-Westergaad space is an outstanding graphic representation of second order tensors and is as useful as the Mohr Diagram. Algebraic manipulations involving invariants lose their fussiness and reduce to simple geometry. Observational and experimental results are clear and can easily be transformed into various invariant formulations.

One of the attractive features of the Hencky strain matrix and natural strain space is its close analogy with stress. Mean natural strain measures volume changes in the same way that mean stress measures hydrostatic pressure. Deviatoric natural strain and stress both measure shear and distortion. This correspondence does not hold with any other common strain measure.

In structural geology only Hossack (1968) has used natural strain space to date, and most workers use a more complex graph due to Flinn (1962) and originally proposed by Zing (1935) for use in sedimentology.

Information on the strain path in structural geology is often restricted to a plane section containing two principal directions, so only the plane in natural strain space containing the appropriate principal values is used; often this is the $H_I H_{III}$ plane. The two-dimensional deformation path is the projection of the three-dimensional path (Fig. 5). Nothing is implied about the third strain in this type of plot. Information on variation of the third strain

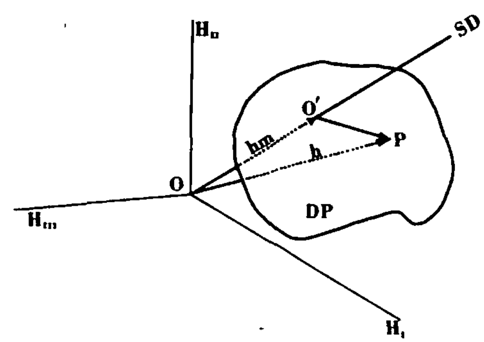

Figure 4. The space diagonal SD is the line at equal angles from the three coordinate axes H_I, H_{II}, H_{III}. Projection of the strain vector h onto the space diagonal is the mean natural strain hm. The deviatoric plane DP is normal to the space diagonal and projection of h onto the deviatoric plane gives the deviatoric strain vector O'P.

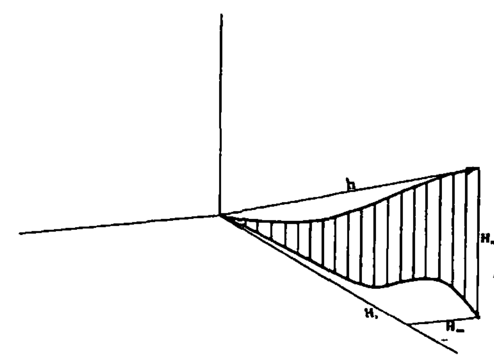

Figure 5. The current strain vector h has principal values H_I, H_{II}, H_{III}. The short axis of the strain ellipsoid has been compressed throughout the strain path so that H_{III} is negative. Note that the strain path observed in the $H_I H_{III}$ plane can be derived by projection from the three-dimensional path.

can be determined either from differently oriented plane sections or from the volumetric strain history if known.

The two-dimensional deformation path of a particular material plane which has not remained a principal plane can be shown in a similar way.

PRINCIPAL STRESS DIRECTIONS

A widely used technique attempts to identify principal stress directions at a particular instant from features formed inside a crystal at this instant, such as twins, kinks, and deformation lamellae (*see* Carter and Raleigh, 1969). More deformation **Di** generally follows and unless the deformation **Di** is known it is *impossible* to determine what the orientation of the principal stress direction was at the time the particular stress state was active, so any attempt to deduce principal stress directions with this method must always make some assumption about the deformation path.

Further, individual crystals can start and stop deforming throughout the history, and there is no way of distinguishing when a particular lamella or twin was formed. The fabric diagrams composed from a large number of deformed crystals record some complex cumulative effects with no necessary or unique relation to any stress or strain principal directions (Dieterich and Carter, 1969).

However, if the deforming rock is essentially isotropic, the orientation ϕr of a principal strain rate is the same as the principal stress direction at that instant. Consequently, if the curve of ϕr versus t can be found, the history of principal stress directions is known. Some methods will be discussed later in this paper.

LINEARITY AND COAXIALITY OF PATHS

If the same material lines remain axes of all the finite and strain rate ellipsoids during an interval, then this portion of the deformation path is defined as *coaxial* (Hsu, 1966). The material line parallel to a strain ellipse long axis might at a later time become a short axis, but the change would be smooth, and one of the strain ellipses in between would have to be a circle.

Coaxially accumulating deformation paths can sometimes be recognized by inspection; several striking examples in blueschist grade rocks from the Cordillera Coast Ranges are illustrated by Misch (1969, Figs. 1, 4, 5, 10; 1970). These crystals show several successive pull-aparts, each separated by optically distinctive grain growth and fracture filling, and if all the differently oriented crystals in the thin section show similar features, then throughout the increasing deformation the same material lines always remained parallel to the principal strains.

The most common type of coaxial path is one in which the finite strain ellipsoid short axis has always been a shortening material line. In such a path it would not be possible to distinguish whether cleavage formed normal to the maximum compressive stress or to the short axis of the finite strain ellipsoid, since they would remain parallel. A noncoaxial path is a better place to investigate the origins of cleavage, both experimentally and in the field.

In an externally fixed coordinate frame the orientation of the accumulating finite strains can vary even though successive strain increments are coaxial, as a result of rotation **Ri**:

$$\mathbf{Ri} = \boldsymbol{\theta} \mathbf{f} \boldsymbol{\theta}^{-1}.$$

The rotation accumulation equation for several increments is simply the product of the incremented rotation

$$\mathbf{Rf} = \mathbf{Ri_3}\,\mathbf{Ri_2}\,\mathbf{Ri_1}.$$

Coaxially accumulating paths have usually been referred to as "irrotational." This is an inappropriate description because (as many have realized) material lines acting as axes for all the finite strain and strain-rate ellipses can still be rotating in an externally fixed coordinate frame by the last two equations.

The calculation of a final pure strain after an increment is

$$\mathbf{Tdf} = \mathbf{Tdi}\,\mathbf{Td},$$

using the Hencky strain matrix gives an even simpler version

$$\mathbf{Hf} = \mathbf{Hi} + \mathbf{H}.$$

We shall now see that the special nature of coaxial paths gives rise to an important graphical visualization, some aspects of which have also been discussed from the point of view of mechanics by Hsu (1966). The equation above can be re-expressed using the principal strains as cartesian components of a final strain vector **hf** in natural strain space (Fig. 6):

$$\mathbf{hi} = \mathbf{hf} - \mathbf{h}.$$

Therefore, finite pure strain increments have a

geometric interpretation as chords on the strain path (Fig. 6).

The chords begin and end at true points on the deformation path, but as the increments get larger there is increasing uncertainty about the curve between the end points.

The strain accumulation equation for a coaxial path is

$$\mathbf{Tdf} = \mathbf{Tdi_n} \cdots \mathbf{Tdi_3}\,\mathbf{Tdi_2}\,\mathbf{Tdi_1},$$

or, in natural strain

$$\mathbf{hf} = \mathbf{hi_n} + \cdots \mathbf{hi_3} + \mathbf{hi_2} + \mathbf{hi_1},$$

so that coaxially accumulating strain increments may be computed by vector addition in natural strain space. The simplicity of coaxially accumulating deformation paths leads to particularly clear representation.

Since $\mathbf{h} = f(t)$, one can take the derivative of the vector function with respect to the scalar variable t.

$$\frac{d\mathbf{h}}{dt} = \dot{\mathbf{e}} = \lim_{\Delta t \to 0} \frac{\Delta \mathbf{h}}{\Delta t}.$$

The strain-rate vector $\dot{\mathbf{e}}$ in natural strain space is a vector tangential to the path, but to determine the magnitude of the strain rate vector requires the values of the principal strain rates.

$$/\dot{\mathbf{e}}/ = (\dot{e}_{I}^{2} + \dot{e}_{III}^{2})^{\frac{1}{2}}.$$

In a coaxial path the maximum principal strain rate $\dot{e}_I$ is the rate of change of the long axis of the finite strain ellipse.

$$\dot{e}_I = \frac{d\,H_I}{dt}:$$

similarly,

$$\dot{e}_{III} = \frac{d\,H_{III}}{dt}.$$

The angle of slope γ of the strain-rate vector $\dot{\mathbf{e}}$ at a particular point on the path is the ratio of the principal strain rates (Fig. 6).

$$\tan\gamma = \frac{d\,H_{III}}{d\,H_I} = \frac{d\,H_{III}}{dt} \cdot \frac{dt}{d\,H_I} = \frac{\dot{e}_{III}}{\dot{e}_I}.$$

To determine the magnitude of the principal strain rates one must know the values of t along the path, but the ratio of the principal strain rates can be determined knowing only the succession of finite strains.

Several important measures of the degree of coaxiality are available; and in some cases it is possible to determine them in naturally deformed rocks. We will now define three measures of path coaxiality.

1. If a coaxial path has the finite, incremental, and strain rate ellipses all parallel at any one instant in a rigidly moving frame, $\phi i = \phi r = \phi f$ = constant at any t. So one measure of coaxiality is the angle between current finite and strain ellipses (ϕf-ϕr) and its history with t.

2. The angle between two successive finite strain ellipses is $\Delta\phi f$, and the rate at which new material lines became axes of successive finite strain ellipses is $\frac{d\,\phi f}{dt}$. In other words, as the finite strain ellipse develops, the ends (or vertices) of the ellipse move around occupying new material points on the perimeter at a rate $\frac{d\,\phi f}{dt}$.

3. New strain-rate ellipses are forming at each instant, and the rate with which new material lines become axes of the strain-rate ellipses $\frac{d\,\phi r}{dt}$ is a third measure of coaxiality.

The concept of coaxiality has important implications for the material properties of rock. For example, Lifshitz (1963), Green (1970), and Weertman (1970) argue that purely diffusive flow laws of the Nabarro-Herring or grain-boundary diffusion type can never lead to steady-state flow, because the shape of the grains changes with time. Green (1970) and Weertman (1970) considered the special case of a coaxial deformation path. Now an incremental strain on the grain is reflected by an incremental change in the grain's shape and

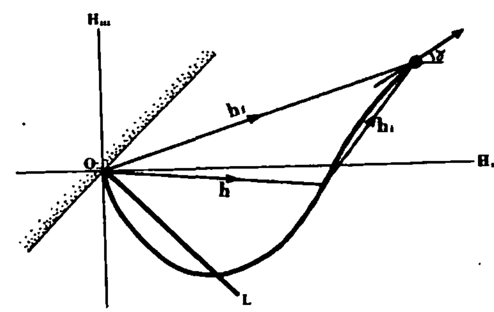

Figure 6. The linear plane strain path OL separates the H_I H_{III} plane into two sectors, and all strain states found above or below OL have an area increase or decrease. Since $H_I \geq H_{III}$, strain states in the stippled region are impossible.

size, and the largest possible alteration to a grain's dimensions with a given number of increments is accomplished in a coaxial deformation path. Consequently, the approximation to steady-state flow gets better and better as the path becomes more and more noncoaxial.

Linearity is the approach of the strain path to a straight line in natural strain space, and ideally is specified by the radius of curvature of the strain path. *Linear coaxial deformation paths* have constant axial ratios of the strain-rate ellipsoids as well as being coaxial.

$$\frac{H_{II}}{H_I} = \frac{\dot{e}_{II}}{\dot{e}_I} = \text{constant for all t.}$$

Such paths plot as straight lines through the origin in natural strain space (Fig. 6).

Linear coaxial deformation paths have been aptly called the "simplest possible strain paths" by Flinn (1962, 1965) and have been more precisely defined by Ramsay (1964, 1967). In the mechanics literature such paths are described as "proportional," "radial," or "linear."

It is apparent that a coaxial path is very much simpler than a noncoaxial one. For noncoaxial paths the finite strain path can still be plotted in natural strain space and the linearity described, but strain increments are no longer chords to the strain path, and strain rates cannot be shown as vectors tangential to the strain path. The matrix equations must be used.

FOLDING AND PLANE STRAIN

Let us apply the preceeding analysis of deformation paths to the important and frequently assumed case of folds growing in plane strain. In the widest sense of plane-strain flow, the normal to a flow plane is a principal strain which is of constant magnitude everywhere at any one instant. Rotation is only possible about an axis normal to the flow plane. All lines normal and all planes parallel to the flow plane remain straight. Folds developing in such a general plane-strain field are cylindrical with hinge lines parallel to the rotation axis. The hinge-line direction remains a principal direction throughout and is a coaxially accumulating principal strain. Consider two-dimensional deformation paths on the profile plane of symmetrical folds developing in a general plane-strain field. From symmetry, material points lying on a fold-hinge surface must be undergoing coaxially accumulating deformation paths, whereas the limbs have new material lines becoming principal strains at each instant, and the paths are noncoaxial.

Plane strain is frequently used in a stricter sense, with no volume strain and the principal strain normal to the flow plane specified as a direction of unchanged length throughout the deformation. If the H_{II} principal direction is parallel to the hinge line of a fold, then in the stricter sense of plane strain $H_{II} = 0$, and $H_I = -H_{III}$. So although the strain paths are everywhere linear it is only at special spots that the deformation paths are coaxial.

Theoretical and computer studies of folding have always assumed cylindrical folds whose hinge lines parallel the axis of rotation in plane strain. But the cylindrical nature of a fold is by itself insufficient to prove plane strain, and one must use the additional evidence provided by grooves and striations on bedding planes (Wegmann and Schaer, 1957) or fibrous syntectonic mineral growths (Wickham and Elliott, 1970).

Why are deformation paths so important for the study of natural folds? Consider a bed with a nonlinear flow law in a more ductile medium. The folds developed in the bed will have a different geometry every time a different history of velocities at the boundaries is used. Consequently the history of velocity boundary conditions is amongst the most important features to be determined on a natural fold (Chapple, 1969).

Examine the profile of a symmetrical fold growing in a plane-strain field. The local velocity v' (in natural time) of one point Q relative to another point P a distance away ds is (Fig. 7)

$$v' = [\dot{D} - I]\, ds .$$

Any convenient length s will serve; for example, the arc length measured from the hinge along any particular bed. The velocity varies smoothly from point to point along the bed, so considering the hinge h as stationary, the velocity vi of the inflection point on the limb is

$$vi = \int_0^s [\dot{D} - I]\, ds .$$

Clearly, if the deformation paths at a sufficient number of points in a fold are known, the history of the velocity boundary conditions may be determined. This result holds regardless of the type of material properties of the rock.

Velocity boundary conditions cannot be derived *a priori* from folding theory; they must be measured from the deformation paths. It is not an intractable problem, but something which can be measured with careful field work and thin-section study. Even a very few points with known deformation paths have enormous significance for the theoretical interpretation of a fold.

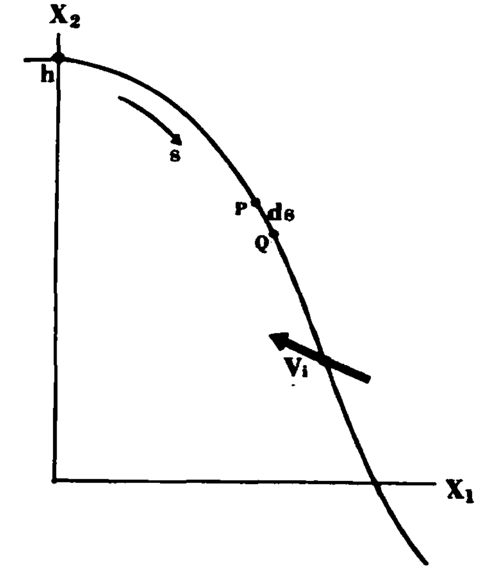

Figure 7. Relation between velocity boundary conditions and deformation paths on a symmetrically folded bed.

DEFORMATION PATHS FROM MICROBOUDINAGE

The remainder of this paper shows how many features of deformation history can be calculated in a practical way from boudinage, pressure shadows, and inclusion trails, and that the concept is important in any discussion of the origin of folded boudins and of cleavage and schistosity. An interesting field example described by Badoux (1963) illustrates some of the possibilities of measuring deformation paths. Because his paper might not be readily accessible, some of his data is reviewed along with the further conclusions derived here.

Badoux measured 300 deformed belemnites from a small area in the Morcles Nappe. The belemnites all lie in bedding surfaces approximately parallel to a strong cleavage (Fig. 8). Let us assume that this cleavage is developed parallel to the intermediate and long axes of the total finite strain ellipsoid. A mineral lineation on the cleavage is parallel to both the long axis of the finite strain ellipsoid and the fold hinge line. The belemnites have all been fractured, and because belemnites break after very small extension, the original and final lengths ℓo, ℓ are measured. From the orientation α and ratio $\frac{\ell}{\ell o}$ for each belemnite the arithmetic mean $\frac{\ell}{\ell o}$ for each 5° of orientation α is calculated, then the total strain ellipse constructed (Fig. 9). The ellipse is well defined, demonstrating that the area over which the belemnites were measured is small enough for the strain to be homogeneous.

Badoux deduced that the fractures were initially filled by quartz segregating into the neck, then metamorphic conditions changed and calcite started crystallizing into the growing crack between quartz and belemnite. It is possible to construct the finite strain ellipse for two separate events: the quartz infilling, and the calcite infilling (Fig. 9).

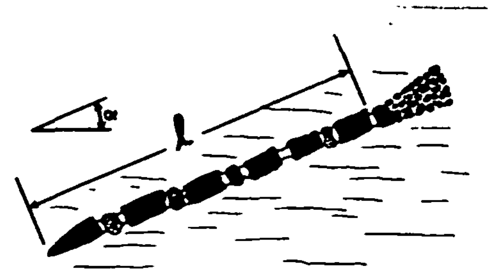

Figure 8. Boudiné belemnite (black) of current length ℓ lying in a cleavage plane at an angle α to a mineral lineation quartz (stippled) and calcite (white) infilling boudin necks. (Redrawn after Badoux, 1963, Fig. 1.)

The axes of the two strain ellipses are parallel; therefore, the two-dimensional deformation path was *coaxial.* When the results are plotted, we see that the strain path is *linear* as well as coaxial (Fig. 10).

The belemnite fragments are only slightly squashed normal to the cleavage; however an independent estimate of the third principal strain is provided, as Badoux considers that the density of the rock remained constant during the deformation. Consequently, if the quartz and calcite are locally derived from adjacent host rock there is no volume change and

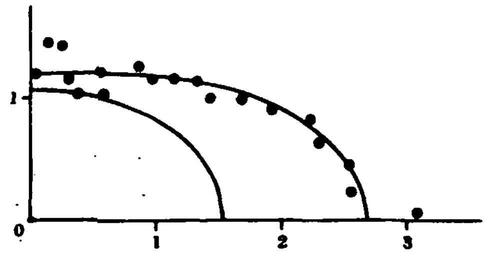

Figure 9. On a polar graph the radius vector is indicated by dots representing the arithmetic mean $\left(\frac{\ell}{\ell_o}\right)$ of all belemnites after the calcite finished infilling for each 5° sector of α. The outer strain ellipse for the calcite infilling fits the data well. The inner strain ellipse for the quartz infilling is shown without data points (redrawn *after* Badoux, 1963, Fig. 3).

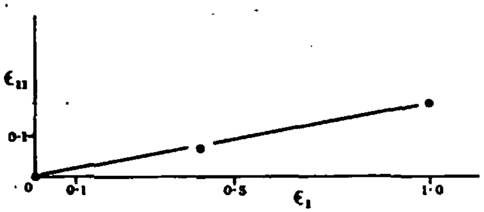

Figure 10. Linear strain path in ϵ_I ϵ_{II} plane defined by the unstrained state at the origin and the strain states after the quartz and after the calcite finished infilling. Without further information it is not possible to define the correspondence between ϵ_I ϵ_{II} and the H_I H_{III} planes.

$$H_{11} + H_{22} + H_{33} = 0 \text{ for all t.}$$

When the strain perpendicular to cleavage is calculated, it appears that very large compressive strains are not recorded by belemnites, probably due to their greater resistance to compression than to extension.

In summary, we know the three finite principal directions and strains after the infilling of calcite if we assume cleavage is a principal plane. For the earlier deformation recorded by the quartz infilling we have the principal direction and magnitude of the strain parallel to the fold hinge line, but the angle of cleavage to bedding at this time is unknown, so the two other principal directions and values are unknown. The deformation path in the profile section might be evaluated by the pressure shadow method discussed below. Suitable boudinage and microboudinage is available from a wide variety of areas which can be analyzed in similar fashion (Misch, 1969, 1970).

DISTRIBUTION OF FOLDS AND BOUDINS

It has been suggested that noncoaxial deformation paths might be indicated by the angular distribution of folded boudins. This is because in an area undergoing homogeneous deformation, layers in some directions may be extending and necking down while in other directions they will be shortening and folding (Flinn, 1962; *reviewed in* Ramsay, 1967). With coaxial deformation paths, any layer undergoing shortening always moves toward that sector of the strain ellipsoid where it will be extended. However during noncoaxial deformation it is possible that layers undergoing extension can later undergo shortening, but even then only in that sector of the finite strain ellipsoid which the long axis has been moving away from. This is where folded boudins should be found (Ramsay, 1967, p. 120, 175).

Talbot (1970) devised an ingenious way of determining the orientation and shape of the finite strain ellipsoid using only the orientation of folded and bouding veins. Folded boudins produced during one period of deformation were never found, and he concluded that the deformation paths are coaxial in all the areas he investigated.

A layer must undergo appreciable extension for a boudin to be finite and observable, and the path must be sufficiently coaxial for the layer to remain in the extension field during this interval. Similarly, for a fold to become visible, the layer must be shortened for an appreciable time along a path sufficiently coaxial not to remove the layer from the shortening field. As a result, the path required for folded boudins is an initial finite strain on a nearly coaxial path followed by an increment of finite strain on another nearly coaxial portion of the path, with a large angle between the two finite-strain ellipsoids. The overall path necessary to produce folded boudins is a noncoaxial path of a very special type, and the absence of folded boudins only means that this special type of noncoaxial path is uncommon during one period of deformation. During *two* periods of deformation the two finite strain ellipsoids are frequently at a large angle, and the combined path may approach the one required. Folded boudins produced during two periods of deformation have been observed by Talbot (1970).

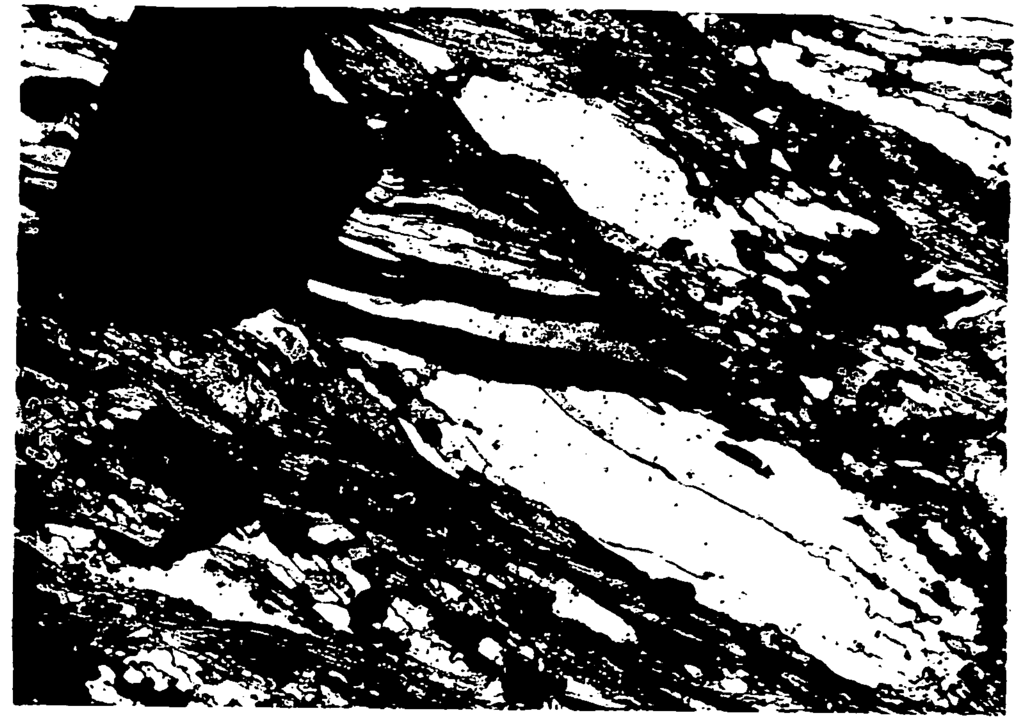

Figure 11. Pressure shadow on a magnetite crystal, crossed nicols. Thin section cut perpendicular to the hinge line of an isoclinal fold in the Weaverton Formation, South Mountain, Maryland.

It appears that the distribution of folds and boudins is not sensitive enough to be of help in distinguishing coaxial and noncoaxial paths.

PRESSURE SHADOWS

Relatively rigid particles such as small pebbles, phenocrysts, porphyroblasts, magnetite or pyrite crystals are often present in a rock at the outset of a deformation. The rigid particle is deflecting the streamlines of the flowing matrix, introducing inhomogeneity into the flow pattern, and giving rise to dissipative forces and a boundary layer about the rigid particle. The deformation path being recorded by growing pressure shadows is that of the boundary layer adjacent to the rigid particle but will show the same trends as in the matrix outside the boundary layer.

The interface between rigid particle and ductile matrix is a sharp discontinuity in material properties. During a deformation increment, any part of the rigid particle's surface whose normal is a direction of incremental extension tends to unstick from the matrix, and this is where crystal growth occurs (Mügge, 1930).

Segment boundaries separate the pressure shadow into portions within which the minerals have similar angular relations to a face of the rigid grain (Spry, 1969, p. 240). Quartz, for example, always tries to grow normal to the face of magnetite crystal, and as the crystal moves during deformation the quartz fiber grows curved to remain normal to the face (Figs. 11 and 12). A sharp curve or junction of two faces on the rigid grain boundary traces out a segment boundary as this point slowly drifts away from its previous positions (small black circles on Fig. 12). Pressure shadows, then, are swept out by the boundary of a rigid grain, and certain points on the grain boundary leave a recognizable path line (Fig. 13). The arc length of the path line is a convenient time-like quantity (*see* equation 9).

Successive orientations of the rigid grain can be traced by (1) maintaining a uniform angle between the rigid grain and the adjacent minerals within the segment while (2) keeping

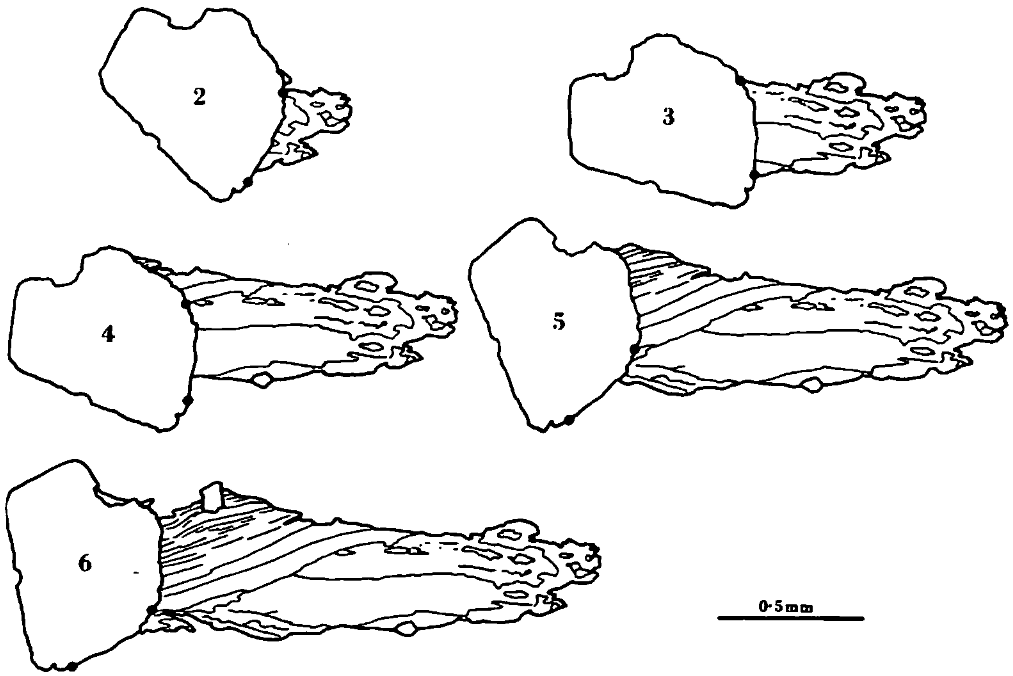

Figure 12. Growth from the initial state without pressure shadow (step 1 not shown) through to final result (6) shown in photograph (Fig. 11).

points of the grain boundary on their path lines (Figs. 12, 13). At any instant the pressure shadow is limited by two points on the rigid grain boundary outside of which minerals do not grow. The normal to the grain boundary at these two points are directions of zero longitudinal strain rate. The variation of the angle between these directions (2V′) in a plane strain situation is ideally the 2V of the strain rate ellipsoid (Fig. 14). Because axes of the strain rate ellipsoid are of very similar length the 2V should in principal be close to 90°. The measured 2V′ is about 15° smaller than the ideal value. This discrepancy could be either a result of seeing a finite increment rather than a strain rate or to complex boundary effects, but the trend of 2V′ may be indicative of how the true 2V is varying.

The principal directions of strain rate bisect the 2V′. In the example illustrated the deformation path is clearly noncoaxial. Accuracy may be established by comparing pressure shadows from both sides of the same grain and from different grains in the same slide.

Published illustrations of pressure shadows suggest that coaxial paths are not rare; good examples are in Burger and others (1970), and Ramsay (1967, p. 181). Pressure shadows suggesting noncoaxial paths around folds are illustrated by Zwart and Oele (1966) and Langheinrich (1964), but unfortunately these

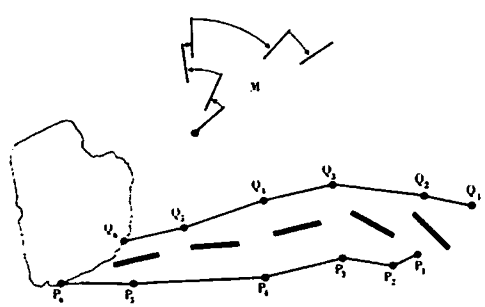

Figure 13. Path lines traced out by two points on boundary of magnetite (small solid circles in Fig. 12). Bars show principal strain-rate directions, and M is the history of re-orientation of the line P Q.

two examples are not really detailed enough to allow analysis in depth.

INCLUSION TRAILS AND FORMATION OF SCHISTOSITY

Garnets and many other porphyroblastic minerals may show inclusion trails running through them. Inclusions in garnet are commonly made up of rounded or elongate tapered blebs of quartz in discontinuous stringers, flat bands, or lines of flattened groups of crystals. Several inclusion trails are usually present in one garnet and all trails curve sympathetically with one another.

The garnet grows by extending fingers among the trails of quartz in the surrounding matrix, and these quartz trails are then incorporated into the garnet (Spry, 1963; 1969, p. 255). The orientation of the trail at the time it was swallowed by the growing garnet is thus preserved.

The continuously increasing ratio $\frac{\text{current radius}}{\text{final radius}}$ of a growing garnet serves as a dimensionless time-like parameter t with which to mark the progress of the deformation.

Changes in color of the garnet or breaks in inclusion trails show up any discontinuities in grain growth and are aptly termed "metamorphic unconformities" by Rosenfeld (1969; 1970). They represent discontinuities in the deformation path as well.

The orientation of the inclusion trail at a particular radius is measured with respect to a coordinate x_1 embedded on fixed material particles of the garnet. However, this is not the straightforward angular measurement it would appear to be; an aspect treated in some detail by Powell and Tregus (1970), and Rosenfeld (1970).

The maximum angular difference $\Delta\psi$ between the innermost and outermost part of inclusion trails reaches values for individual garnets as high as 540° in the Scottish Highlands (Flett, 1912; *see* Spry, 1969, p. 256) and 625° in the Northern Appalachians (Rosenfeld, 1968).

An inclusion trail clearly tells something about the deformation path, but what it tells depends upon interpretation of how the quartz trail orientation formed in the matrix before it was enveloped by the growing garnet. This is summarized in Table 1. Many quartz trails seem to form as part of a schistosity which develops in the matrix as the garnet grows, so the inclusion trails preserved in the garnet record the orientation history of the current schistosity. The problem now is to determine the cause of the schistosity.

Most strain determinations measured from snowball garnets involve the assumption that schistosity is a set of finite simple shear planes. An equation derived originally by Schmidt in 1918 was somewhat modified by Spry (1963; 1969, p. 256), but Rosenfeld (1970) shows that all these formulas are incorrect and that the amount of finite simple shear γ_s is

$$\gamma_s = 2\Delta\psi\,.$$

If the schistosity planes have always been the simple shear planes, then a simple shear deformation path is indicated. But can schistosity be regarded as forming parallel to simple shear surfaces, and what is the origin of schistosity?

There is very strong evidence for rocks deformed once that schistosity develops normal to the short axis of the finite strain ellipsoid. In particular whenever it is possible to calculate the strain from deformed oölites, fossils, and conglomerates, schistosity is always found normal to the short axis of the total finite strain ellipsoid (evidence reviewed by Ramsay, 1967).

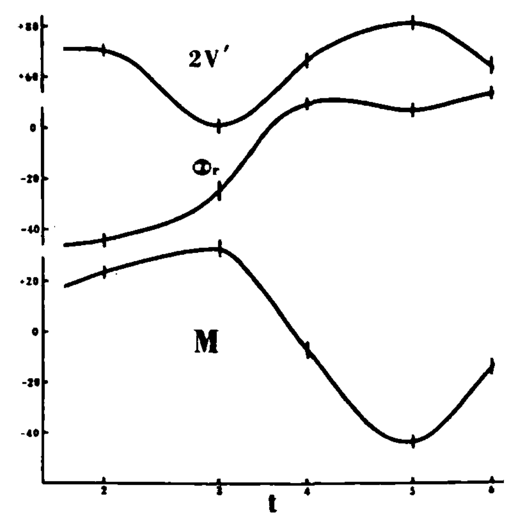

Figure 14. Variation with t in 2V′, orientation of principal strain-rates ϕr, and orientation of the magnetite crystal. Angular datum is arbitrary (ϕ_r = 0, M = 0).

TABLE 1. INCLUSION TRAILS IN METAMORPHIC MINERALS

Origin of included fabric	Assumptions	Results
		Attitude of schistosity versus t
Older Schistosity or Inherited Lithological Layering	Principal directions of finite strain	Axial ratio
	Principal directions, coaxial path	Axial ratio versus t
	Finite simple shear on known shear planes	γs
	Simple shear deformation path on known shear planes	γs versus t
Current Schistosity	Finite simple shear on schistosity planes	γs
	Simple shear deformation path on schistosity planes	γs versus t
	Schistosity parallels current finite principal strains	ϕf versus t
	Schistosity parallel current incremental principal strains	ϕi versus t
Growth over Pressure Shadow	Analyse as Pressure Shadow	ϕr and 2V' versus t

Further evidence is provided from the strain distribution across ductile shear zones (Ramsay and Graham, 1970). *If* schistosity is always normal to the short axis of the strain ellipsoid then the garnet inclusion trails give a history of lines which have been in principal planes of the finite strain ellipsoids, and a graph of the coaxiality of the deformation path can be constructed, ϕf versus t.

In rocks that have been deformed by two separate periods the total finite strain ellipsoid in general has a different orientation from either individual strain ellipsoid. From the theory above *only one schistosity* should be expected, normal to the total finite strain ellipsoid. I only know of one instance when this effect is claimed to occur (Ramsay, 1963, p. 375), and in all other areas each individual deformation phase develops its own schistosity and cross-cuts any earlier ones. It appears that schistosity cannot always be normal to the shortening direction of the total finite strain ellipsoid.

There are three other independent possible causes of schistosity: (1) After a certain definite amount of finite strain, deformable grains and clastic particles suddenly and discontinuously acquire a preferred orientation and cleavage as a result of the "shape factor" (Elliott, 1970a). (2) Schistosity is sometimes due to rigid body reorientation of platy minerals through a ductile matrix, and this has no necessary simple relation to any stress or strain ellipsoids. (3) Subparallel pressure solution surfaces contribute toward a visible cleavage and are the dominant process in some rocks.

Another possibility is that schistosity may appear normal to the short axis of the last incremental strain ellipsoid which was just sufficiently large to produce the three above effects to a noticeable degree. In this case the mineral inclusion trails record the material lines which have been axes of the incremental strain ellipse, so ϕi versus t could be measured. This incremental strain ellipsoid may even be large enough to approximate the orientation of the total finite strain ellipsoid.

The quartz trails in the matrix could be part of a lithological layering such as bedding, or an old foliation inherited from an earlier period of metamorphism. The inclusion trails give us the history of orientation of a lithological layer but without further information nothing more can be said. If the finite principal directions at any instant can be identified, the axial ratio of the total finite strain ellipse can be calculated from Harker's formulae (Ramsay, 1962).

It is not clear whether cleavage and schistosity are tracking stress, strain rate, finite strain, or an incremental strain of some finite magnitude. Distinction between the various hypotheses is only possible in a region of noncoaxial deformation paths.

Several authors have noted that internal inclusion trails do not look like remnants of the schistosity presently seen in the matrix as the shape and size of the quartz in the trails is different (Spry, 1963). It is well established that garnet and other porphyroblasts have remained essentially rigid during deformation, and quartz-filled pressure shadows are found on garnets (Harker, 1939, p. 219; Wilson, 1971, Figs. 1 and 3). The garnet growing over the pressure shadow would include the quartz fibers and a new pressure shadow would form at the end of the newly enlarged garnet (Fig. 15). Inclusion trails which were "frozen" pressure shadows could be interpreted in the same way

as the shadows discussed earlier, but there is a practical difficulty as they are much less complete. Approximations have to be made, such as assuming that the fibers grew roughly in the direction of current incremental principal extension.

If inclusion trails in snowball garnets are remnants of fibrous growth in pressure shadows this is one of the few bits of direct evidence that diffusion-flow laws might be operative under fairly high-grade metamorphic conditions—evidence that usually would be destroyed by the annealing recrystallization so widespread at these temperatures (Elliott, 1970b, 1972).

An important case arises when the deformation path is coaxial during the garnet growth. The inclusion trail within the garnet would then be straight and parallel to the external schistosity—regardless of whether the quartz inclusion trails formed as a schistosity parallel to the strain ellipse or as quartz fibers in a pressure shadow. But such textures have traditionally been interpreted as postkinematic (Fig. 16a). Posttectonic recrystallization episodes postulated on this evidence alone can be reinterpreted as synkinematic grain growth during a coaxially accumulating deformation path.

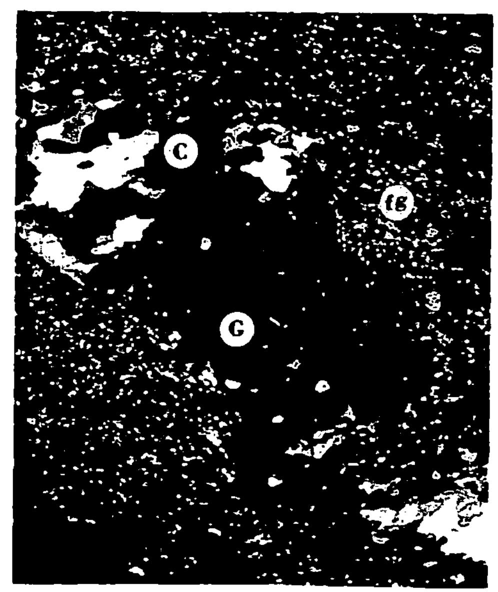

Figure 15. Photo of coarse quartz (c) in pressure shadow at upper left and bottom right of garnet, surrounded by fine-grained quartz (fg). Garnet (G) is engulfing the pressure shadow quartz. Crossed nicols. Garnet-mica gneiss collected by R. Balk in the Benardston quadrangle, Massachusetts.

A particularly strong case for synkinematic grain growth arises when the inclusion trails and the porphyroblast composition are either smoothly varying or constant (Fig. 16a, b).

Inclusion trails which change from curved to straight are conventionally interpreted as indicating a change from synkinematic to static postkinematic grain growth. Alternatively, the grain growth could be entirely synkinematic and reflect the change from a noncoaxial to a coaxial deformation path (Fig. 16c).

The initiation and development of structures in higher-grade metamorphic terrains may be followed by an extended interval of approximately homogeneous flattening (Ramsay, 1962; Mukhopadhyay, 1965). Homogeneous flattening implies a coaxially accumulating deformation path. Is it possible that the straight portions of inclusion trails recorded from such terrain as "static postkinematic" are in fact evidence of the coaxial portions of such paths?

In the spectacular snowball garnets of southeastern Vermont—some of the finest in the world—Rosenfeld (1969) reports that inclusion trails in garnets measured around the major Ascutney fold have different senses of curvature on each limb, except at one place where the inclusion trails are straight. At this locality he concludes that there was static postkinematic grain growth, although synkinematic grain growth holds everywhere else. Unlike the symmetric folds discussed earlier, material points which have coaxial deformation paths in an asymmetric fold will not lie on the hinge surface; if they exist they are somewhere on the long limb side of the hinge. Is it possible that the straight inclusion trails found at one spot

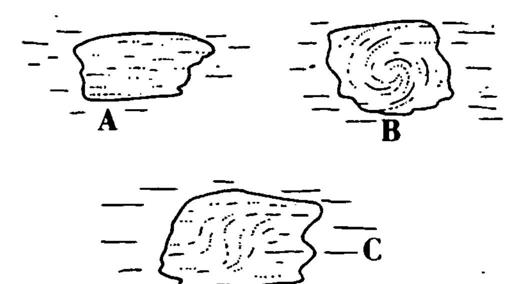

Figure 16. Typical patterns of inclusion trails in metamorphic minerals when thin section is cut normal to the rotation axis.

on the major Ascutiny fold are at a point of coaxially accumulating deformation path, and that the garnet growth here is synkinematic as it is everywhere else?

There are still valid criteria for distinguishing synkinematic from postkinematic grain growth. Straight inclusion trails produced during synkinematic grain growth along a coaxially accumulating path should only occur in restricted zones bordered by regions with oppositely curved synkinematic inclusion trails. The habit remains useful, for true postkinematic crystals may grow randomly across the schistosity, while synkinematic grains often are elongated in the schistosity.

DISCUSSION

Analysis by means of matrices and natural strain space gives a clear picture of the various aspects of deformation paths. Linearity and coaxiality emerge as two particularly important properties which are significant not only for the evolution of cleavage, boudinage, and folds but also have ramifications on the flow laws or stress-strain rate relations. There is a clear need for many more field observations from a variety of rocks to see which of the various possible systematic patterns of deformation history are important in naturally deformed rocks.

One of the most powerful ways of comparing computer experiments on folding with natural folds is by means of their deformation paths, something which has not as yet been attempted. Several new methods for determing deformation paths from boudinage, pressure shadows, and inclusion trails are proposed. Unfortunately some of the methods in current use, such as interpreting principal stresses within crystals or the distribution of folded boudins, do not appear to have much promise for describing deformation paths. Suitable material might be available in metamorphic rocks of most grades; examples noted in this paper range from greenschist and blueschist up to garnet-grade rocks. Inclusion trails in metamorphic minerals may be more valuable for measuring path coaxiality than they are for defining the time relations between metamorphic and deformation periods.

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A Plate Tectonic Model for the Origin of Porphyry Copper Deposits

RICHARD H. SILLITOE

Abstract

The theory of lithosphere plate tectonics, embodying the concepts of sea-floor spreading, transform faulting, and underthrusting at continental margins and island arcs, is employed as a basis for an actualistic, though speculative, model for the origin and space-time distribution of porphyry copper and porphyry molybdenum deposits.

Porphyry ore deposits, occurring in the western Americas, southwest Pacific and Alpide orogenic belts, are thought to constitute a normal facet of calc-alkaline magmatism. Chemical and isotopic data cited are consistent with the generation of the components of calc-alkaline igneous rocks and porphyry ore deposits by partial melting of oceanic crustal rocks on underlying subduction zones at the elongate compressive junctures between lithospheric plates.

It is proposed that the metals contained in porphyry ore deposits were derived from the mantle at divergent plate junctures, the ocean rises, as associates of basic magmatism, and transported laterally to subduction zones as components of basaltic-gabbroic oceanic crust and small amounts of suprajacent pelagic sediments; evidence supporting the presence of significant amounts of metals in the oceanic crust is listed.

It is suggested that the temporal and spatial distribution of porphyry ore deposits is dependent on two principal factors, namely the erosion level of an intrusive-volcanic chain, and the time and location of magma generation, and the availability of metals, on an underlying subduction zone. The erosion factor is believed to offer an explanation for the paucity of porphyry ore deposits in pre-Mesozoic orogenic belts, and for the relative abundance of exposed porphyry deposits of Upper Cretaceous-Paleogene age in post-Paleozoic orogens. Provinces with a high concentration of porphyry copper deposits, such as southern Peru-northern Chile and the southwest United States, may be interpreted as regions beneath which anomalously copper-rich oceanic crust was subducted at the time of porphyry copper emplacement; one possible explanation for the episodic formation of volumes of copper-rich oceanic crust is the presence of a heterogeneous distribution of metals in the low velocity zone of the upper mantle. Porphyry ore deposits seem to have formed during a series of relatively short, discrete pulses, perhaps correlable with changes in the relative rates and directions of motion of lithospheric plates. In some regions, such as Chile, porphyry ore deposits are arranged in parallel, linear belts, which may be explicable in terms of shifting loci of magma and included metal generation on a subduction zone, and which seem to be largely independent of control by tectonic lineament intersections. The time intervals during which the formation of porphyry deposits took place are shown to be broadly coincident with periods of lithosphere plate convergence, and porphyry deposits may still be forming above currently active subduction zones.

A number of potential regions for the discovery of porphyry ore deposits are suggested, and the importance to exploration of analyzing orogenic belts in terms of plate tectonics is emphasized.

Introduction

THE sea-floor spreading hypothesis of Dietz (1961) and Hess (1962) has recently been further developed by Isacks, Oliver and Sykes (1968), Le Pichon (1968), McKenzie and Parker (1967) and Morgan (1968), resulting in the theory of lithosphere plate tectonics (the new global tectonics), the only existing global tectonic model compatible with much recently accumulated geological and geophysical data. The model considers the Earth's surface to be divided into six large and several smaller lithospheric plates (Fig. 2), which accrete at ocean rises by uprise of subcrustal basic magmas, slide past one another along transform (horizontal shear) faults, and are destroyed at trench systems by descent into the asthenosphere down inclined subduction (Benioff) zones (Fig. 3).

The concepts of the new global tectonics have rendered the stabilist geosynclinal theory of orogeny outmoded; it is now realized that a generalized model.

for geosynclinal development, commencing with the accumulation of a thick volcanic and sedimentary pile, and followed by plutonism and deformation and finally by epeirogenic uplift and volcanism (e.g., Beloussov, 1962) does not satisfactorily explain the evolution of most orogens (Coney, 1970). In terms of the theory of plate tectonics, geosynclines may be equated with oceans and continental margins, and oceans and island arcs. Mountain systems are generated as a consequence of the underthrusting of oceanic lithosphere beneath an adjacent plate at continental margins (e.g., the cordilleran system of North and South America), or at island arcs (e.g., Japan). In some instances, eventual continent-continent collision (e.g., the Himalayas) or continent-island arc collision (e.g., New Guinea) are involved in orogenic development (Dewey and Bird, 1970; Dewey and Horsfield, 1970). It should be stressed, however, that each orogenic belt displays an essentially unique sequential history, even though certain sequences of events are more common than others.

Currently accepted concepts of the metallogenesis of post-Precambrian orogenic belts are based on stabilist geotectonic doctrine (Bilibin, 1968; McCartney and Potter, 1962; McCartney, 1964), and support the association of distinct types of mineralization with each stage of geosynclinal development. Porphyry copper deposits, for instance, are considered to typify the post-orogenic, late tectonic stage (McCartney and Potter, 1962). In view of the recent advances in geotectonic theory, it would therefore seem opportune to reexamine the metallogenesis of orogenic belts in terms of the new global tectonics. This paper outlines a plate tectonic model to account for the genesis and distribution in both space and time of one important class of mineralization, the porphyry copper and porphyry molybdenum deposits. Some aspects of the model were presented elsewhere as an abstract (Sillitoe, 1970). It should perhaps be stressed that the proposed model is of a speculative nature, and does not purport to embody rigorous proofs of its validity.

Geological and Genetic Characteristics of Porphyry Copper Deposits

Over one half of the world's copper production is currently derived from porphyry copper deposits, large tonnage (commonly exceeding 500 m tons), low grade, roughly equidimensional deposits of disseminated and stockwork-veinlet, pyrite-chalcopyrite mineralization, carrying at least trace amounts of molybdenum, gold and silver. They are spatially and genetically related to passively-emplaced hypabyssal felsic stocks, commonly porphyries. The geological characteristics of the porphyry molybdenum deposits, from which molybdenum is the principal metal recovered, are similar to those of the porphyry coppers. Hypogene ore grade in the porphyry coppers rarely exceeds 1% Cu, and is commonly below 0.5% Cu.

The host intrusions of porphyry copper deposits, and their various types of country rocks, may both be ore-bearing, and are characterized by widespread, zonally-arranged hydrothermal alteration, commonly of potassic, phyllic, argillic and propylitic types (Meyer and Hemley, 1967; Lowell and Guilbert, 1970), and by hydrothermal brecciation.

The close association of intrusion and mineralization in porphyry copper and molybdenum deposits is emphasized by K-Ar dating which has demonstrated the two processes to be temporally indistinguishable in several instances (e.g., Livingston, Mauger and Damon, 1968; Moore, Lanphere and Obradovich, 1968; Laughlin, Rehrig and Mauger, 1969). In at least some deposits, part of the mineralization may in fact be syngenetic with respect to the associated intrusive rock (e.g., Ely, Nevada; Fournier, 1967). The intimate temporal association of intrusion and mineralization lends support to the orthomagmatic model of porphyry copper genesis (Nielsen, 1968; Lowell and Guilbert, 1970). According to this model, a felsic magma, becoming water-saturated as it intrudes towards the surface zone, undergoes crystallization of its outer parts, which are subsequently brecciated by the release of accumulated fluids, which also produce the alteration and mineralization. Meteoric waters are involved in the formation of the outer zones of hydrothermal alteration and mineralization.

The Origin of Calc-Alkaline Igneous Rocks and Porphyry Copper Deposits

World Distribution of Porphyry Ore Deposits

Figure 1 shows the location of the majority of exploited porphyry copper and molybdenum deposits, and of many important prospects, which in Figure 2 are related to Mesozoic-Cenozoic orogenic belts and currently active lithospheric plate boundaries. It can be appreciated that the majority of the world's porphyry deposits are located in the circum-Pacific orogenic belts and in the central portion of the Alpide orogenic belt. The western Americas belt, containing most of the known porphyry copper deposits, extends from western Argentina and central and northern Chile, through Peru, Ecuador, Panama, Mexico, the western United States (Arizona, New Mexico, Nevada, Utah, Colorado, Idaho, Washington and Montana), to British Columbia, the Yukon and Alaska. Marked concentrations of deposits occur in Sonora-Arizona-New Mexico and in British

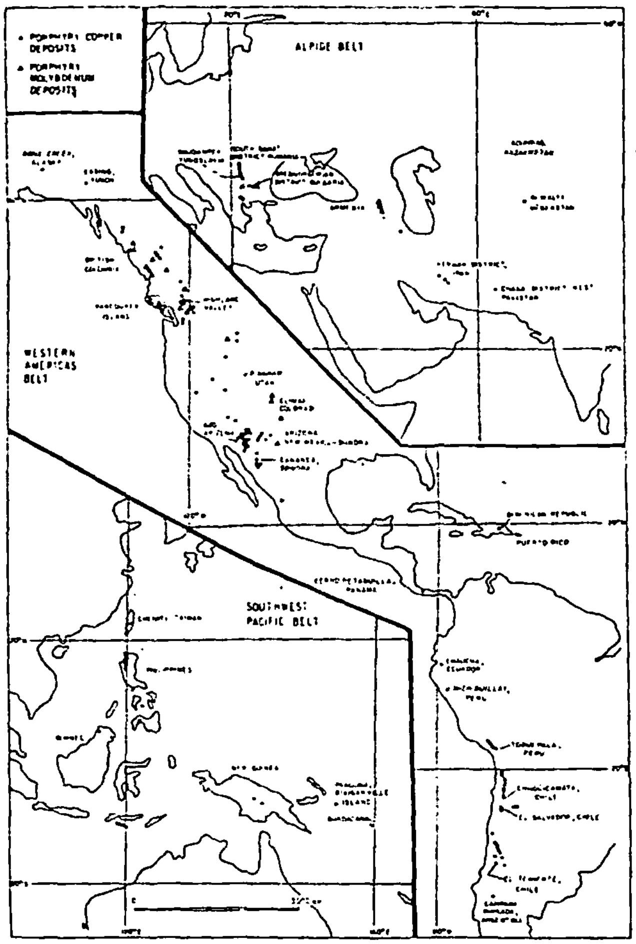

FIG. 1. Porphyry copper and molybdenum deposits and prospects in the western Americas, southwest Pacific and Alpide belts.

Columbia. Deposits in the Dominican Republic and Puerto Rico may be considered as an offshoot of the western Americas belt. Two other belts, for which published information on the porphyry deposits is scant, are located in the Taiwan, Philippines, Borneo, West Irian, Papua-New Guinea and Solomon Islands region (the southwest Pacific belt), and in the South Banat district of Romania, Yugoslavia, central Bulgaria, Armenia, Iran and West Pakistan (the Alpide belt). The only well-authenticated occurrences of porphyry deposits outside of these post-Paleozoic orogenic belts are those in Uzbekstan and Kazakhstan, USSR.

Relationships between Calc-Alkaline Igneous Rocks and Porphyry Copper Deposits

The post-Paleozoic history of the continental margins and island arcs where porphyry copper deposits are located was characterized by widespread calc-alkaline volcanism which gave rise to basalts, andesites, dacites, rhyolites and, in some parts, felsic ignimbrites. These volcanic rocks are commonly observed to be intruded or underlain by extensive batholiths and smaller intrusions of a similar composition. Hamilton and Myers (1967) and Hamilton (1969a, b) have convincingly demonstrated the consanguinity of the calc-alkaline volcanic suite and the spatially related felsic plutonic rocks, the latter interpreted as the roots of major eruptive chains.

The location of porphyry copper orebodies in the cupolas of plutons of intermediate composition was emphasized many years ago (Emmons, 1927), and subsequent work tends to confirm the high-level, subvolcanic nature of their environment of formation. Some porphyry copper deposits may have been emplaced at very shallow depths, perhaps at less than 1,500 m (Fournier, 1968). This is emphasized by the occurrence at Bingham, Utah of a porphyry-copper stock and nearby penecontemporaneous, comagmatic volcanic rocks (Moore, Lanphere and Obradovich, 1968).

It is suggested that the accumulation of copper and molybdenum in high-level felsic stocks was a normal part of calc-alkaline magmatism in post-Paleozoic orogenic belts.

The Origin of Calc-Alkaline Igneous Rocks

Many workers (e.g., Benioff, 1954; Coats, 1962; Dickinson and Hatherton, 1967; Dickinson, 1968; Ringwood, 1969) agree that the magmas which have given rise to calc-alkaline volcanic rocks—and by analogy their plutonic equivalents (Hamilton, 1969a)—were generated on subduction zones which underlay the eruptive chains. The magmas were probably generated by partial fusion consequent upon frictional heating of subducted, water-saturated oceanic lithosphere which originally was generated at ocean rises, and transported laterally into ocean trenches (Fig. 3). Restricted volumes of fused ocean-floor sediments (layer 1) and the lowest melting fractions of layers 2 and 3 of the oceanic crust seem to be the most likely source materials for calc-alkaline magmas (Oxburgh and Turcotte, 1970). Several chemical and isotopic parameters determined for calc-alkaline volcanic and intrusive rocks are now cited in support of an origin by partial melting on a subduction zone.

Recent volcanic rocks in the circum-Pacific belt show a systematic increase in their potash to silica ratios landwards from the trench (Dickinson and Hatherton, 1967; Dickinson, 1968); these ratios correlate with the depth from the site of eruption to the Benioff zone dipping beneath the island arc or continental margin, suggesting an origin on the Benioff zone, and the absence of widespread crustal contamination. A comparable landward increase in potash has also been demonstrated for post-Paleozoic volcanic and intrusive rocks in parts of western

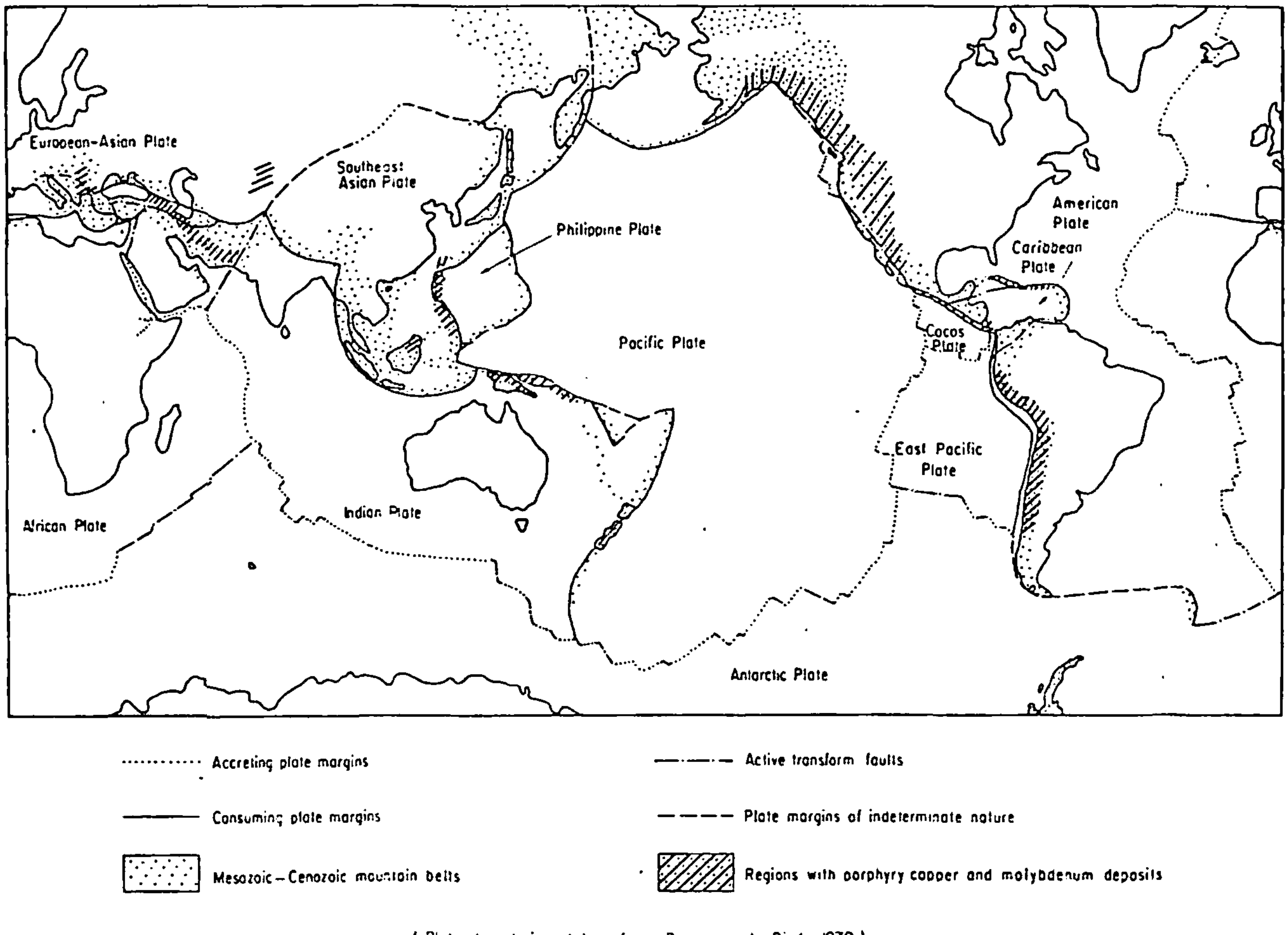

FIG. 2. The western Americas, southwest Pacific and Alpide porphyry belts in relation to Mesozoic-Cenozoic orogenic belts and accreting and consuming plate boundaries

North America (Moore, 1959, 1962; Moore, Grantz and Blake, 1963; Bateman and Dodge, 1970).

Initial Sr^{87}/Sr^{86} ratios (0.703–0.706) obtained for andesitic volcanics (Ewart and Stipp, 1968; Pushkar, 1968; Peterman, Carmichael and Smith, 1970), and those (0.705–0.709) obtained for felsic rocks from the British Columbia, Sierra Nevada and Boulder batholiths (Fairbairn, Hurley and Pinson, 1964; Hurley et al., 1965; Doe et al., 1968) are incompatible with an origin by partial melting or wholesale assimilation of continental crust, but would seem to be in accord with a derivation by partial melting of oceanic crust on a subduction zone. The trace element content of andesites is also consistent with a Benioff-zone origin (Taylor, 1969; Taylor et al., 1969).

The Origin of Porphyry Copper and Molybdenum Deposits

In view of the close temporal and spatial relationship between the genesis of porphyry ore deposits and calc-alkaline magmatism, summarized in the preceeding sections, the components of porphyry-copper stocks, including the contained metals, are likewise postulated to possess an origin by partial melting of oceanic crust on a subduction zone. Initial strontium isotope ratios in the range 0.706–0.708 obtained for several porphyry-copper stocks in the southwest United States and for the stock associated with the Questa, New Mexico porphyry molybdenum deposit (Moorbath, Hurley and Fairbairn, 1967; Laughlin, Rehrig and Mauger, 1969) support this contention. A deep, homogenized, probably mantle-source for sulfide sulfur in porphyry copper and molybdenum deposits in the southwest United States is suggested by δS^{34} values close to the meteoric standard (Field, 1966; Jensen, 1967; Laughlin, Rehrig and Mauger, 1969).

In summation, therefore, porphyry copper and molybdenum deposits are considered to be confined to orogenic belts characterized by calc-alkaline magmatism, and resulting from plates of oceanic

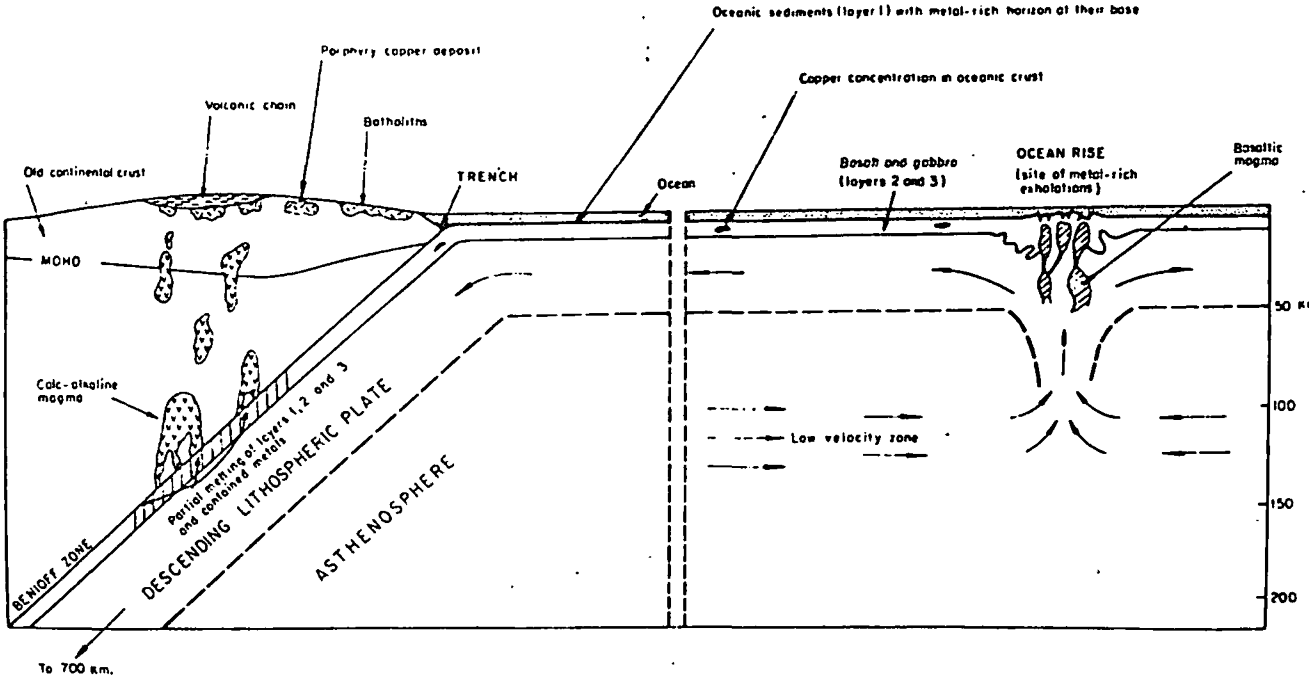

FIG. 3. Schematic representation of the genesis of porphyry copper deposits in the context of plate tectonics.

lithosphere underthrusting adjacent lithospheric plates, in some cases with ensuing continental collision—the compressive type of elongate contact between two lithospheric plates.

The Source of Metals in Porphyry Ore Deposits

The source of metals in post-magmatic sulfide ore deposits has long been a topic of contention. For the porphyry copper deposits in particular, a continental crustal provenance of copper by extraction from andesitic volcanics (Ney, 1966), geosynclinal sediments (United Nations, 1970), or shale horizons (Jensen, 1971), during igneous intrusion, has been inferred. However, considering the wide spectrum of host rock types of porphyry copper deposits, namely calc-alkaline volcanics, intrusive and metamorphic rocks, and marine sediments, partially calcareous, and taking into account the probable absence of certain of these source rock types from beneath some porphyry copper provinces, the provision of the metals by a specific rock type in the continental crust seems improbable. Furthermore, the only source of metals available to all porphyry copper deposits, which takes into consideration the thin crust and almost total absence of sialic crust beneath some porphyry copper provinces (e.g., the Solomon Islands; Coleman, 1966), would seem to be the upper mantle.

An important aspect of the model outlined above for the origin of porphyry ore deposits, is the implication that a large percentage of the metals were extracted from oceanic crust during partial melting as it sank into the mantle on a subduction zone. Collection of metals by saline fluids expelled from oceanic crust during subduction may also be operative. Initially, the metals were largely derived from the mantle at the ocean rise system and carried to th margins of ocean basins as components of layers 1, 2 and 3 of the oceanic crust (Fig. 3). The wedge of mantle above a Benioff zone may also act as a source for basaltic magmas (Oxburgh and Turcotte, 1970), and minor quantities of copper and molybdenum. The metals released during partial melting ascended as components of calc-alkaline magmas (Fig. 3), and were ultimately concentrated in chloride-rich fluid phases associated with the roof-zones of certain intrusions. The fluid phase was released upwards during consolidation of the magma to give rise to the typical upright cylinders of porphyry copper and molybdenum mineralization.

A corollary of this premise is that regions possessing high concentrations of porphyry copper deposits, such as the southwest United States and southern Peru-northern Chile, and considered as copper-rich metallogenetic provinces (e.g., Turneaure, 1955), are not regions of the Earth where the subjacent continental crust or upper mantle are enriched in copper, but regions beneath which anomalously copper-rich oceanic crust, including pelagic sediments, has been subducted. Variation in the amount of copper consumed in a zone of subduction, and therefore potentially available for mineralization, may depend on the rate of sea-floor spreading and hence the volume of oceanic crust entering the subduction zone, or the intensity of volcanism and metal production at the ocean rise, or, more fundamentally, may reflect an inhomogeneous distribution of this metal in the upper mantle, perhaps in the low velocity zone (Fig. 3); chemical heterogeneity of the mantle has recently been demonstrated (Peterman and Hedge, 1971).

Evidence Bearing on the Derivation of Metals from Oceanic Crust

The following points are thought to provide evidence favoring the oceanic crust as a source for the copper and molybdenum contents of porphyry ore deposits:

1. Recent workers (Cann, 1968; Oxburgh and Turcotte, 1968; Thayer, 1969; Christensen, 1970) have proposed that the third layer of the oceanic crust has a gabbroic composition, perhaps with dolerite dike complexes in its lower part (Dewey and Bird, 1970). It is overlain by basalts of layer 2. Basic igneous rocks have average copper and molybdenum contents of the order of 100 ppm and 1.5 ppm, respectively, five times that of granitic rocks in the case of the former element (Turekian and Wedepohl, 1961; Vinogradov, 1962). Oceanic basalts (layer 2) in the Atlantic and Pacific Oceans average 77 ppm Cu (Engel, Engel and Havens, 1965).

2. Sulfide phases containing up to approximately 10 percent copper are present in vesicles and as globules in pillow basalts from layer 2 on the ocean rises (Moore and Calk, 1971).

3. Sulfide grains of a similar nature to those occurring in ocean-floor basalts (but with up to 25 percent copper) have been discovered in recent Hawaiian basalt flows (Desborough, Anderson and Wright, 1969), and in basaltic oozes which flowed into a drill hole in the crust of Alae lava lake, Hawaii (Skinner and Peck, 1969). The Hawaiian islands are mantle-fed volcanoes in the center of the Pacific lithospheric plate (Menard, 1969).

4. Large alpine-type mafic-ultramafic (ophiolite) complexes such as those in Cyprus, Turkey, Papua and elsewhere are thought to represent fragments of surficial rinds of ancient ocean-floor plates (Dietz, 1963; Thayer, 1969; Bird and Dewey, 1970). Support for such a belief stems from records of similar rock types on the mid-Atlantic and Indian Ocean ridges (Bonatti, 1968; Engel and Fisher, 1969). The Mesozoic Troodos ophiolite complex of Cyprus is thought to correspond to layers 2 and 3, formed beneath an ocean rise in the Tethys Ocean (Gass, 1968). The well-known cupriferous pyrite deposits of Cyprus are interbedded in a succession of basaltic pillow lavas, which, according to Gass (1968), would represent a part of layer 2 of the oceanic crust. The close genetic relation between the basaltic volcanics and the copper deposits (Vokes, 1966; Hutchinson and Searle, 1970) is evidence that concentrations of copper are present in the oceanic lithosphere.[1]

5. Manganese nodules on the ocean floors have copper contents as high as 2.5%. Analyses of nodules from the Pacific Ocean showed from 0.03 to 1.6% Cu (United Nations, 1970a).

6. Normal pelagic clays (layer 1) possess metal contents in excess of most sedimentary rocks, and average abundances of 323 ppm Cu and 18 ppm Mo have been recorded for Pacific clays (Cronan, 1969).

7. Base metals, including copper and molybdenum, of probable mantle derivation are concentrated on the East Pacific Rise (Boström and Peterson, 1966, 1969), and on the rest of the ocean rise system (Boström et al., 1969). Layer 1 sediments on the flanks of the East Pacific Rise (and presumably away from the rise beneath a blanket of normal ocean-floor sediments) have average copper and molybdenum contents of 990 ppm and 100 ppm, respectively.

8. The floor of the Red Sea, also a locus for the generation of new oceanic crust (e.g., Vine, 1966), possesses a series of metal-rich brine pools and sediments (Miller et al., 1966; Degens and Ross, 1969), possibly also charged directly from the mantle.

9. The early Pliocene Boléo bedded copper deposit in Baja California (Wilson and Rocha, 1955) may possibly represent another, albeit somewhat older, example of copper which has risen directly from the mantle, in this case related to the northwestward rotation of Baja California away from the rest of the North American continent, along segments of the East Pacific Rise mutually offset by transform faulting, a process initiated in the late Miocene to middle Pliocene (Moore and Buffington, 1968; Larsen, Menard and Smith, 1968). The high manganese- and iron oxide contents of the Boléo ores (Wilson and Rocha, 1955) provide a further similarity to the metal concentrations of the East Pacific Rise and the Red Sea.

It is also conceivable that the metals (6 ppm Cu; White, 1968) in the brines of the Salton Sea geothermal system, lying just north of the Gulf of California, and possibly underlain by the East Pacific Rise, have a comparable mantle source, although their derivation by low-grade metamorphism of clastic sediments has been proposed (Skinner et al., 1967).

[1] This suggests the possibility that other massive pyritic sulfide deposits associated with basaltic pillow lavas may have formed on the crests and flanks of ocean rise systems.

TABLE 1. Ages of Porphyry Copper and Molybdenum Deposits

Name of Porphyry Deposits or Regions	Age[a] of Porphyry Deposits	Source of Data
British Columbia		
Most porphyry copper and molybdenum deposits	Upper Triassic-Middle Jurassic Upper Jurassic Late Lower Cretaceous Upper Paleocene-Upper Eocene	White, Harakal and Carter (1968); Brown (1969)
Vancouver Island	Lower Eocene-Lower Oligocene	Carson (1969)
Western United States and Sonora		
Most porphyry copper deposits Bisbee, Arizona Ely and Yerington, Nevada Bingham, Utah Front Range porphyry molybdenum deposits	Upper Cretaceous-Paleocene Middle Jurassic Lower Cretaceous Lower Oligocene Upper Oligocene-Lower Miocene	Creasey and Kistler (1962); McDowell and Kulp (1967); Moorbath, Hurley and Fairbairn (1967); Livingston, Mauger and Damon (1968); Moore, Lanphere and Obradovich (1968); Tweto (1968); Wallace et al. (1968); Laughlin, Rehrig and Mauger (1969)
Panama		
Botija	Lower Oligocene	Ferenčić (1970)
Ecuador		
Chaucha	Upper Miocene	Müller-Kahle and Damon (1970)
Peru		
Southern Peru Michiquillay	Paleocene ? Lower Miocene	Laughlin, Damon and Watson (1968); Stewart and Snelling (in prep.)
Argentina	Tertiary (S)	United Nations (1970)
Farellón Negro, Catamarca	Upper Miocene-Pliocene (S)	Llambias (1970)
Chile	Upper Cretaceous Paleocene Upper Eocene-Oligocene Upper Miocene-Pliocene	Sillitoe, Quirt, Clark, Farrar and Neumann (in prep.)
Bougainville Island		
Panguna	? Pliocene (S)	Macnamara (1968)
Taiwan		
Chemei	Miocene or later	Po and Lee (1970)
Philippines	Tertiary (S)	Bryner (1969)
Atlas	Upper Paleocene	
West Pakistan		
Chagai district	? post-Oligocene (S)	Schmidt (1968)
Iran		
Kerman region	Upper Oligocene-Miocene (S)	Bazin and Hübner (1969)
Armenia	Upper Eocene Lower Oligocene Lower Miocene	Bagdasaryan, Gukasyan and Karamyan (1969)

[a] Time scale according to Harland, Smith and Wilcock (1964).
(S) Stratigraphic estimate.

The Distribution of Ages of Porphyry Ore Deposits

Published ages for the main groups of porphyry copper and molybdenum deposits are summarized in Table 1. It is apparent that deposits were emplaced at intervals throughout the Mesozoic and Cenozoic, with a particularly large number of deposits of late Cretaceous-Paleogene age.

It is suggested that two principal factors control the space-time distribution of porphyry ore deposits. These are: 1. the level of exposure—largely dependent on erosion rate—of a plutonic-volcanic chain; and 2. the time and location of magma generation, and the quantity of metals incorporated in magmas, on a subduction zone. It is, of course, recognized that additional variables, such as the quantity of chloride-rich fluid present during the final stages of consolidation of an intrusion, are of considerable importance.

Evidence has been advanced to show that mechanisms implicit in the new global tectonics were operative in pre-Mesozoic times (Bird and Dewey, 1970). Nevertheless, with the exception of Uzbekstan and Kazakhstan, porphyry ore deposits have not yet been definitely described from the older orogenic belts. This apparent absence is tentatively attributed to the effects of Mesozoic-Cenozoic erosion, which has been sufficient to remove the upper parts of batholiths, the loci of porphyry deposits. However, it is predicted that porphyry deposits will be encountered in parts of certain pre-Mesozoic orogens where erosion has been less severe.[8] The apparent predominance of late Cretaceous-Paleogene ages of porphyry deposits (Table 1) might also be dependent on the erosion factor, whereby many early Mesozoic deposits have been eroded away, and *some* post-Paleogene deposits have yet to be exhumed from beneath their volcanic cover. Support for this proposal is derived from conditions in Chile north of latitude 30°S, where the erosion level becomes progressively deeper from the recent volcanic chain on the Andean crest westwards to the Jurassic intrusions on the Pacific littoral (Sillitoe and Sawkins, 1971). The majority of the exposed porphyry copper deposits are Paleogene in age; Jurassic deposits, if they ever existed, have been lost by erosion, and deposits yet to be exposed may exist in the recent volcanic chain. However, the second factor, discussed below, may also be an important contributary, or even the dominant, cause of the relative abundance of Paleogene deposits in northern Chile.

If the importance of the erosion factor has been correctly evaluated, then porphyry deposits in regions with a high erosion rate, such as the southwest Pacific belt characterized by a tropical climate, could be expected to yield a predominance of particularly young ages; more radiometric dating is needed in order to test this proposal.

Explanations of many features of the distribution of porphyry copper and molybdenum deposits can be attempted in terms of the second factor, the lateral and secular pattern of magma generation, and the availability of copper and molybdenum, on subduction zones. In northern Chile, discrete post-Paleozoic intrusive episodes are manifested by a series of north-south-trending belts of batholiths and stocks. The ages of these belts decrease from Lower Jurassic near the coast to late Tertiary in the Andean Cordillera (Ruiz et al., 1965; Farrar et al., 1970). The ages of porphyry copper deposits in this region possess an analogous space-time distribution, although Jurassic deposits are as yet unknown. Thus the possibility arises that each discrete pulse of magma generation had the potential to give rise to porphyry ore deposits; the extended time interval and episodicity of porphyry deposit genesis, particularly in western North America, as reflected by the ages in Table 1, support this conclusion. The data presently available suggest that the periods of porphyry copper formation in Chile were separated by quiescent intervals with durations of about 15–25 m.y. Similar pulse-like igneous intrusion has also been reported from parts of western North America, pulses lasting approximately 10–15 m.y. and being separated by 30 m.y. intervals (Damon and Mauger, 1966; Evernden and Kistler, 1970; Gabrielse and Reesor, 1964). Such magmatic periodicity may be correlable with changes in the thermal regime on subduction zones induced by variations in the relative spreading rate or motion pattern of the plates. In this context, changes in the relative motion of plates every 10–20 m.y. in the northeast Pacific, deduced by Francheteau, Sclater and Menard (1970), might be significant.

The eastward migration of the foci of intrusion and porphyry copper emplacement evident in northern Chile, which perhaps reflects a parallel trend in the position of magma generation on the underlying subduction zone, whether or not caused by a change in its position or inclination relative to the continental margin, is less well defined in the western United States. In the latter region, Gilluly (1963) recognized an overall decrease in the age of Mesozoic-Cenozoic intrusion landwards from the continental margin, but subsequent programs of radiometric dating have shown many exceptions to this generalization. A broadly comparable pattern of eastward younging is apparent from the porphyry deposits. The belt of mid-Tertiary porphyry molybdenum deposits in the Front Range lies east of the main cluster of late Cretaceous-Paleocene porphyry copper deposits in Sonora-Arizona-New Mexico (Fig. 1), and the Lower Cretaceous porphyry deposits at Ely and Yerington, Nevada are located in the western part of this porphyry province. The great concentration of late Cretaceous-Paleocene porphyry copper deposits in the southwest United States is visualized as being due to the subduction of areas of exceptionally copper-rich oceanic crust. This contention is supported by the occurrence in the same province of unmineralized (with the exception of Bingham, Utah) mid-Tertiary stocks, which I consider to have been intruded at a time when lesser amounts of copper were available on the subjacent subduction zone. Theories invoking the extraction of copper from the continental crust or upper mantle fail to account for the concentration within a limited time period of most of these porphyry copper deposits; furthermore, post-Paleozoic stocks of all ages in the province would be expected to be similarly endowed with porphyry copper deposits. Continuing the same line of argument, large amounts of molybdenum, and only minor copper, are thought to have been available on a subduction zone vertically

[8] Pre-Mesozoic porphyry copper-type deposits have been reported from northwest of St. John. New Brunswick (Ruitenberg, Shafiqullah and Tupper, 1970), and from east-central Queensland (Cornelius, 1969), but no particulars of the occurrences were given.

beneath the Front Range in mid-Tertiary time. If the locus of magma generation on a subduction zone does not migrate systematically with time, then no clear pattern of porphyry copper ages is to be expected, a situation which may explain the apparently random distribution of ages in British Columbia.

Porphyry Copper Deposits/Plate Tectonics Interrelationships

Details of the world distribution of the evolving system of ocean rises and trenches which existed during Mesozoic and early Cenozoic times are not yet well known. Nevertheless, evidence derived from both continents and ocean basins has enabled the establishment of some aspects of plate tectonics at this time. The distribution of porphyry deposits in selected regions will now be related in general terms to the plate tectonic model:

Western South America

Evidence derived from the interpretation of magnetic anomaly patterns in the Pacific Ocean basin (Heirtzler et al., 1968; Morgan, Vogt and Falls, 1969) and the paleontological study of sediment cores in the South Atlantic Ocean basin (Maxwell et al., 1970) has demonstrated the convergence of the American and East Pacific plates along the western margin of South America (Fig. 2) since at least the late Cretaceous. This interval of active underthrusting embraces the period of formation of porphyry copper deposits in Chile, Argentina, Peru and Ecuador. In western South America, porphyry copper deposits may well be forming still beneath the active volcanic chains, for underthrusting is continuing (e.g., Plafker and Savage, 1970).

Central America and the Caribbean

The Panama trench is sediment-filled, and underthrusting is inactive (Le Pichon, 1968). The trench was abandoned in the Miocene when the pattern of sea-floor spreading in the North Pacific changed (Vine, 1966; Le Pichon, 1968), but was part of a continuous trench system bordering the west of the American continent intermittently during the Mesozoic and early and middle Cenozoic, during which time the Cerro Petaquilla and Botija porphyry copper deposits of Panama were emplaced.

In Puerto Rico, porphyry copper formation seems to have been associated with a phase of igneous intrusion of Eocene age (Mattson, 1965). Although present-day eastward movement of the American plate nearly parallels the Puerto Rico trench (Chase and Bunce, 1969; Molnar and Sykes, 1969), reconstructions of plate motions in this region indicate that underthrusting normal to the trench prevailed in mid-Cretaceous-early Tertiary times (Freeland and Dietz, 1971).

Western North America

It is now generally accepted (Vine, 1966; Yeats, 1968; Hamilton, 1969b; Page, 1970) that thrusting of the East Pacific ocean floor beneath the American plate in western North America took place at times during the Mesozoic and early and middle Tertiary. Underthrusting terminated in the western United States in the Miocene (Vine, 1966; Atwater, 1970) by the overrunning of the East Pacific Rise by the trench system. Porphyry copper and molybdenum deposits in the western United States range in age from middle Jurassic to Lower Miocene (Table 1), in excellent agreement with the time of plate convergence. In view of the absence of an active subduction zone off this coast, it is concluded that porphyry copper formation is not currently active in North America, north of the tip of Baja California, except in the Alaska Peninsula where a trench system is still active, and possibly landward of the Juan de Fuca plate. It is noteworthy that calc-alkaline volcanism, typical of convergent plate junctures, ceased in the western United States south of Oregon in the Miocene (Christiansen and Lipman, 1970; Lipman, Prostka and Christiansen, 1970).

The character of Mesozoic-middle Cenozoic plate interactions off British Columbia cannot be elucidated from studies of magnetic anomaly patterns (Atwater, 1970), but it is noteworthy that calc-alkaline volcanism in British Columbia terminated in the Eocene (Souther, 1970), at about the same time as the decline in the formation of porphyry copper deposits—excluding those on Vancouver Island.

The small Juan de Fuca plate, sandwiched between North America and the northeast Pacific (Fig. 2), has descended, and may still be descending, along a trench system paralleling the coast of Oregon, Washington and Vancouver Island (Morgan, 1968; Tobin and Sykes, 1968) giving rise to calc-alkaline magmatism in the Cascades. The Lower Eocene-Lower Oligocene porphyry deposits on Vancouver Island (Carson, 1969) do not fit well into the overall space-time distribution pattern of porphyry deposits in British Columbia, and might be ascribed to earlier activity in the vicinity of this localized compressive system, as might apparently young porphyry coppers in Washington. Extrapolation of plate motions back into the early Cenozoic (Atwater, 1970) has shown that subduction of the Juan de Fuca plate was preceded by more rapid underthrusting of the Farallon plate, at a trench which did not extend further north than Vancouver Island. However, her model approximately predicts the commencement of subduction in the Upper Eocene, mid-way through the interval of formation of the porphyry deposits.

Lineament Intersections and Porphyry Copper Deposits in Western America

The locations of several southwest North American porphyry copper deposits have been attributed to major orogen- and fault-zone intersections (Billingsley and Locke, 1941; Mayo, 1958; Schmitt, 1966). More specifically, the locations of several porphyry copper deposits (e.g., Ajo, Pima-Mission and Silver Bell) have been considered to have been influenced by elements of the west-northwest- trending Texas lineament, particularly by its intersection with the Wasatch-Jerome orogen (Mayo, 1958; Schmitt, 1966; Guilbert and Sumner, 1968; Wertz, 1970). Schmitt (1966) and Guilbert and Sumner (1968) have interpreted the Texas lineament as a continental manifestation of now-extinct transform faults in the North Pacific basin. Although several porphyry copper deposits outside of the southwest United States (e.g., Chuquicamata, Chile; Taylor, 1935) lie adjacent to important faults, none have been described as being located by major structural intersections. It is suggested that the control of porphyry copper emplacement by extinct transform faults and major structural intersections is not universally applicable, and is subordinate to a fundamental dependence on elongate zones of plate convergence. In Chile, for example, the linear, longitudinal array of porphyry copper deposits (Fig. 1) provides strong support for a subduction-zone origin, and no indication of control by structural intersections is evident (Sillitoe, unpublished). In the southwest United States, the less regular, disperse pattern of porphyry copper deposits may be explained in terms of partial fusion and consequent magma and metal generation over a greater downdip extension of the underlying subduction zone; this situation might be expected if the subduction zone were flat-dipping and imbricate as invoked by Lipman, Prostka and Christiansen (1971). It is not denied, however, that lineaments may have influenced locally the uprise of magma and included metals.

The Southwest Pacific Belt

In view of the young ages (Table 1) suggested for the porphyry copper deposits in Bougainville and Taiwan, it seems probable that their formation is linked to Benioff zones occupying positions closely similar to those currently active (Fig. 2). If the porphyry copper deposits in West Irian and Papua-New Guinea prove to be post-Miocene in age, then they would seem to be related to the southward underthrusting of the Pacific plate (Fig. 2). On the other hand, if the deposits were formed in pre-Miocene times, they would be related to a northward dipping subduction zone, which became extinct during the Miocene, by the collision of its overlying island arc (Bismarck arc) with the Australian continent (Dewey and Bird, 1970).

The Alpide Belt

The Alpide belt, in terms of the new global tectonics, is one of the least known and most complex of the compressive plate boundaries. On a global scale, the compressive forces in the Alpide belt have been attributed to relative movements between the African and Eurasian plates related to sea-floor spreading in the Central and North Atlantic Oceans (Hsü, 1971; Smith, 1971). Lithosphere was consumed along the northern and northeastern edges of the Arabian plate at the Zagros thrust zone in Iran and West Pakistan and its westerly continuation in Turkey (Dewey and Bird, 1970). Porphyry copper deposits in Iran and West Pakistan, north of the Zagros zone, were emplaced while subduction was active. The porphyry deposits in Romania, Yugoslavia and Bulgaria appear to be related to a Mesozoic-Tertiary subduction zone which, according to Dewey and Bird (1970; Fig. 14), is marked by ophiolite complexes, and extended westwards from the southern shore of the Black Sea. It might be conjectured that all the porphyry ore deposits in the Alpide belt were generated during phases of subduction related to the closure of the western Tethyan-Indian Ocean.

In the case of orogenic belts in which the collision of continents with island arcs or with other continents has contributed to their development, as in the Alpine-Mediterranean system, calc-alkaline igneous rocks and associated ore deposits may have been concealed by overthrust slices or by flysch deposits during or after collision.

Concluding Remarks

In terms of the plate tectonic model outlined above for the genesis of porphyry ore deposits, several suggestions for exploration may be made. A consideration of the distribution of Mesozoic-Cenozoic subduction zones (Fig. 2) indicates several areas as potential porphyry provinces, in addition to the orogenic belts of western America and their southward continuation into the Antarctic Peninsula. Probably the most obvious of these regions are Japan [4] and New Zealand where porphyry deposits have not yet been discovered despite extensive exploration. A recent compilation of ages of magmatism in island arcs (Mitchell and Bell, 1970) shows that Upper Cretaceous-Eocene or mid-Tertiary periods of volcanism accompanied by intrusion are represented, in addition to in island arcs where porphyry copper deposits are already known, in the Aleutians, Izu-Bonin, Sumatra-Java, Banda, North and South Celebes, New Hebrides and Fiji; these areas are considered important for porphyry copper exploration. The Lesser Antilles island arc, Kamchatka and Burma and Thailand are also considered to be promising targets.

In the Alpide zone, Turkey,[4] Greece and Afghanistan seem to be likely areas for porphyry copper discovery. A detailed analysis of the Alpine-Mediterranean orogen in terms of the new global tectonics would provide a cogent tool in the search for porphyry copper deposits. For example, if a subduction zone existed during Mesozoic and early Tertiary times at the contact of the African and European-Asian plates, as depicted by Dewey and Bird (1970; Fig. 14), then porphyry copper deposits might be expected in the north of Morocco and Algeria. If the Indus suture marks the site of the subduction zone active during the closure of the eastern part of the Tethyan-Indian Ocean (Mitchell and Reading, 1969), then porphyry deposits might be present on its northern side.

It is hoped that this model for the origin of porphyry copper and molybdenum deposits, though liable to modification in the light of further information relating to porphyry deposits and plate tectonics, will stimulate confirmatory research, and attempts to apply the new global tectonics to other classes of ore deposit. A number of the stages in the proposed model would seem to be amenable to testing by further work. Potentially important research might include age-determination and Sr^{87}/Sr^{86} studies of porphyry deposits, and isotopic and chemical studies of oceanic crustal rocks. These should be accompanied by investigations to elucidate the magmatic and metal-concentrating processes operative at the ocean rise system, and the nature of lithosphere sinking, and partial melting and metal extraction, in subduction zones.

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INSTITUTO DE INVESTIGACIONES GEOLÓGICAS,
CASILLA 10465, SANTIAGO, CHILE,
PRESENT ADDRESS:
DEPARTMENT OF MINING GEOLOGY,
ROYAL SCHOOL OF MINES,
IMPERIAL COLLEGE,
LONDON, S. W. 7, ENGLAND,
April 20; December 1, 1971.

[4] Since this paper was submitted, porphyry copper prospects have been reported from Okinawa and Turkey.

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Volcanogenic Sulfide Deposits and Their Metallogenic Significance

R. W. HUTCHINSON

Abstract

Volcanogenic sulfide deposits may be described as stratabound, lenticular bodies of massive pyritic mineralization, containing variable amounts of chalcopyrite, sphalerite, and galena in layered volcanic rocks. Often they are found to be immediately overlain by thin-bedded siliceous and iron-rich sedimentary rocks, and they are commonly underlain by extensive zones of altered, sulfide-impregnated lava. They are believed to have formed subaqueously by volcanic-fumarolic activity which occurred periodically during volcanism.

Three distinct varieties of such deposits can be distinguished by their compositions, relative and absolute ages, and rock associations. Pyrite-sphalerite-chalcopyrite bodies are found in differentiated, mafic-to-felsic volcanic rocks; pyrite-galena-sphalerite-chalcopyrite bodies occur in more felsic, calc-alkaline volcanic rocks, and pyrite-chalcopyrite bodies occur in mafic, ophiolitic volcanic rocks.

The time-tectonic-stratigraphic interrelationships of these varieties can be related to evolutionary processes of crustal development. Deposits of the pyrite-sphalerite-chalcopyrite variety are numerous, important, and best developed in Archean greenstones, suggesting that they were generated under conditions of thin "proto-crust,"possibly by degassing of as yet poorly differentiated "proto-mantle." Although they recur in younger volcanic successions, they become scarcer and smaller in later geologic time. However, their place is taken by the two other varieties which are notably rare or absent in the Archean. Thus, in Proterozoic volcanic rocks the pyrite-galena-sphalerite-chalcopyrite type appears. In Phanerozoic orogens, pyrite-chalcopyrite bodies are common in the ophiolites that typify an early stage of orogenic activity and are probably generated in oceanic ridge-rift environments during the initial stages of separation of continental crustal blocks. Both of the earlier varieties reappear in Phanerozoic belts; the pyrite-sphalerite-chalcopyrite type in early stages of subduction along continent margins, and the pyrite-galena-sphalerite-chalcopyrite type in later, more felsic, calc-alkaline volcanics that characterize somewhat later tectonism.

In addition to their obvious application in mineral exploration, these concepts may have certain scientific applicability. If, on a very broad or general basis, mineral deposits (like fauna) are products of evolutionary change, then in the absence of fossils or other correlation aids they might be used as gross-scale indicators for correlation or age determination purposes. This could be particularly applicable to highly metamorphosed Precambrian terrane throughout the world.

Introduction

A mind nimble and versatile enough to catch the resemblance of things, which is the chief point, and at the same time steady enough to fix and discern their subtle differences: . . . Francis Bacon in "The Scientific Mind."

WALDEMAR LINDGREN (1919, p. 819) considered certain massive base metal sulfide bodies to be "the most enigmatic of ore deposits." Subsequent work has substantiated Lindgren's view but still has not entirely resolved the enigma to which he referred. More recently, H. C. Gunning (1959) considered the question of the origin of these ores one of the major unresolved problems facing ore deposits geologists.

Differences among massive base metal sulfide deposits have previously been noted and discussed. Stanton (1958) was one of the earliest to present data on the different base metal contents of various deposits. Both Saager (1967) and Waltham (1968) classified the numerous deposits of the early Paleozoic Norwegian Caledonide belt into various types based mainly on their metal content, sulfide mineralogy, and texture. Gräbe (1972) attempted to explain the differences in base metal content and mineralogy by variations in surficial processes, including differences in their paleotopographic, lithofacies, and Eh environments of deposition. Gilmour (1971) related differences in the deposits to differing conditions of sedimentation or erosion, to differing petrochemistry and character of igneous activity, and to evolutionary tectonics, an approach similar to that followed here.

One purpose of this paper is to review briefly recent ideas concerning the origin of certain of these ore bodies and to summarize the broad similarities which identify them as members of a major family. A second purpose is to outline a number of important differences among these deposits that permit division of the major family into distinct types. A final purpose is to consider the significance of these differences and to interpret them in relationship to evolutionary processes of crustal tectonism and related volcanic activity.

Volcanogenic Massive Base Metal Sulfide Deposits

Latest studies of the massive, pyritic base metal sulfide deposits in volcanic rocks suggest that they are of volcanogenic origin, formed in recurrent episodes of sea-floor fumarolic activity during prolonged periods of subaqueous volcanism (Horikoshi, 1969, p. 322; Schermerhorn, 1970, p. 277–278). All members of this major volcanogenic family of ore deposits are therefore broadly related, comparable to one another, and identifiable by their common characteristics. All occur as irregularly shaped but stratabound lenses of dense pyritic sulfide containing various amounts of zinc, copper, and lead as well as important silver and gold in layered subaqueous volcanic sequences. These lenses are accompanied by two other types of mineralization. The lenses are commonly overlain by thin, siliceous iron- and manganese-rich sedimentary strata (Hutchinson and Searle, 1971, p. 199 and fig. 3; Anderson and Nash, 1972, p. 853–855) and are usually underlain by extensive zones of altered and sulfide-impregnated lava (Purdie, 1967; Suffel, 1948, p. 760). Metal zoning is prominent and the stratigraphic tops of the massive lenses are pyritic and zinc- or zinc-lead rich, whereas their stratigraphic bottoms are copper-rich and often pyrrhotitic (Price and Bancroft, 1948, p. 749–750; Scott, 1948, p. 774; Hutchinson and Searle, 1971, p. 202; Horikoshi and Sato, 1970, p. 188; Anderson and Nash, 1972, p. 855 and fig. 6). So consistent is this relationship that it apparently provides a reliable means of top determination in deformed sequences, as, for example, in the Ming deposit at the Rambler Mine in Baie Verte, Newfoundland where zinc-rich mineralization occurs on the footwall of the body (Heenan and Truman, 1972, p. 3), but where structural information suggests that the section is overturned (Kennedy, 1972). Primary breccia textures are also common, particularly at the tops of the massive bodies. These are apparently related to some form of irruptive volcanic activity, such as steam explosions (Horikoshi, 1969, p. 322) in the underlying lavas, with accompanying fragmentation of earlier deposited sulfides. This was apparently followed by redeposition involving turbidite activity, gravity-induced sedimentary slump, and downward movement on the old volcanic paleoslope (Schermerhorn, 1970, p. 276). Sharp contacts are also remarkably common and may be present at the top of the massive bodies, within the bodies between the zinc-lead- and copper-rich massive mineralization, or at the base of the massive lenses (Hutchinson and Hodder, 1970, p. 35; Hutchinson, 1965, p. 976 and 985).

These primary relationships, which characterize all members of the family, are commonly disrupted and complicated to varying degrees or even obliterated by subsequent metamorphism (Suffel et al., 1971; Vokes, 1969; p. 1130; Kalliokoski, 1965). Where metamorphism has been minimal, however, evidence for a synvolcanic or intravolcanic origin is compelling. This evidence includes soft-sediment slump and deformation structures in the overlying sedimentary strata in which the sulfides clearly behaved as heavy, primary components and also graded bedding within pyrite-rich beds (Hutchinson and Searle, 1971, p. 201). Brecciation in the tops of the massive lenses and in the underlying lavas is of "primary synvolcanic" rather than "tectonic" generation (Sinclair, 1970); primary structures, for example pillows in the nearby lavas, although fractured and brecciated, are not otherwise dislocated (Duke, 1971, p. 74). Moreover, the ores in undeformed successions, as on the north flank of the Troodos complex in Cyprus are cut by postmineralization dikes (Hutchinson and Searle, 1971, p. 201; Bear, 1960, p. 97, 99). These are chilled against ore and occasionally can be traced upward where they become feeders to fresh lavas overlying the sulfide bodies. The dikes are petrographically similar to the overlying lavas, and commonly the lavas above massive ore bodies are extremely fresh, unaltered, and barren of sulfides, as in Cyprus and at the Lake Dufault mine near Noranda,

BASE METAL TYPE	PRECIOUS METAL ASSOCIATION	ASSOCIATED VOLCANIC ROCK TYPES	TYPE OF VOLCANISM	TYPE OF SEDIMENTATION	TECTONISM	AGE	EXAMPLES
(1) Zn-Cu-pyrite	both Au (with high Cu) and Ag (with high Zn)	-fully differentiated suites of intermediate bulk composition(?); -tholeiitic to calc-alkaline -basalt-andesite-dacite-rhyolite, etc.	-initial deep, subaqueous mafic platform; with differentiation toward felsic volcanism, building domical centres	-chemical; cherts, iron formations -clastic; immature, first cycle, volcanogenic greywackes, volcanoclastics	-early eugeosynclinal-orogenic stage; -major subsidence	Archean Proterozoic(?)	Timmins, Ont. Noranda, Que. United Verde, Ariz.
Zn-Cu-pyrite	ditto	ditto	ditto	ditto	ditto; early subduction	pre-Ordovician mid-Devonian	Rambler, Nfld. W.Shasta, Calif.
(2) Pb-Zn-Cu-pyrite	mainly Ag	-intermediate to felsic calc-alkaline volcanic suites; -andesite-dacite-rhyolite-porphyry-crystal tuff, etc.	-felsic centres of explosive, pyroclastic and ignimbritic activity; subaqueous to subaerial	-epiclastic predominates; immature volcanogenic greywackes, manganiferous shales, graphitic shales and argillites, siltstones -chemical minor, cherts, iron formations -sulphate gangues common	-later eugeosynclinal-orogenic stage; infilling with uplift balances subsidence(?)	Proterozoic Ordovician	Mt. Isa, Queensland Errington, Vermilion, (Sudbury Basin) Bathurst, New Brunswick
Pb-Zn-Cu pyrite	ditto	ditto	ditto	ditto	ditto; later subduction	Triassic Tertiary	E.Shasta, Calif. Kuroko, Japan
(3) Cu-pyrite	mainly Au	-poorly differentiated mafic-ultramafic(ophiolitic) suites; -tholeiitic -basaltic pillow lavas, serpentinite, etc.	-deep subaqueous, quiescent fissure eruptions	-chemical predominates; cherts, ironstones, manganstones -clastic insignificant	-early stage of continental plate rifting; tension, separation, graben	l.-Ordovician u.-Cretaceous Juro-Cretaceous Cret.-Eocene	W.Newfoundland Cyprus Island Mountain, California Phillipines

FIG. 1. Comparative Chart: Some geological characteristics of different volcanogenic sulfide deposits.

Quebec (Hutchinson and Searle, 1971, p. 201; Purdie, 1967).

Differences Among Volcanogenic Massive Base Metal Sulfides

Despite these broad, unifying similarities of all members of the family, there are readily recognizable differences among the various deposits which serve to divide the family into three distinct types. The distinguishing geological features include: first, the base metal and precious metal assemblages present in the ores themselves; second, the petrographic character of the associated volcanic rocks and the inferred nature of the volcanism that deposited them; third, the nature and proportion of intercalated sedimentary strata within the host volcanic succession; fourth, the tectonic environment in which sulfide bodies, volcanic rocks, and sedimentary rocks alike were originally laid down; and finally, but not least, their greatest abundances through time and also their ages relative to one another within Phanerozoic orogenic belts. These criteria are tabulated in Figure 1 which also lists examples of the types.

The zinc-copper type

The earliest type of massive, volcanogenic base metal sulfide deposits recognizable using these criteria is rich in either or both zinc and copper. They contain only minor lead but produce both gold and silver, usually accompanying zinc-rich sulfides; gold being relatively more important in copper-rich deposits (Table 1).

Their host volcanic rocks include broadly differentiated suites that span the entire compositional range from basalt to rhyolite. The exact petrochemical nature and derivation of these lavas is not certain despite considerable investigation. Wilson et al. (1965, p. 170–174) considered these volcanics to be of calc-alkaline affinity, as did Goodwin and Ridler (1970, p. 22). Other work (Baragar, 1968) suggests that they are of tholeiitic descent. Baragar and Goodwin (1969, p. 140) suggest that the "average" Archean volcanic rock is slightly more salic than tholeiitic basalt and more mafic than andesite, thus perhaps resembling island-arc tholeiites more than oceanic tholeiites (Jakes and White, 1972; Gill, 1970). Further clarification of these relationships is necessary, but present data suggest that these volcanic rocks may have affinities to both tholeiitic and calc-alkaline families with chemical properties intermediate between the two.

Whatever their petrogenesis, these lavas were apparently deposited in successions ranging up to 40,000 feet in total thickness (Goodwin, 1968, p. 3). At the base of the successions are broad mafic platforms, generally of basaltic composition, probably formed mainly by widespread fissure eruptions from deep fractures (Goodwin and Ridler, 1970, p. 23). Subsequent volcanism became increasingly domal with differentiation toward the later felsic stages (Goodwin, 1965). The intercalated sedimentary rocks include chemical sediments, particularly cherts and the various facies of iron-formation, siliceous tuffs, and immature first-cycle volcanogenic gray-

TABLE 1. Base and Precious Metal Content of Three Types of Volcanogenic Massive Base Metal Sulfide Deposits

Type, District, Name		%Pb	%Zn	%Cu	Ag	Au	Ag/Au	Remarks—Reference
1. Zn-Cu Type								
Horne Mine; Noranda, Quebec		n.s.	n.s.	2.3	n.s.	0.18	very low	Approx. grade 60 m.t. ore produced.
Quemont; Noranda, Quebec		n.s.	2.69	1.52	0.92	0.17	5.4	Ore reserve estimate, 9 m.t. 1947; Taylor, 1957, p. 406.
Lake Dufault "A"; Noranda, Quebec		n.s.	7.2	4.0	2.2	0.03	73.3	Initial ore reserve estimate 2.3 m.t.; Purdie, 1967, p. 53.
Waite; Noranda, Quebec		n.s.	4.47	7.11	0.70	0.03	23.3	Total production 1.1 m.t.; Price and Bancroft, 1948, p. 748.
Amulet Lower A; Noranda, Quebec		n.s.	5.69	5.13	1.41	0.041	34.4	Millhead ore assays, 2.1 m.t.; Suffel, 1948, p. 757.
Amulet C; Noranda, Quebec		n.s.	14.30	2.62	4.19	0.020	209	Millhead ore assays, 240,000 t.; Suffel, 1948, p. 757.
Amulet F; Noranda, Quebec		n.s.	10.88	4.28	1.48	0.017	87.0	Millhead ore assays, 150,000 t.; Suffel, 1948, p. 757.
Mattagami Lake; Matagami, Quebec		n.s.	10.4	0.69	1.13	0.014	81.0	18.7 m.t. ore reserves 1966; Canadian Mines Handbook 1967–68, p. 213.
Manitou Barvue: Val d' Or, Quebec	a)	0.38	4.74	0.03	2.57	0.034	76.	Prod. from 6.1 m.t. zinc-lead ore; Ramsay and Swail, 1967, p. 19.
	b)	n.s.	n.s.	0.9	0.068	0.012	5.5	Prod. from 3.1 m.t. copper ore; Ramsay and Swail, 1967, p. 19.
Mattabi; Sturgeon Lake, Ontario		0.84	7.67	0.91	3.13	0.007	447	12.8 m.t. ore reserves, 1970; recoverable lead; Canadian Mines Handbook, 1972–73.
Flin Flon; Flin Flon, Manitoba		n.s.	4.24	2.99	1.25	0.089	14.0	26 m.t. ore reserves, 1946; Geology Staff, Hudson Bay Mining and Smelting Co., Ltd., and Stockwell, 1948, p. 295.
Cuprus Mine; Flin Flon, Manitoba		n.s.	6.4	3.75	0.84	0.038	22.1	Avg. grade 510,000 t. ore mined; Geological Staff, Hudson Bay Mining and Smelting Co., Ltd., 1957, p. 253.
Stall Lake Mine Snow Lake, Manitoba		n.s.	0.5	5.2	0.31	0.44	7.0	2.1 m.t. reserves, 1968; Coats et al., 1970, p. 971.
Rod Deposit No. 2 Snow Lake, Manitoba		n.s.	2.88	7.1	0.47	0.05	9.4	0.3 m.t. reserves, 1968; Coats et al., 1970, p. 971.
Chisel Lake Mine Snow Lake, Manitoba		0.7	12.00	0.39	1.23	0.49	25.1	3.2 m.t. reserves, 1968; Coats et al., 1970, p. 971
Jerome District; Arizona		n.s.	?	8.5	2.50	0.068	37	Avg. of 5 different mines, 19.2 m.t. produced 1908–51; Anderson and Creasey, 1958, p. 100.
UVX Mine; Jerome, Arizona		n.s.	n.s.	?	<2.0	0.04	<50	Avg. grade 3 m.t. ore; Anderson and Creasey, 1958, p. 136.
West Shasta District; California								All data from Kinkel et al., 1956.
Balaklala		n.s.	1.3	2.8	1.0	0.028	35.8	1.2 m.t. production, several ore bodies.
Early Bird		n.s.	n.s.	3.4	2.0	0.034	58.8	35,000 t. prod.
Golinsky		n.s.	8.9	3.57	4.61	0.134	34.4	3,000 t. prod.
Iron Mountain		n.s.	3.50	2.00	1.00	0.02	50	380,000 t. flotation ore.
Iron Mountain		n.s.	2–5	7.50	1.00	0.04	25	1.6 m.t. prod. Old Mine Orebody.
Keystone		n.s.	8.0	6.0	2.7	0.06	45	122,000 t. prod.
Mammoth		n.s.	4.20	3.99	2.24	0.038	59	3.3 m.t. copper ore 1905–25.
Mammoth		n.s.	21.10	2.40	5.79	0.078	74.2	84,000 t. Zn ore 1914–15.
Shasta King		n.s.	?	2.29	1.01	0.034	30	69,000 t. prod. 1918–19.
Rambler Mine; Baie Verte, Nfld.	a)	n.s.	2.16	1.30	0.85	0.15	5.6	Main Zone prod. 440,000 t.; Heenan and Truman, 1972, p. 2.
	b)	n.s.	n.s.	2.67	0.57	0.08	7.1	Ming Zone ore reserves 905,000 t.; Heenan and Truman, 1972, p. 7.

TABLE 1.—(*Continued*)

Type, District, Name		%Pb	%Zn	%Cu	Ag		Au	Ag/Au	Remarks—Reference
2. Pb-Zn-Cu-Ag Type									
Errington Mine; Sudbury Basin		0.75	3.24	1.02	1.49		0.017	88	Ore reserves 7.5 m.t. after 15% dilution; Martin, 1957, p. 365.
Vermilion Lake Mine; Sudbury Basin		1.10	4.56	1.10	1.78		0.020	89	Ore reserves 2.8 m.t. after 20% dilution; Martin, 1957, p. 365.
Mt. Isa; Queensland	a)	8.3	7.2	n.s.	6.3		n.s.	very high (>300?)	19.2 m.t. lead ore mined to 1964; Bennett, 1965, p. 233.
	b)				130.	g	0.03	4,300	Lead ore; personal commun., R. L. Brown, Mt. Isa Mines Ltd., 1972.
	c)			3.5	8	g	0.06	133	Copper ore; personal commun., R. L. Brown, Mt. Isa Mines Ltd., 1972.
Iron King; Prescott, Arizona		2.50	7.34	0.19	3.69		0.123	30	Based on metal recovered from 5 m.t. prod.; Gilmour & Still, 1968, p. 1241.
Buchans; Newfoundland		7.85	15.5	1.45	3.52		0.05	70	Average, all ore bodies to 1962; Swanson and Brown, 1962, p. 623.
Heath Steele; New Brunswick	a)	2.90	7.10	1.10	3.20		0.02	160	Ore reserves 4.2 m.t., 1965; Pb-Zn ore; Canadian Mines Handbook, 1965.
	b)	1.20	3.50	1.30	1.90		0.02	95	Ore Reserves 3.0 m.t., 1965; Cu ore.
Brunswick #6		2.20	6.01	0.39	2.02		n.s.	high (>100?)	6 m.t. ore oreserves in pit, 1972; Canadian Mines Handbook, 1972–73, p. 57.
Brunswick #12	a)	3.52	8.60	0.30	2.47		n.s.	high (>100?)	16.8 m.t. ore reserves, above 1,500 level, 1971.
	b)				1.85		0.011	165	Avg. of 28, 10-foot samples across ore body, ddh. 1,275; personal commun. P. A. Kavanagh.
	c)	n.s.	n.s.	1.11	0.85		n.s.	?	9.4 m.t. high copper-ore reserves.
Afterthought; E. Shasta, California		2.17	16.5	3.23	5.55		0.03	185	166,000 t. produced; Albers and Robertson, 1961, p. 81.
Bully Hill-Rising Star; E. Shasta, California		n.s.	2.2	4.2	3.82		0.066	58	580,000 t. produced but high Zn ore left in mine: figures for Ag and Ag/Au ratio therefore too low; Albers and Robertson, 1961, p. 90.
Blue Moon; Foothill Belt, California		0.48	12.29	0.37	3.75		0.062	61	56,000 t. prod.; Eric and Cox, 1948, p. 145.
Penn Mine; Foothill Belt, California		5.58	7.89	2.05	2.37		0.07	34	84,000 t. prod. 1943–1946; Heyl et al., 1948, p. 65.
Suffield; E. Tps., Quebec		0.59	6.45	1.28	2.40		0.018	134	Avg. grade 1 m.t. reserve, 1951; Carrière, 1957, p. 466.
Matsumine, Japan									
black ore		high	high	med.	209.3	(g)	1.27	165	Average content of precious metals in four ore varieties; Dowa Mining Co., 1970, p. 44.
sub-black ore		med.	med.	high	64.3	(g)	0.66	97	
yellow ore		low	low	high	27.9	(g)	0.47	60	
pyritic ore		low	low	low	9.8	(g)	0.32	31	
Lake George Mines; Captains Flat, N.S.W.		6.	10.	0.67	1.80		0.055	33	Avg. grade 4 m. tons 1937–1962; Glasson and Paine, 1965.
Blow Orebody; Mt. Lyell, Tasmania		n.s.	n.s.	1.28	2.67		0.065	41	Avg. grade 5.5 m. tons prod. to 1963; Solomon and Elms, 1965, p. 482.
Tasman and Crown-Lyell Extended Co. Lode; Mt. Lyell, Tasmania		28.	20.	0.5	17.		n.s.	very high	"Typical grade," Solomon and Elms, 1965, p. 483.
Rosebery and Hercules deposits; Tasmania		6.	20.	0.95	6.		0.1	60	Avg. grade 5.2 m. l. tons prod. from both deposits to 1963; Hall et al., 1965, p. 485.

TABLE 1—(*Continued*)

Type, District, Name		%Pb	%Zn	%Cu	Ag	Au	Ag/Au	Remarks—Reference
3. Cupreous Pyrite Type								
Skouriotissa; Cyprus					2.02	0.33	6.1	Assay of nearly pure native sulfur, 915 level; Wilson and Ingham, 1959, p. 174.
Mavrovouni; Cyprus		n.s.	0.4	4.2	0.25	0.025	10.0	Average ore grade, 1947; Wilson and Ingham, 1959, p. 148.
Mavridia; Cyprus	a)	0.004	0.4	0.58	6.0 g	0.8	7.5	Low-grade ore; Bear, 1963, p. 64.
	b)	0.015	0.33	1.1	5.5 g	2.0	2.75	High-grade ore; Bear, 1963, p. 64.
Limni; Cyprus		n.s.	n.s.	1.41	1.59 d	1.97	8.1	50,000 tons semi-oxidized ore; Bear, 1963, p. 74.
Ergani Maden; Turkey		n.s.	n.s.	10.02	21.9 g	1.4	15.6	Typical analysis of high-grade ore; Griffitts et al., 1972, p. 711.
Weiss; Turkey	a)	n.s.	3.2	5.2	2.3 p	1.0	2.3	Analysis of ore; Cagatay, 1968.
	b)				0.592	0.083	7.1	Average; ten analyses of massive ore.
Kure; Turkey		n.s.	0.01	2.37	11.0 g	2.2	5.0	Typical massive ore; Bailey et al., 1967, p. 58.
Island Mtn., California		n.s.	n.s.	3.39	1.092	0.651	1.7	Average grade 132,000 tons of ore 1915–1930; Stinson, 1957, p. 25.
Hixbar; Philippines		n.s.	n.s.	9.36	0.98	0.11	8.9	Average grade 63,000 tons pre 1941; Bryner, 1969, p. 652.

Notes: values in precious metals are in troy oz. per ton unless otherwise noted as d—pennyweights per ton; g—grams per ton; p—parts per million; n.s.—not stated; ?—data uncertain but metal present.

wackes (Goodwin and Ridler, 1970). The tectonic environment in which these rocks and sulfides were deposited clearly involved major subsidence on a very large scale with deposition of two or more cycles of the differentiated mafic to felsic subaqueous volcanic rocks (Ridler, 1970). This apparently results from large scale "eugeosynclinal" volcanism, presumably occurring in a very early stage of a tectonic cycle.

The best examples of the zinc-copper type are of Archean age. They include the great base metal deposits of the Keewatin greenstone belts in the Superior Structural Province of the Canadian Shield, including the Horne Mine at Noranda, the Kidd Creek deposit at Timmins, the Mattagami Lake Mines ore body at Matagami Lake, and many others. It is interesting, as indicated in Figure 1, that ore bodies of this type are less numerous in Aphebian rocks. Although possibly present at Flin Flon, Manitoba (Sangster, 1972) and Jerome, Arizona (Anderson and Nash, 1972), the reliability of the Pb-isotopic age dates on which the Aphebian age of the Flin Flon ores is mainly based has very recently been questioned (Slawson and Russell, 1973). Hence the Flin Flon deposits, and perhaps those at Jerome as well, may be of older Archean age, metamorphosed during a later Aphebian orogeny. There are no known examples of the zinc-copper type in still younger Proterozoic rocks.

The type reappears, however, in some older Phanerozoic (i.e. Paleozoic) orogenic belts. Recurrent examples are the deposits of the West Shasta District in California, which lie in Middle Devonian volcanic rocks (Kinkel et al., 1956, p. 16) and perhaps the three ore bodies of the Rambler Mine in western Newfoundland (Heenan and Truman, 1972), although the exact age and petrochemical affinities of the volcanic rocks in the Fleur de Lys Supergroup that are host to the Rambler ore bodies (Kennedy et al., 1972; Church, 1969) are uncertain and require further investigation. In Australia, the deposits near Bathurst, N.S.W. occurring in basaltic to andesitic volcanic and pyroclastic rocks of Upper Ordovician to Lower Silurian age (Stanton, 1955, p. 684) appear to be examples of this type in the Brisbane-New England eugeosyncline. Notwithstanding its recurrence, examples of the zinc-copper type in early Phanerozoic orogenic belts are both rarer and smaller than their Precambrian analogues, and there are apparently no examples in volcanic rocks of still younger Phanerozoic age.

The lead-zinc-copper-silver type

The second type of volcanogenic massive base metal sulfide deposits is rich in both lead and zinc, contains minor copper, and yields important silver rather than gold, as might be anticipated from its significant lead content (Table 1). Carbonate-rich gangue is abundant in older examples of these ores (Bennett, 1965, p. 241; Martin, 1957, p. 367), and sulfate gangues, including both bedded gypsum-anhydrite strata and barite, are often found in younger deposits (Matsukuma and Horikoshi, 1970, p. 165, 174–175; Swanson and Brown, 1962, p. 624; Smith, 1948, p. 121). Volcanic rocks associated with this type are considered more felsic in overall composition than those accompanying the former. Basalts are usually rare in the vicinity of the ore bodies, although they may be present deeper in the same section or in laterally correlative rocks (McAllister, 1960, p. 88). Intermediate and felsic lavas predominate near the ore bodies, along with porphyritic subvolcanic intrusions (Swanson and Brown, 1962, p. 620–621), breccias (Martin, 1957, p. 364; Swanson and Brown, 1962, p. 621), and various pyroclastic rocks (McAllister, 1960, p. 96), all of which are distinctive and abundant. Broadly speaking the petrochemical affinities of these volcanic rocks appear to be calc-alkaline (Strong, 1972b; Tatsumi et al., 1970, p. 35–36).

The explosive volcanism (Horikoshi, 1969; McAllister, 1960, p. 89) that generated these rocks apparently took place in domical centers that were perhaps briefly and locally subaerial. Endogenous rhyolite domes containing both felsic lavas and their porphyritic subvolcanic equivalents (Matsukuma and Horikoshi, 1970, p. 157–158; Smith and Skinner, 1958, p. 152) appear to have been a prominent product of volcanism. Epiclastic sedimentation during deposition of these sequences was apparently a more extensive process than in the case of the zinc-copper type. As a result, epiclastic sedimentary rocks predominate over the chemically deposited cherts and iron-formation and the volcaniclastic sedimentary rocks, although these are still present (McAllister, 1960, p. 95, 97; Martin, 1957, p. 366; Newhouse, 1931, p. 403). The volcaniclastic rocks include as before tuffs and volcanogenic graywackes, but in this environment manganiferous and carbonaceous or graphitic argillites and shales are also common and abundant (Bennett, 1965; Smith and Skinner, 1958, p. 152; Martin, 1957, p. 367). In fact, the ores of this type are often found in these fine, impure clastic sedimentary rocks near the felsic volcanic rocks rather than in the rocks of the endogenous dome itself. Siltstones, coarser epiclastic rocks, and rarely limestones are also associated with deposits of this type (Swanson and Brown, 1962, p. 619–620; Smith, 1948, p. 114).

The tectonic environment of deposition was again predominantly eugosynclinal and volcanic but apparently represented a somewhat later stage than that of the zinc-copper type. As well as volcanism, this later-stage tectonism apparently involved considerable sedimentation both by degradation of the volcanic dome and from a bordering cratonic area. Presumably, infilling of the subsiding trough by both volcanic and sedimentary processes in general kept pace with and compensated for subsidence, resulting in very shallow subaqueous or even partially subaerial volcanism. Initial uplift of the margins of the subsiding trough may have accompanied its central subsidence, accounting for the greater abundance of craton-derived epiclastic sedimentary material.

This type appears first in Proterozoic rocks, and the Errington and Vermilion Lake deposits within the Sudbury Basin of Ontario (Martin, 1957) and perhaps those at Mt. Isa in Queensland, Australia (Bennett, 1965) are examples. It is equally important in earlier Paleozoic time, and perhaps the best examples are the numerous large deposits of the Bathurst District in northern New Brunswick (McAllister, 1960; Smith and Skinner, 1958) and at Buchans, Newfoundland (Swanson and Brown, 1962; Newhouse, 1931) occurring in volcanic rocks of Ordovician age. In Australia, examples are the important deposits at Captains Flat, N.S.W. (Table 1), which occur in andesitic to rhyolitic and porphyritic volcanic-pyroclastic rocks of Upper Silurian age (Glasson and Paine, 1965, p. 424). This type too may become rarer and the deposits smaller in younger rocks, although recurrent examples are listed in Figure 1. They include the ore bodies of the East Shasta District (Albers and Robertson, 1961) and the Foothill Copper Zinc Belt (Heyl, 1948a) in California, of Triassic and Jurassic ages, respectively, and the Kuroko district of Japan (Matsukuma and Horikoshi, 1970; Horikoshi, 1969), an extremely important Tertiary example.

The cupreous pyrite type

Cupreous pyrite deposits represent the third type of volcanogenic massive base metal sulfide ores. These deposits contain virtually no lead, only minor zinc, and carry significantly higher proportions of gold relative to silver than the others (Table 1). The associated igneous rocks are of mafic-ultramafic composition and of ophiolitic affinity. Lenses of massive pyrite-chalcopyrite ore are contained in spilitized basaltic pillow lavas which are apparently the product of relatively deep water, quiescent, fissure eruptions on the sea floor (Bear, 1960, p. 60), presumably like those occurring along modern oceanic rift-ridge systems (Chase, 1969). These lavas resemble modern oceanic tholeiites (Smitheringale, 1972; Gass, 1968, p. 39). Unlike both of the other types, felsic volcanic rocks are rare or absent. The associated sedimentary rocks are volumetrically insignificant and include predominantly chemically deposited radiolarian cherts, ironstones, and manganstones (Bear, 1960, p. 20; Wilson and Ingham, 1959, p. 26). Very few clastic or pyroclastic strata are present, although aquagene tuffs and hyaloclastic and pillow breccias are intercalated with the basalt flows (Carlisle, 1963; Smitheringale, 1972, p. 576; Bailey et al., 1967, p. 32–34). The tectonic depositional environment was different from both former types. The ophiolitic lava sequence is much thinner than in the two former types, probably less than 5,000 ft. in Cyprus (Bear, 1963, p. 17; Bear, 1960; p. 60), and about 10,000 in western and north central Newfoundland where it may be abnormally thick (Upadhyay et al., 1971, p. 33). This suggests minor subsidence, probably tension-induced and the result of crustal attenuation and rifting (Gass, 1970), rather than major eugeosynclinal subsidence as in the other types.

Examples of this type are all Phanerozoic, appearing first in early Paleozoic ophiolite complexes, like the cupreous pyrite deposits in basaltic lavas of the Notre Dame Bay area (Peters, 1965; Smitheringale, 1972; Riccio, 1972) and of the Bay of Islands (Duke, 1971) in north central and western Newfoundland, respectively. The best examples of this type, however, are Mesozoic and include the numerous deposits of Cyprus (Bear, 1963) and those at Küre (Bailey et al., 1967, p. 31) in Turkey. The Island Mountain copper deposit in Franciscan rocks of the Trinity Alps in California (Stinson, 1957) and other similar occurrences in southwestern Oregon (Shenon, 1933) may be examples of Mesozoic age in the North American Cordillera. Deposits of this type in Tertiary rocks are those at Ergani Maden in Turkey (Griffitts et al., 1972, p. 703) and perhaps the Balabac, Barlo, and Hixbar deposits in the Philippines (Johns, 1963; Bryner, 1967; 1969).

The recently discovered, metal-rich, gel-like precipitates beneath heated saline brines on the floor of the Red Sea deeps (Degens and Ross, 1969) may be a modern example and analogue of this type (Hutchinson and Searle, 1971, p. 203), although where these gels have been sampled they are richer in zinc than most, but not all (Bear, 1963, p. 40), of the examples cited.

Comparison of Subvarieties and Interpretation

If, as suggested, these ores are all members of a major family distinguishable by its coeval volcanic affiliations, then differences among the three types should be relatable to differences in volcanic petrochemistry. Moreover, if the age relationships are correct or even approximately so, the variations among the ores should be explainable by evolutionary changes in volcanism, its style, and its products. These changes in volcanism, in turn, result presumably from evolutionary tectonism. Conversely, it follows that the ores themselves should provide another clue to the processes of evolutionary volcanism and tectonism that have affected the earth's crust through time and during the development of Phanerozoic orogenic belts.

Based on all the foregoing geological relationships, it is suggested that the zinc-copper type was formed under relatively thin, primitive crustal conditions. The associated volcanic rocks were derived from magmatic activity in an underlying, relatively thick, and poorly differentiated upper mantle that may have been of intermediate "average composition" (Baragar and Goodwin, 1969, p. 140). These conditions are diagrammatically represented in Figure 2. Magmas originating in this mantle during prolonged periods of Keewatin volcanism therefore underwent a complete cycle (or cycles) of differentiation. In this manner, the characteristic Archean volcanic successions were laid down, commencing with widespread basaltic lavas of the mafic platform and progressing toward felsic, domal activity in the later differentiation stages (Goodwin, 1965; 1968). Judging from the extent and repetition of these successions in Archean terrane throughout the world (Anhaeuser et al., 1969, p. 2192), these conditions must have prevailed over much of the earth's surface during Archean time, and the large and numerous zinc-copper sulfide deposits that are so characteristic of these rocks in Canada may represent the products of extensive degassing (Cloud, 1972, p. 543) of the primitive upper mantle on a scale not since repeated. Presumably it was by these processes, combined with subsequent Kenoran orogeny and its attendant granitic intrusion and gneissification, that earth generated a thick, supracrustal sialic plate by the close of Archean time.

It is suggested that in Proterozoic time eruptive volcanism breaching this thickened continental plate became rarer and more restricted, whereas epiclastic sedimentation involving degradation of the plate became more extensive and important. However, in deeper, eugeosynclinal portions of what were perhaps broad, rift-controlled basins (Card and Hutchinson, 1972, p. 69) or eugeosynclinal belts marginal to old Archean cratonic blocks (Hoffman et al., 1970), volcanism from the now differentiated, ultramafic mantle produced tholeiitic and basaltic lavas (Dimroth et al., 1970, p. 45; Frarey and Roscoe, 1970, p. 146). Locally, in shallower shelves flanking these basins and geosynclines, volcanism, presumably derived from anatectic melting of the sialic plate, occurred in centers of more felsic, explosive, and possibly even subaerial eruption (Stevenson, 1971; Williams, 1957; Sauvé, 1953). The latter conditions are diagrammatically illustrated in Figure 3. Here the resulting volcanic sequences were of more felsic, calc-alkaline

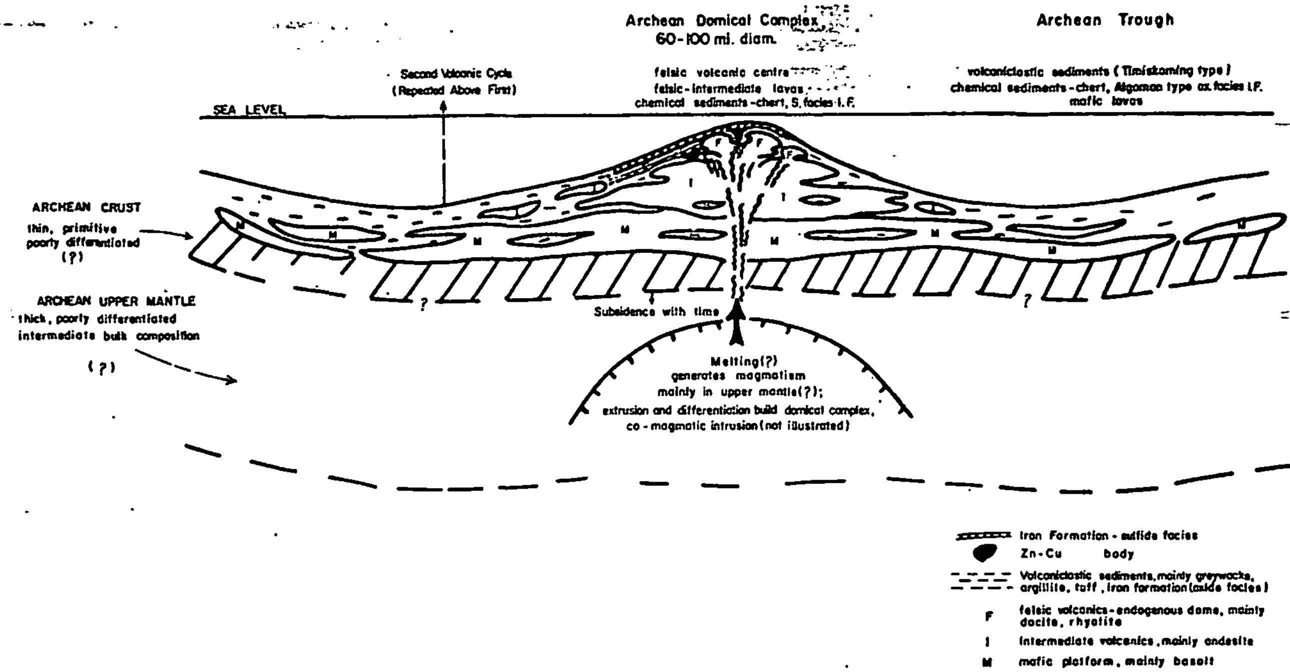

FIG. 2. Diagrammatic illustration: Zn-Cu type of volcanogenic massive sulfide deposit in Archean tectonic setting.

composition and contained a greater proportion of fine epiclastic sedimentary strata which graded laterally into increasingly coarse clastic shelf-facies rocks deposited at the margins of the subsiding basins (Dimroth et al., 1970, p. 67–68). The filling in of these basins and troughs by combined volcanism and

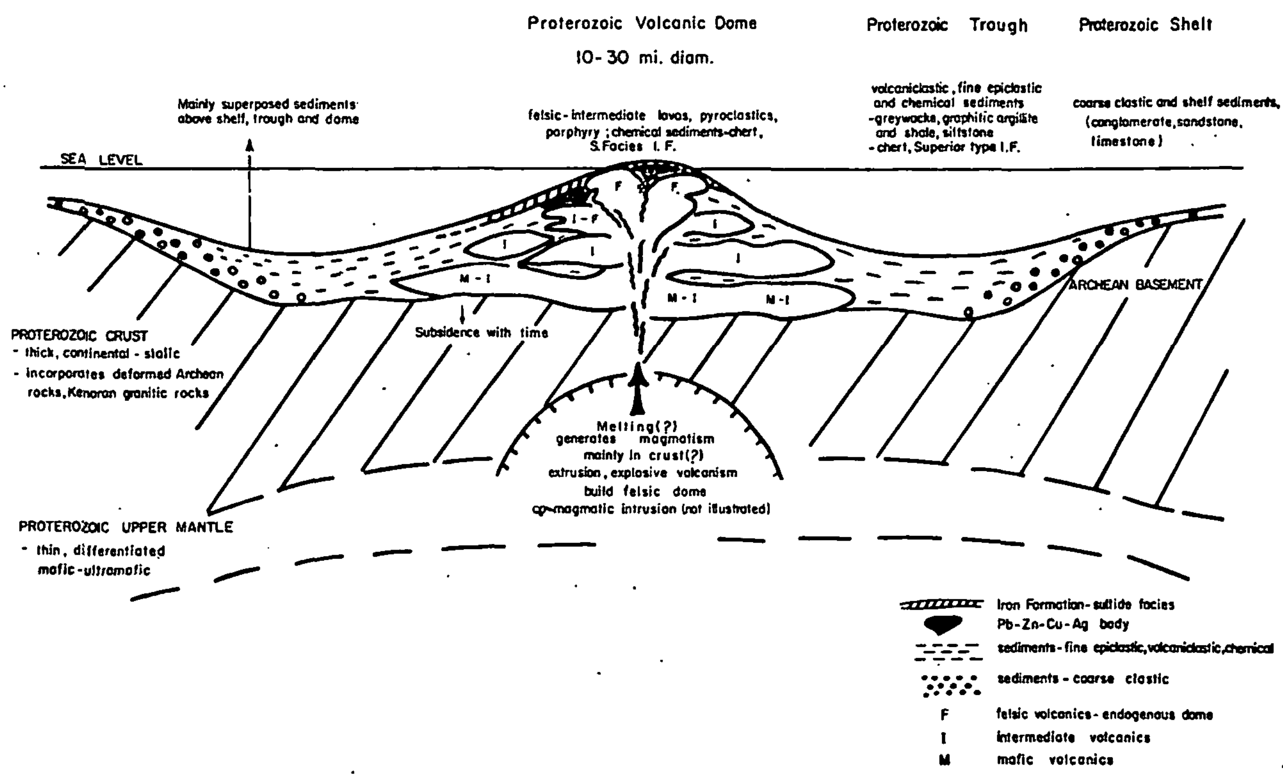

FIG. 3. Diagrammatic illustration: Pb-Zn-Cu-Ag type of volcanogenic massive sulfide deposit in Proterozoic tectonic setting.

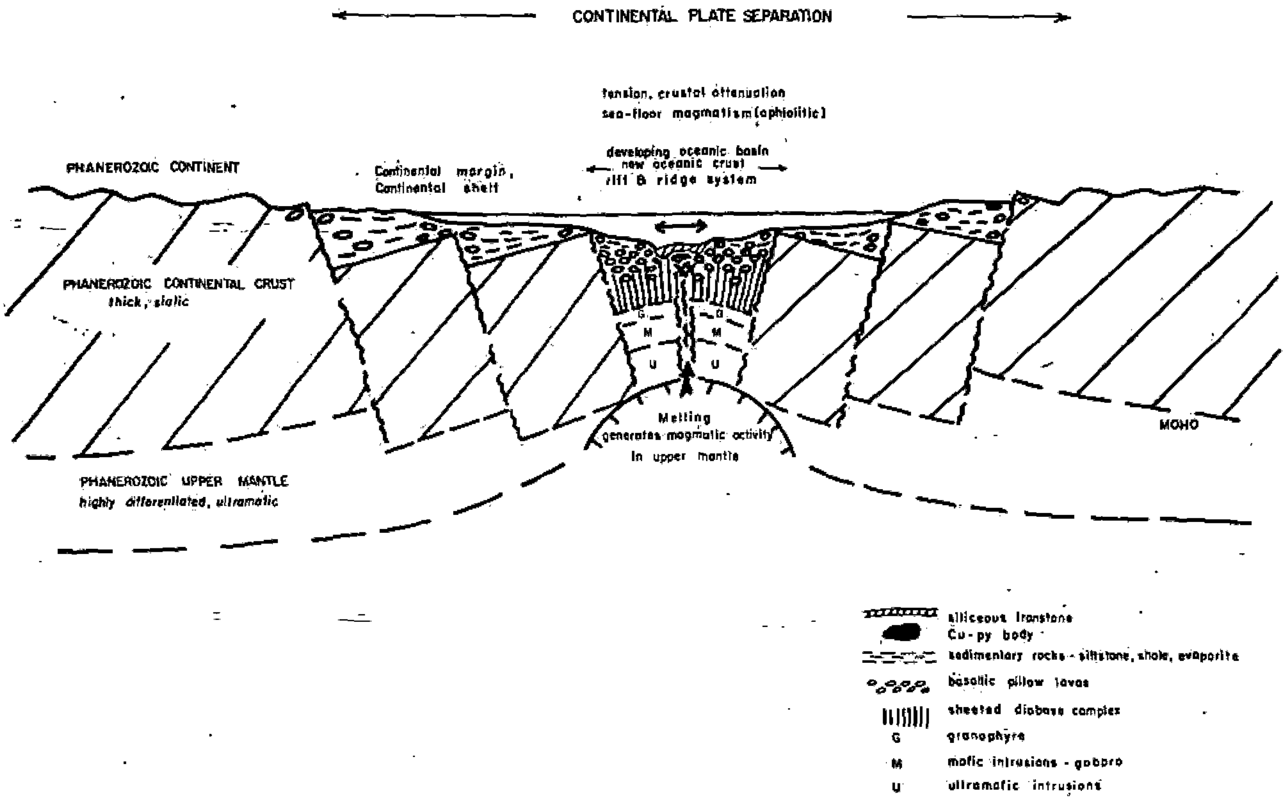

FIG. 4. Diagrammatic illustration: Cupreous pyrite type of volcanogenic massive sulfide deposit in Phanerozoic, separating oceanic rift environment.

sedimentation more or less kept pace with subsidence. These conditions were favorable for the generation of the second lead-zinc-copper-silver type of volcanogenic massive base metal sulfide deposits.

Finally, it is suggested that in still later Phanerozoic time, plate-tectonic activity in its presently known form (Isacks et al., 1968), accompanied by crustal attenuation, rifting, and separation of the rifted plates, created conditions favorable for generation of the third type of volcanogenic massive base metal sulfide deposits. The conditions are illustrated diagrammatically in Figure 4. Cupreous pyrite deposits formed under these tectonic conditions along the developing oceanic rift-ridge systems (Hutchinson and Searle, 1971, p. 198; Smitheringale, 1972, p. 586) from sea-floor fumarolic activity associated with ophiolitic magmatism. The magmatism included eruptive activity, which deposited the pillowed, spilitic basalts and intercalated radiolarian cherts and siliceous ironstones, subvolcanic feeder-dike intrusion which formed the sheeted diabase complexes, and deeper intrusion of layered gabbro and ultramafic rocks (Strong, 1972a; Upadhyay et al., 1971, p. 28, 32; Pantazis, 1967; Dewey and Bird, 1971). Serpentinites are also formed in this setting (Wyllie, 1969), and tholeiitic basalts derived from the now highly differentiated and ultramafic mantle (Engel et al., 1965) predominate in this newly formed oceanic crust. The common association of all these oceanic crustal rock types with cupreous pyrite deposits is thereby explainable. Subsequently, complex plate interactions (Bird and Dewey, 1970, p. 1044) and underthrusting of this oceanic crust by sialic blocks (Hsu, 1970; Gass, 1967, p. 130–131) occurred along plate boundaries. Slices of oceanic crust were thereby thrust onto the continent margins where they were preserved, along with their cupreous pyrite ores, from destruction in subduction zones.

The reappearance of both earlier types in Phanerozoic orogenic belts indicates that the mechanics of plate interaction somehow recreate the earlier tectonic environments, thus permitting their local regeneration. Current investigations of plate boundary relationships (James, 1971) and of magmatism in continent margin-island arc environments (Jakes and White, 1971; Kuno, 1966) suggest that the permissive developments occur in subduction zones where continental crustal plates override descending oceanic crustal plates. Successive stages of this process are apparently involved in regenerating the two differing types. The distinction between the two in this tectonic setting may be most useful for classification and for recognition of various evolutionary stages, insofar as the broad process of subduction is presumably continuous in nature.

Nevertheless, extensive magmatism accompanies the earlier stages of subduction during initial infilling of unstable, subsiding eugeosynclinal troughs along continental margins. These early eugeosynclinal troughs may have formed above subduction zones along which oceanic crustal plates were underthrust beneath island-arc systems (Dewey and Bird, 1970, figs. 2c, d, 12a). These arcs were presumably separated by marginal ocean basins from the advancing continent margin itself, possibly like those rimming the western Pacific today (Karig, 1971), and it is interesting that zinc-rich fumarolic volcanism is active in such an environment in New Britain (Ferguson and Lambert, 1972). Whether from anatectic melting and consumption of the descending oceanic plate in the Benioff zone or from partial melting of mantle between the top of the underthrust plate and the crust, this magmatism generates extensive basaltic-andesitic volcanic successions (James, 1971, p. 3343, 3342, fig. 10). These conditions are shown diagrammatically in Figure 5. This environment resembles the primitive Archean conditions already discussed, involving thin and unstable crust, major subsidence, and extensive volcanism of intermediate overall composition, perhaps spanning the entire compositional range from mafic to felsic with progressing differentiation. These tectonic conditions favor generation of the zinc-copper type of volcanogenic massive base metal sulfide deposits, which consequently reappears with the volcanic rocks deposited during the early eugeosynclinal stages of Phanerozoic orogenic belts.

In later stages of subduction, when the continental crustal block has farther overridden the oceanic plate, melting may involve increasing amounts of sialic crustal material, either by anatexis of sialic blocks underthrust at the trench and carried downward by subduction (James, 1971, p. 3343) or by partial melting in the base of the thickened continental plate itself. These stages may have involved

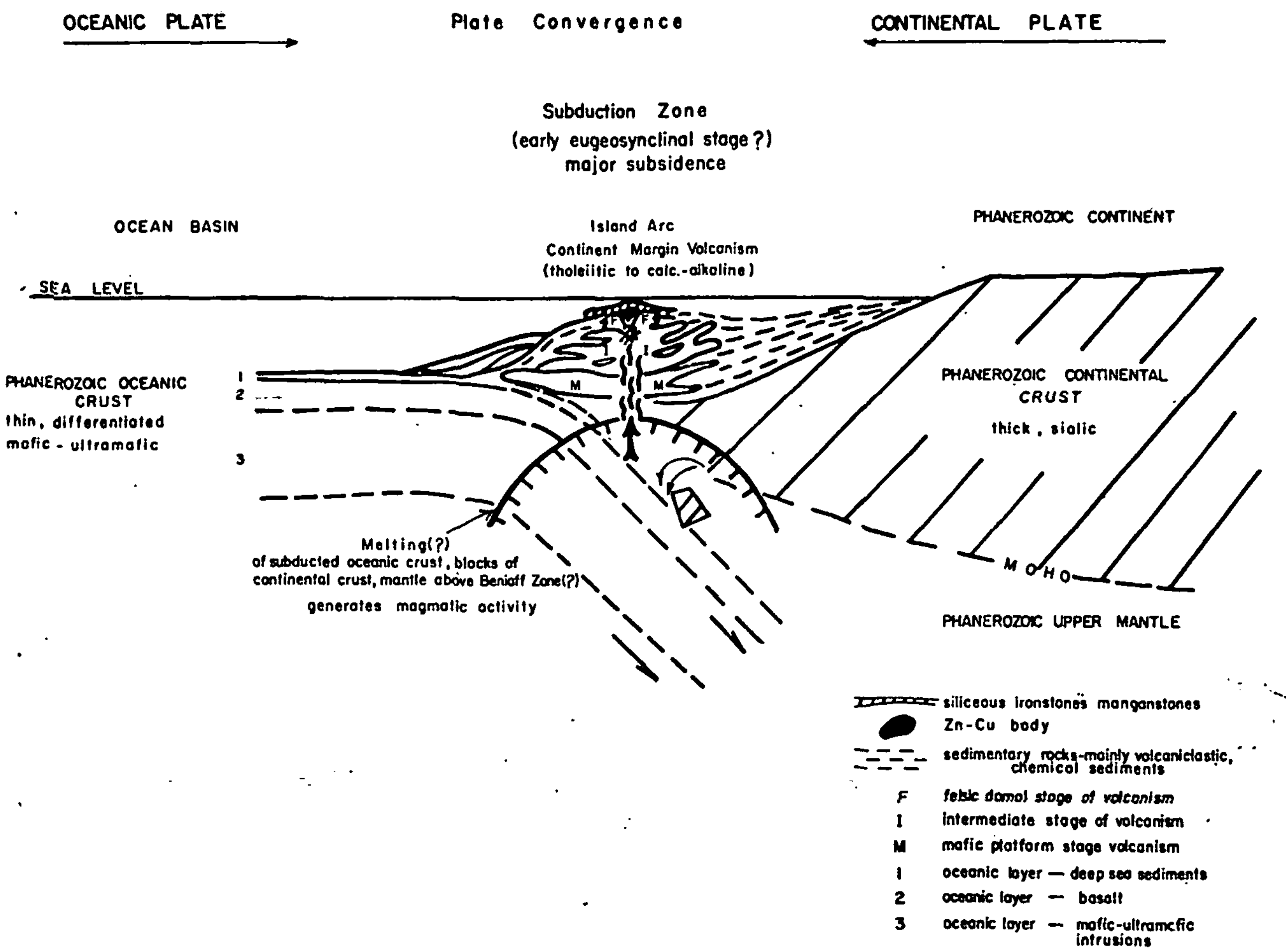

FIG. 5. Diagrammatic illustration: Zn-Cu type of volcanogenic massive sulfide deposit regenerated in early stage of Phanerozoic continent margin mobile belt.

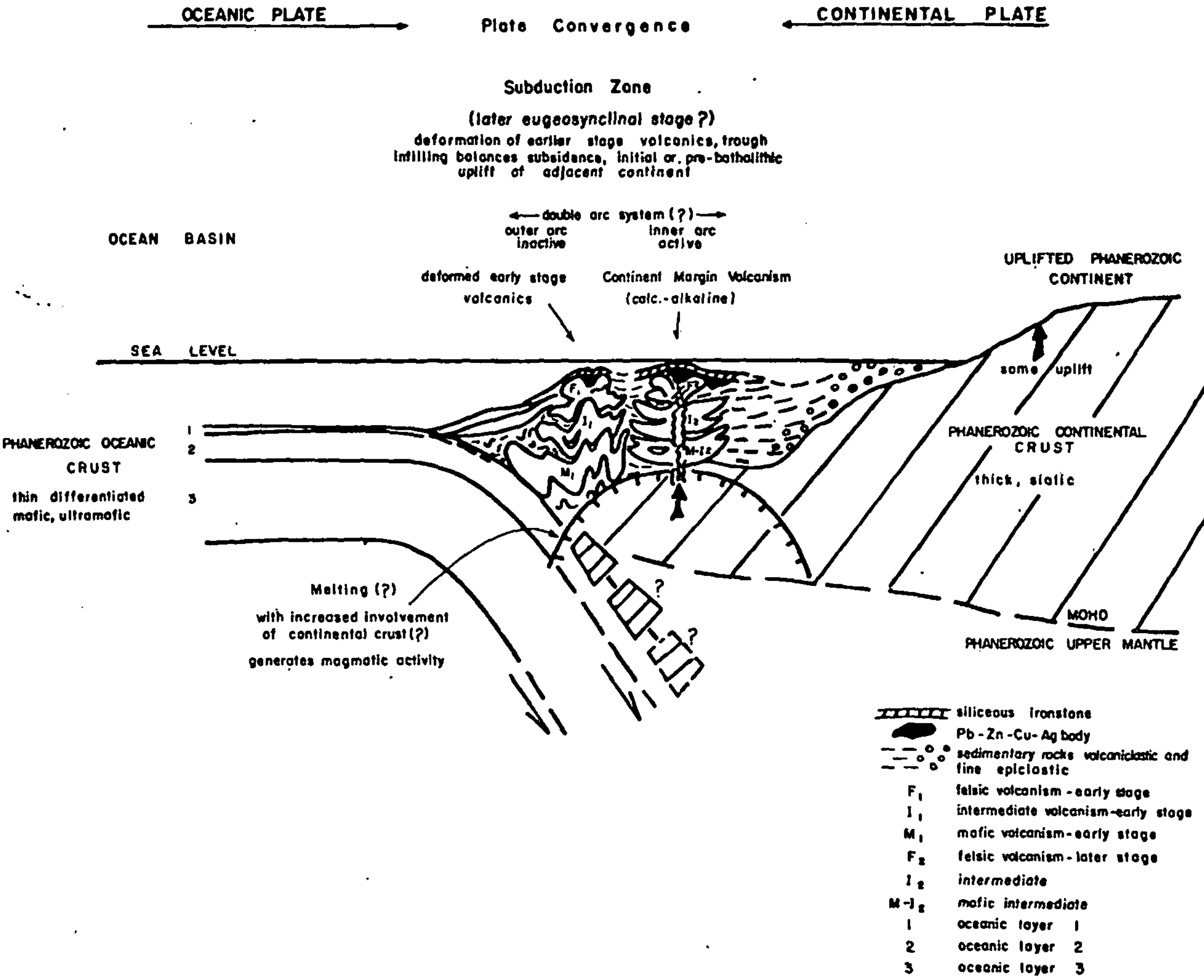

FIG. 6. Diagrammatic illustration: Pb-Zn-Cu-Ag type of volcanogenic massive sulfide deposit regenerated in later stages of Phanerozoic continent margin mobile belt.

island arc-continent collisions (Dewey and Bird, 1970, p. 2640) with underthrusting of the oceanic plate beneath major continental blocks, perhaps like the present Andean margin of South America or the Cordilleran margin of North America during Mesozoic time. These conditions are diagrammatically illustrated in Figure 6. Whatever their magmatic lineage, the resulting extrusives are more felsic and calc-alkaline and are usually displaced continentward from the earlier, more mafic volcanic sequences (James, 1971, p. 3342). These earlier sequences, with their contained copper-zinc deposits may now have undergone varying deformation due to initial uplift and batholithic intrusion that accompanied this later evolutionary stage (James, 1971, p. 3342). Increased epiclastic sedimentation, derived both from the earlier deformed volcanics and from the adjacent and intruded continental margin, deposited increasing amounts of sedimentary rocks which are intercalated with the felsic volcanic strata. Under these conditions, infilling of the trough would keep pace, more or less, with subsidence, and the lead-zinc-copper-silver type of volcanogenic massive base metal deposit is apparently regenerated.

Examples in Phanerozoic Orogenic Belts

The Cordilleran belt in California

As suggested by some examples already named, the Cordilleran region of California provides one particularly interesting example of the sequential evolutionary tectonic-volcanic developments in a Phanerozoic orogenic belt during which all three types of the massive base metal sulfide deposits were generated. The broad geological relationships and sulfide deposits involved are shown in Figure 7.

The important massive zinc-copper sulfide deposits of the West Shasta district in California occur in volcanic rocks of Middle Devonian age, as determined from faunal evidence (Kinkel et al., 1956, p. 16, 37). Together, these volcanic rocks and sulfide bodies apparently represent a relatively early Cambrian to Devonian (pre-Antler orogeny) stage (Burchfiel and Davis, 1972, p. 98) of the much longer period of eugeosynclinal volcanism in the Cordilleran belt of the western United States that persisted from early Paleozoic until the end of Triassic time (Gilluly, 1965, p. 10). Volcanism in this early stage produced fully differentiated basalt-andesite-rhyolite successions like the Copley-Balaklala rocks of the West Shasta (Gilluly, 1965, p. 33). It is particularly interesting that deposits of the zinc-copper type reappear in these older volcanic rocks of an early Phanerozoic orogenic belt, just as the great zinc-copper deposits of the Canadian Precambrian occur in the lithologically comparable oldest Keewatin (Archean) lavas.

Contrasting with the West Shasta deposits are the lead-zinc-copper-silver deposits of East Shasta and of the Foothill Copper-Zinc Belt in California. These are examples of the second type and all occur in younger volcanic sequences of Triassic and mid-Jurassic age respectively (Albers and Robertson, 1961, p. 35; Taliaferro, 1934, p. 284). Although metamorphosed, the metavolcanic rocks that contain these sulfide deposits are predominantly of intermediate to felsic composition and in most places are of pyroclastic rather than extrusive nature, as in the Bully Hill rhyolite of East Shasta and the Logtown Ridge volcanics of the Foothill Belt (Albers and Robertson, 1961, p. 33; Taliaferro, 1943, p. 283). These ores and their volcanic hosts are products of later stage eugeosynclinal-tectonic evolution in the Cordilleran belt (Burchfiel and Davis, 1972, p. 103). This stage succeeded deformation of the earlier Paleozoic volcanic successions during the Antler orogeny (Gilluly, 1965, p. 7, 34; Burchfiel and Davis, 1972, p. 101–102) and was accompanied by the initiation of thrusting, uplift, and large-scale plutonism in the Sierra Nevada (Gilluly, 1965, p. 13; Hamilton, 1969, p. 2420). Abundant coarse, epiclastic sedimentary rocks therefore accompanied the volcanics. Sulfide deposits of the lead-zinc-copper-silver type reappear under these later tectonic-volcanic conditions, which resemble those of later Precambrian, post-Kenoran (Aphebian) time, when the older analogues of these deposits were formed, as in the Sudbury Basin and perhaps at Mt. Isa.

Finally, the small but high-grade deposit of pyrrhotite-chalcopyrite at Island Mountain in Trinity County and perhaps several similar but smaller occurrences in the Klamath Mountains in Delnorte County, all in northwestern California, occur with glaucophane schists, cherts, graywackes, shales, and serpentinites of the Franciscan formation (Stinson, 1957, p. 24; Eric, 1948, p. 225–227). The exact age of the Franciscan rocks has long been controversial but is now considered to be Late Jurassic-Cretaceous (Hamilton, 1969; Guilluly, 1965, p. 34; Bailey, 1960). The basaltic pillow lavas of the Jurassic Dothan formation in the Klamath region of southwestern Oregon-northwestern California are also of similar tholeiitic affinity (Wells et al., 1949). The Dothan and Franciscan rocks are clearly ophiolitic and of oceanic crustal derivation (Hamilton, 1969, p. 2414; Bailey et al., 1970) and were probably incorporated onto the continent during major underthrusting of the continental plate that began in Late Jurassic time and culminated in Nevadan orogeny in Late Cretaceous time (Hamilton, 1969, p. 2416; Gilluly, 1956, p. 40–41). As previously discussed, cupreous pyrite deposits are characteristic in ophiolite suites and are of Phanerozoic age, particularly Mesozoic and younger; thus the presence of deposits of this type in the Franciscan rocks of California is a typical example. It would be particularly interesting to know the exact age and tectonic relationships between these rocks and their cupreous pyrite deposits on the one hand and the lead-zinc-copper-silver deposits in the more felsic Triassic-Jurassic volcanics of the East Shasta and Foothills districts. Present information suggests that the Franciscan rocks are slightly younger than the calc-alkaline volcanic rocks

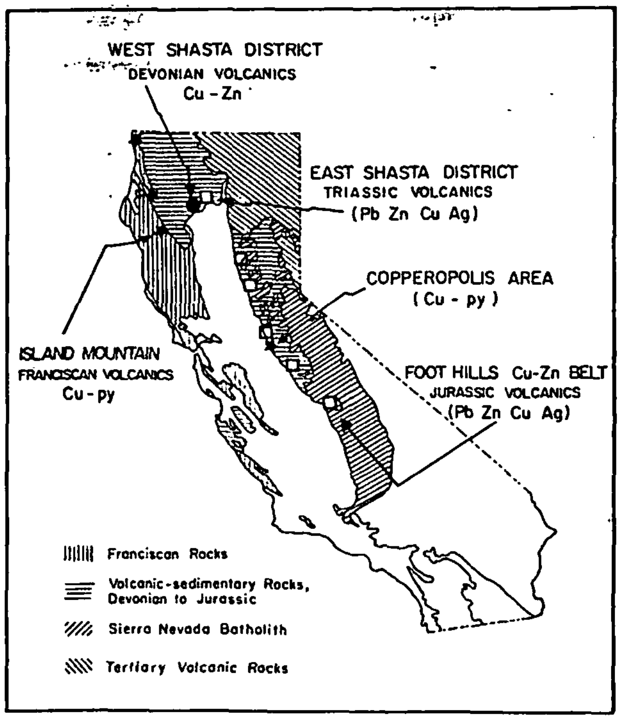

FIG. 7. Distribution of some massive volcanogenic sulfide deposits in Northern California.

of the Foothills belt. Thus they were apparently formed as oceanic crust and subsequently incorporated into the continent (Bailey et al., 1970, p. C79) at a relatively late stage in the prolonged period of subduction during which the Pacific plate underthrust the North American continent from early Paleozoic through Mesozoic time.

Tectonic evolution from early Paleozoic to late Mesozoic time in the Cordilleran belt generated the various volcanic successions and their different types of sulfide deposits in the same relative age sequence as that recognizable on the broader time scale of crustal evolution from Archean through Phanerozoic. The copper-zinc type was first to appear and is oldest in both cases, the lead-zinc-copper-silver type was next, and the cupreous pyrite deposits were last and youngest.

The Appalachian belt in Newfoundland

Early Paleozoic volcanic rocks of the Appalachian orogenic belt in west-central Newfoundland provide a second example of the generation of all three types of massive base metal sulfide deposits during a relatively short period of tectonic evolution from very late Precambrian to Late Ordovician time. The various ore bodies of the Rambler mine (Heenan and Truman, 1972) belong to the copper-zinc type judging from their lack of lead, the presence of both zinc and copper in recoverable amounts, and their relatively low silver-to-gold ratios (Table 1). Moreover, the deposits lie in rocks of the Pacquet Harbour Group which, together with rocks of the underlying Mings Bight and overlying Cape St. John Groups, constitute a sequence of mafic to felsic, apparently differentiated and calc-alkaline extrusive and pyroclastic rocks (Kennedy et al., 1972, p. 519). Intercalated with these volcanic rocks are abundant pelitic metasedimentary schists, and the succession as a whole resembles those of Archean (and younger) age in which the copper-zinc type is found. All these rocks belong to the Fleur de Lys Supergroup of Eocambrian to Cambrian age (Stevens, 1970) and rest on an earlier deformed granitic gneiss basement (Kennedy et al., 1972, p. 517). Thus, these volcanic rocks and zinc-copper sulfide deposits represent a very early stage of eugeosynclinal tectonism in the Appalachian belt, probably generated during subduction along an arc-trench type continent margin. The Fleur de Lys rocks were subsequently deformed prior to deposition of Lower Ordovician volcanic rocks of considerably different character (Kennedy et al., 1972, p. 521).

Ophiolitic volcanic rocks of the Bay of Islands-Baie Verte-Betts Cove-Lushes Bight sections are of Early to Early-Middle Ordovician age and were emplaced following deformation of the Fleur de Lys rocks (Church, 1972, p. 73). Pillow lavas belonging to these ophiolitic suites contain numerous cupreous pyrite deposits of the third type including the Tilt Cove, Whalesback, Little Deer, Little Bay, and Betts Cove mines (Strong, 1972b; Williams et al., 1972; Kanehira and Bachinski, 1968; Kennedy and DeGrace, 1972), and possibly the Gullbridge mine (Upadhyay and Smitheringale, 1972, p. 1071). These rocks are of oceanic crustal derivation (Smitheringale, 1972) formed during the opening by rifting of a proto-Atlantic Ocean in Early Ordovician time. Like the Franciscan of California, they were subsequently thrust onto the continent by later closing of the ocean, accompanied by underthrusting of the oceanic crust by the continent margin along a southeast-dipping subduction zone (Church, 1972, p. 81; Upadhyay et al., 1971). Their incorporation into the continent thus marks the beginning of a second, later Middle to Late Ordovician, period of subduction.

Finally, the intermediate to felsic volcanic and pyroclastic rocks of the Buchans Group belong to the central volcanic belt of Newfoundland and contain more epiclastic sedimentary rocks than the earlier successions (Swanson and Brown, 1962, table I, p. 2). They have been variously assigned to the Silurian (Williams, 1967, fig. 6) and to the Middle Ordovician (Swanson and Brown, 1962), but whatever their age, they are younger than the ophiolitic rocks and are clearly of calc-alkaline affinity, the products of island-arc volcanism (Williams et al., 1972, p. 6). They contain the important lead-zinc-copper-silver deposits of the Buchans mine and may be correlative with the similar Roberts Arm volcanics which contain the comparable polymetallic sulfide bodies at Pilley's Island (Strong, 1972b). These rocks and sulfide deposits apparently represent a later stage of Appalachian tectonic evolution, when renewed subduction again generated calc-alkaline volcanism. This, however, was perhaps accompanied by beginning of uplift and initiation of Siluro-Devonian plutonic activity that marked the Acadian orogeny (Williams et al., 1972, p. 9–10).

Again, in the Appalachian orogenic belt, as in the Cordilleran of California and in Archean rocks, the earliest and most primitive stages of tectonic-magmatic evolution were favorable to formation of the zinc-copper type of sulfide deposits. Similarly, the lead-zinc-copper-silver type accompanied a later stage of Appalachian tectonic evolution and subduction. An interesting difference in the Appalachian belt, however, is the appearance of the cupreous pyrite type in ophiolitic volcanic rocks at an intervening evolutionary stage between the former two types. The exact significance of this ophiolitic magmatism is not known (Church, 1972, p. 81), but it somehow reflects the generation of oceanic crust in Early Ordovician time. The different sequence of magmatism and sulfide generation must reflect some important difference between the tectonic history of the Cordilleran and Appalachian belts. Perhaps this difference arises from the close proximity of Europe and Africa to North America in early Paleozoic time (Church, 1969). Tectonic evolution in an orogenic belt that developed between two nearby continental plates ought somehow to be different from those in a belt, like the Cordilleran, that developed mainly along an oceanic-continental plate boundary.

Tasman geosynclinal belt, Australia

In the early Paleozoic Tasman geosyncline of Eastern Australia the cupreous pyrite type is not recognized, although two serpentine belts (Voisey, 1965, fig. 3) may represent the ophiolite environment. However both the copper-zinc and lead-zinc-copper-silver types are known. Again, these two types appear in the same relative age sequence as in the former North American examples. The numerous small deposits of the Bathurst area, N.S.W. (Stanton, 1955) appear to be mainly of the copper-zinc type, although specific metal content of the ores is not available. They occur in a thick Late Ordovician to Early Silurian eugeosynclinal volcanic-sedimentary succession. The host effusive-pyroclastic rocks are basaltic to andesitic, and the sedimentary rocks are predominantly graywackes and tuffaceous shales. The deposits lie in volcanic rocks at two definite stratigraphic horizons separated by several thousand feet of intervening sedimentary strata. It is especially interesting that copper-zinc-rich bodies completely lacking in lead occur along the lower of the two horizons, whereas lead is an important constituent in some of the upper horizon bodies (Stanton, 1955, p. 691). The sedimentary strata above the lower horizon are limey, and this may have been an important factor in concentrating lead in the deposits of the upper horizon. In any case these relationships clearly indicate increasing lead generation and/or deposition in later stages of tectonic evolution within a single volcanic-sedimentary complex.

The same trend toward higher lead content in later stages of tectonic evolution is illustrated on a broader regional scale in the Tasman belt by comparing the Bathurst N.S.W. deposits with those at Captains Flats, N.S.W. The important lead-zinc-copper-silver deposits of Captains Flat occur in somewhat younger andesitic to rhyolitic rocks of Upper Silurian age which are considerably more felsic than those at Bathurst (Glasson and Paine, 1965, p. 424). The associated sedimentary strata contain more fine-to-coarse clastic epicrustal detritus than at Bathurst. The comparison illustrates the same trends as in the Appalachians and Cordillera of North America; increasingly felsic and perhaps shallower subaqueous volcanism, increasing amounts of continent-derived detritus, and increasing lead-silver content in the ores through time. These trends must result from tectonic evolution in orogenic belts. Thus they are presumably due to increasing involvement of continental crust during prolonged subduction along colliding continental-oceanic plate margins.

The very important Mt. Morgan deposits in Queensland cannot, sensu stricto, be considered as volcanogenic massive sulfide deposits, although massive mineralization is present (Cornelius, 1969, p. 896). Rather, they appear to be an extensive copper-gold-rich, silicified, brecciated, and sulfide-impregnated zone very like the mineralized and brecciated alteration pipes so common beneath many massive volcanogenic bodies. Nevertheless they lie in predominantly felsic volcanic and pyroclastic rocks of mid-Devonian age that are of very shallow water or subaerial deposition (Mt. Morgan Ltd. Staff, 1965, p. 365) and may represent a very late evolutionary stage of Siluro-Devonian eugosynclinal volcanism in the Tasman belt. Under these conditions, fumarolic volcanism may have been subvolcanic, failing to break through to surface, or if it did so, it failed to deposit a massive body in the oxidizing, shallow water, or subaerial environment. Deposits like Mt. Morgan, sharing some characteristics of both porphyry copper and massive sulfide deposits, are believed to have formed in just such environments (Hutchinson and Hodder, 1972). In this context it is interesting that Cornelius (1969, p. 901) compared the Mt. Morgan deposit to the massive pyritic United Verde bodies at Jerome, Arizona, which have recently been reinterpreted as volcanogenic deposits (Anderson and Nash, 1972).

The deposits of Mt. Lyell and Rosebery in Tasmania at the southern extremity of the Tasman belt (Solomon and Elms, 1965; Hall et al., 1965) occur in Cambrian volcanic rocks and apparently belong to an older cycle of eugeosynclinal volcanism than the New South Wales-Queensland deposits. Although the Mt. Lyell ore bodies have produced mainly copper, both these and the Rosebery deposits appear to belong to the lead-zinc-copper-silver type. This is indicated by their relatively high silver content and by the presence of some lead-rich ores in both areas (Table 1). Moreover the host volcanic rocks in both districts are mainly of intermediate to felsic composition (Solomon and Elms, 1965, p. 480) with abundant porphyritic and pyroclastic members, and the intercalated sedimentary rocks contain abundant epiclastic detritus (Hall et al., 1965, p. 486).

Aspects for Additional Study

The foregoing sections are a preliminary attempt to explain the relative age and evolutionary relationships among the observed types of volcanogenic massive sulfide ores and their host volcanic-sedimentary rocks. They may also provide a model of evolutionary crustal tectonics and related volcanism which can be critically examined in the light of future investigations of the ores and their geological environments.

Particularly necessary for critical evaluation of these or subsequent hypotheses are petrochemical data for the volcanic and subvolcanic rocks associated with these ores. Admittedly, representative sampling and reliable petrochemical data are extremely difficult to obtain for such thick, widely differentiated, and invariably altered rock successions, nor is it easy to interpret from analytical data the original petrochemistry and petrography of these rocks. Nevertheless, such information is fundamental to a thorough understanding of these relationships, and initial studies, however imperfect, are essential. Particularly lacking are petrochemical and petrographic data for the volcanic rocks associated with the zinc-copper and lead-zinc-copper-silver types, in both their Precambrian and Phanerozoic settings. Adequate data for the ophiolitic rocks associated with the cupreous pyrite type are already available and clearly show the oceanic tholeiitic character of these basalts.

Present information for the Keewatin volcanics that contain the zinc-copper type are somewhat contradictory. Wilson et al. (1965) considered these rocks to be calc-alkaline, whereas subsequent work (Baragar, 1968) suggests tholeiitic affinities for basalts of the mafic platform, but calc-alkaline parentage for the intermediate to felsic rocks of the Archean domical complexes (Baragar and Goodwin, 1969). Considering the primitive crust-mantle conditions herein suggested, it is possible that all these affiliations are valid, but additional more accurate analytical data and more extensive and representative sampling are necessary to substantiate this explanation. Similar comparative petrochemical data for volcanic sequences of Proterozoic age containing the lead-zinc-copper-silver deposits are also lacking. Tholeiitic basalts are present in Aphebian orogenic belts (Dimroth et al., 1970, p. 116) but are not often found closely associated with the lead-zinc-copper-silver ores. Very little petrochemical data are available for volcanic successions of Aphebian age due, no doubt, to the relative rarity of such successions. Additional petrographic-petrochemical studies of the Eastern Creek volcanics beneath the Urquhart shale at Mt. Isa (Bennett, 1965), of the Aphebian volcanic successions at Sudbury (Card and Hutchinson, 1972, fig. 6; Stevenson, 1971, pp. 44–46), and of other Aphebian volcanic successions (Hoffman et al., 1970, p. 206) are essential for elucidating these relationships.

Finally, similar comparative petrochemical-petrographic data for volcanic sequences in Phanerozoic orogenic belts are also deficient. It will be important to compare such data for volcanic rocks containing sulfide deposits of the same type but of differing age, as, for example, comparison of Archean volcanic rocks with those of the pre-Ordovician Fleur de Lys Supergroup in Newfoundland and with those of Middle Devonian age in West Shasta (Kinkel et al., 1956, p. 23). This comparison should reveal major, time-dependent evolutionary trends in volcanic petrochemistry. Similar comparisons for volcanic rocks containing the other two types of sulfide deposits, but again of differing age, would be equally informative. A second interesting comparison is between volcanic rocks of approximately similar or successive ages that contain the three different types of sulfide deposit, as, for example, in the early Paleozoic Appalachian belt or in the mid-Paleozoic-early Mesozoic Cordilleran belt. Comparisons of this kind should elucidate magmatic differentiation trends that develop during tectonic evolution in Phanerozoic orogenic belts. Information of this kind is generally available for evolutionary magmatism across plate boundaries (Kuno, 1966), and the trend from oceanic tholeiites to calc-alkaline series lavas and finally to alkaline rocks has been recognized in progression from the oceanic to the continent side of arc-trench environments. Similarly, Gass (1970) has discussed the trend of magmatic evolution across a developing continental rift. These data and trends have not yet, however, been directly related to the distribution and variation in sulfide ore deposits that occur in volcanic rocks of these different affiliations and tectonic environments.

Also needed are additional and more reliable quantitative data concerning the metal content of the ores themselves. Ratios among the three major base metals in the three different types require further investigation and documentation, as do silver-gold ratios in the same ores. This information is not readily available, occasionally because it is considered confidential, but more often because when any of these elements is too scarce in the ore for commercial recovery, it is commonly overlooked in development, mining, and assaying. Metal production data are inaccurate because they reflect mining, milling, and smelting recoveries as well as metal content in the ore. There is the added problem of obtaining representative samples from these prominently zoned ore bodies. Yet these data are essential for investigation of the changing ratios of base and precious metals in these ores, both through time and in space across plate boundaries. Metal ratios will, in all likelihood, show evolutionary changes parallel to those in volcanic petrochemistry because, if the ideas set out here are correct, the original fumarolic volcanism which generated the ores must be yet another expression of the changes in magmatism during tectonic evolution.

Finally, and perhaps most critical of all, is the need for information about age relationships among the various types of these deposits and their associated rocks. Both absolute radiometric dating of the rocks and ores and stratigraphic information concerning their ages relative to one another are critical in correctly unravelling and explaining these evolutionary relationships.

Due to the prominent metal zoning that is so characteristic of the entire family of volcanogenic massive base metal sulfide deposits, chalcopyrite-pyrite-pyrrhotite mineralization is common in the deeper portions of all three different types. Consequently, metal content or metal ratios *alone* cannot be reliably used to assign the deposits to one or another of the types. It is essential that the other variable geological characteristics (Fig. 1) which aid in distinguishing the three types also be considered, along with metal content, in making such assignments.

Possible Significance and Applicability

Although not yet refined, the preceding concepts may have useful applications in the search for these types of mineral deposits. Suitable geological environments may be more accurately defined and favorable time-stratigraphic epochs recognized.

The concepts also have potential scientific significance. They may improve our understanding of earth's tectonic evolution, particularly in very early and obscure Precambrian time. They may also help in unravelling the complex tectonism of younger orogenic belts such as the Cordilleran of western North America. A specific example of the latter is the Foothill Copper-Zinc Belt in California (Fig. 7; Heyl, 1948a) where known sulfide deposits are almost all of the lead-zinc-copper-silver type in highly schistose felsic volcanic or pyroclastic rocks of the Logtown Ridge volcanics of the mid-Jurassic Amador Group. Notable exceptions however are the important North Keystone, Keystone Union, and Empire deposits in the Copperopolis area (Heyl, 1948b) which have previously been considered part of the belt. These ore bodies contain only pyrite and chalcopyrite and lack both lead and zinc, with only very low values in gold and silver (Reid, 1907, p. 398). Moreover, they occur with chloritic metavolcanic and serpentinized ultramafic rocks in which asbestos is presently mined nearby (Leney and Loeb, 1971). These relationships suggest that neither the ore bodies nor their associated mafic-ultramafic rocks are in the same category as the other sulfide deposits of the Foothill Copper-Zinc Belt, a conclusion reached long ago in the earliest published accounts (Reid, 1907, p. 397–398). Alternatives are that the Copperopolis area is (1) a small slice of pre-Franciscan ophiolite infolded with the older mid-Jurassic calc-alkaline volcanic rocks, or (2) a small slice of Franciscan ophiolite that has been thrust into its present site during Cretaceous deformation of the Great Valley-Coast Range region (Hamilton, 1969, p. 2414–2416; Bailey et al., 1970, p. C79), or (3) an allochthonous slice of Franciscan rocks moved more than 100 miles eastward across the Great Valley from the continent margin, possibly by gravity sliding off the uplifted Coast Range. The second alternative seems most probable and the third least so, but further detailed structural, stratigraphic, and petrochemical investigations appear warranted.

Finally, if massive base metal sulfide deposits are covolcanic with their hosts and, like fauna, are the products of evolutionary change, then they may provide a means of relative age determination and correlation on a regional scale. This could be particularly useful in very ancient rocks or in any rocks where faunal or absolute dating is impossible or unreliable due to metamorphism. For example, in the Grenville and Churchill structural provinces of the Canadian Shield one suspects the coexistence of both older Archean and younger Aphebian rocks, subsequently metamorphosed and infolded together. Here the types of base metal sulfide occurrences alone may serve as useful age indicators and correlation tools. When considered with data on volcanic petrochemistry, sedimentary petrology, stratigraphy, and tectonics from the associated rocks, the relative ages of different belts may be assigned with more confidence.

An example may be the Jerome-Prescott area of Arizona. The metal content of the United Verde and UVX deposits near Jerome and the characteristics of the associated volcaniclastic and volcanic rocks (Anderson and Nash, 1972) closely resemble those of the zinc-copper type in Archean rocks. In contrast with these, the Iron King deposits east of Prescott (Gilmour and Still, 1968) have metal contents and associated rocks that are like those of the lead-zinc-copper-silver type in Proterozoic rocks. Moreover, these two volcanic successions are separated by the Shylock Fault, a regional structural discontinuity (Anderson and Creasey, 1958, p. 77). Yet radiometric age dates on plutonic and pyroclastic rocks near Jerome (Anderson and Nash, 1972, p. 847) suggest a Proterozoic (late-Aphebian) age for the host rocks of the United Verde deposits. Considering the foregoing geological characteristics, these age dates may reflect a later metamorphic event, and this area may contain both older Archean rocks and volcanogenic sulfide deposits (at Jerome) and younger Aphebian rocks and deposits (east of Prescott) which were deformed and infolded together about 1,760 m.y. ago. Similar relationships may exist in the Flin Flon region of Manitoba (Sangster, 1972; Slawson and Russell, 1973), and further investigations of this problem are warranted in both and in other areas.

Speculations Concerning Metal Source

If three types of massive volcanogenic base metal sulfide deposits have evolved with crustal tectonics as suggested herein, some broad speculations about the derivation of metals and their behavior in the geochemical cycle are relevant.

The abundance of copper and iron in cupreous pyrite deposits suggests that these metals in the volcanogenic family of deposits are mainly of "oceanic crust-mantle" derivation. Apparently they remain sufficiently abundant in the mantle, despite 4 b.y. of crustal evolution, to provide an adequate source of copper and iron for the cupreous pyrite ores in the ophiolitic environment.

Conversely, the virtual absence of lead in the cupreous pyrite type, compared to its abundance with zinc, but lower copper in the lead-zinc-copper-silver type, suggests the reverse situation for lead. Apparently lead is mainly of supracrustal derivation in these ores and has perhaps been recycled into them through reworking by leaching, anatectic melting, or palingenesis of sialic crust. After 4 b.y. of crustal evolution lead has apparently been impoverished in the mantle and differentiated into supracrustal rocks.

Low lead values in the Archean zinc-copper ores contrasted with high lead content in the younger Proterozoic-Phanerozoic lead-zinc-copper-silver type (Table 1) immediately suggest the possibility that in Archean time lead had not yet been generated through radioactive decay in sufficient abundance to be important in these early ores. Lead isotopic data, however, does not wholly support this speculation (Kanasewich, 1968, p. 215). Additions of new radiogenic lead from early Proterozoic to late Mesozoic time appear to have been minor and when extrapolated back to Archean time are probably insufficient to explain the major shift in metal content of the ores. Nevertheless, radiogenic lead additions must have contributed to the greater abundance of this metal in the younger deposits. It will be necessary to compare isotopic lead ratios from Archean deposits with those of Proterozoic and younger age to evaluate the extent of these additions.

If the increase in lead in younger rocks cannot be explained by radiogenic lead additions, it follows that lead was simply not concentrated and deposited under Archean tectonic conditions. The "fixing" of lead in later tectonic environments must somehow be related to evolutionary changes in lithosphere, hydrosphere, and atmosphere that accompanied tectonic evolution. The change from a reducing, carbon dioxide-rich atmosphere-hydrosphere in Archean time to oxidizing, carbon dioxide-poor conditions in Proterozoic time (Cloud, 1972, p. 543–546) apparently caused increased deposition of carbonate rocks (Cloud, 1968, p. 455) and must also have affected the solubility and stability relationships of the various lead carbonate, sulfate, and sulfide species in the oceans. The abundance of stratiform sulfates with the younger lead-zinc-copper-silver deposits also suggests an oxygenated depositional environment and contrasts with the rarity of sulfates in the Archean zinc-copper type formed under earlier, more reducing conditions. Finally, higher heat flow and surface temperatures presumably accompanied the more extensive Archean volcanism, causing lead to enter preferentially the silicate and oxide structures, rather than the very insoluble sulfide form (Rankama and Sahama, 1950, p. 733).

The abundance of zinc with copper in the Archean deposits, with lead in the lead-zinc-copper-silver type, and occasionally also with copper in the cupreous pyrite type, suggests that zinc has both mantle and supracrustal affinities. In the primitive tectonic environment of Archean time it accompanied copper and iron and was presumably of mantle derivation. During evolutionary crust-mantle differentiation zinc was mainly partitioned into supracrustal rocks, there joining lead. In Phanerozoic orogenic belts zinc was subsequently recycled with lead into the younger lead-zinc-copper-silver deposits, where the proportions of these metals, and of copper, may depend on the extent of epicrustal vs. mantle involvement at successive stages of subduction in arc-trench environments. The evolutionary partitioning of zinc between crust and mantle has apparently not been so complete as that of lead however, and enough zinc remained in the mantle to accompany occasionally iron and copper in cupreous pyrite deposits formed at spreading rift-ridge systems.

There is an alternative possibilty for the derivation of zinc in a few, small, zinc-rich deposits of cupreous pyrite type. In Cyprus, where the stratigraphic positions of numerous sulfide bodies within the 3,000 foot-thick pillow lava sequence is reasonably well-known, the zinc-rich bodies lie at deeper horizons in Basal Group or Lower Pillow Lavas (Bear, 1963, tables 10, 11; p. 61, 71, 73), whereas larger, copper-rich bodies lie at shallower levels in or at the top of Upper Pillow Lavas. Insofar as lower stratigraphic levels in ophiolitic pillow basalt successions may represent early stages of sea-floor spreading, it may be inferred from the relationships in Cyprus (Gass, 1968) that zinc-rich subaqueous fumarolic activity accompanied these earlier stages, whereas copper-rich activity accompanied later stages. Thus the formation of zinc-rich cupreous pyrite bodies may occur at an early stage of rifting when there has been only slight separation between rifted continental plates. In this setting, zinc may have been derived, not from mantle, but from the adjacent sialic plate, zinc content decreasing correspondingly in later (higher stratigraphic) bodies formed at subsequent stages of greater separation.

Some supporting evidence for this possibility is available. The metal-rich sediments in the Red Sea deeps are zinc-rich (Bischoff, 1969), forming in the initial stages of rifting in a very young ocean where there has been minimum separation between the adjacent Arabian and Nubian sialic plates (Hutchinson and Engels, 1970). These relationships led Blissenbach (1972) to suggest that deposits of this type form only during early stages of separation, when thick evaporite sequences on the nearby continent margin provide a source for metal-carrying saline brines. The stratigraphic relationships in Cyprus lend support to this suggestion. Conversely, other examples of metal-bearing sea-floor sediments are rich in iron and manganese and have copper greatly in excess of lead, although zinc was not determined (Bostrom and Peterson, 1966). These are considerably more remote from continental blocks, lying on the crest of the East Pacific Rise in a mid-oceanic rift-ridge environment. Moreover they may not be comparable to those in the Red Sea because they are probably deep-sea pelagic sediments rather than direct precipitates from heated, heavy metal-rich brines.

Iron is abundant in all three types, but greatly predominates in the cupreous pyrite deposits and is least abundant, proportionate to the other metals, in the lead-zinc-copper-silver deposits. This reflects the overall abundance of iron and its stability in both mantle and supracrustal environments. It remains abundant in the mantle after 4 b.y. of crustal evolution and from here is directly contributed into cupreous pyrite deposits. It has also, however, been extensively concentrated into supracrustal rocks through time, and can be recycled from these into deposits of the other two types.

Speculations Concerning Age

Despite the need for additional information about the absolute and relative ages of their volcanic host rocks, the indicated time-age relationships of the three types of volcanogenic massive base metal sulfide deposits are interesting. These relationships suggest that each type occurs throughout a certain time range and, moreover, that there may have been optimum times for the formation of each. These time ranges and optima reflect various stages of the earth's crustal tectonic evolution.

Deposits of the zinc-copper type are commonest, largest, and richest in Archean rocks, are of less importance in earlier Proterozoic (Aphebian) rocks, but are apparently absent in younger Proterozoic (Helikian-Hadrynian) rocks. They reappear in the earliest Paleozoic stages of Phanerozoic orogenic belt evolution, but are again apparently lacking in rocks younger than mid-Paleozoic age, perhaps representing an "extinction" of this type. Deposits of the lead-zinc-copper-silver type appear first in early Proterozoic time, may reach a maximum in early Paleozoic orogenic belts (New Brunswick), and continue into belts of Teritary age (Japan). Finally, deposits of the cupreous pyrite type appear first in earliest Paleozoic or Eocambrian orogenic belts, apparently reach a peak in the Jura-Cretaceous, and also continue in ophiolitic environments of Tertiary, and perhaps Quaternary (Red Sea) age.

These relationships are represented diagrammatically in Figure 8. They are reminiscent of faunal ranges, first appearances, greater proliferations, and perhaps extinctions. Thus they, too, suggest an evolutionary development of the three types of massive base metal sulfide deposits in which

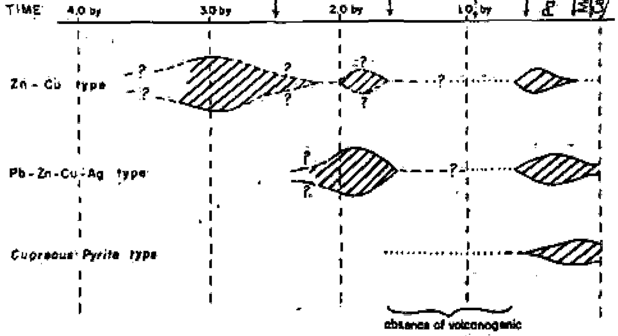

FIG. 8. Apparent time ranges and maxima of massive volcanogenic sulfide deposits.

the zinc-copper type appeared earlier, the other two following later, in response to later stages of crustal evolution. If the optimum times for generation of the three types can be confirmed by additional studies, they would be useful in area selection for mineral exploration and in evaluation of mineral potential. It would be particularly interesting to know whether there was an optimum time for generation of the important zinc-copper type *within* the long span of poorly understood and subdivided Archean time. This information would be helpful in appraising mineral possibilities in Archean terrane in Africa, Australia, and Brazil where this type of deposit is not yet known.

The similar sequence in relative ages of the three types, both through time and within Phanerozoic orogens, suggests that tectonic evolution in Phanerozoic orogenic belts along plate margins above subduction zones recapitulates, on a shortened time scale, the same broad changes involved in earth's overall crustal evolution. This situation, too, is reminiscent of paleontology where "ontogeny duplicates phylogeny." Perhaps it affords scope, within the current dynamic concepts of modern plate theory, to accommodate the older static views of classic uniformitarianism.

A particularly puzzling aspect of the time-age relationships is the apparent absence of volcanogenic massive base metal sulfide deposits of any type in younger Proterozoic (Helikian-Hadrynian) rocks, from 1.6 to about 0.7 b.y. (Fig. 8). This might be expected from the rarity of subaqueous volcanic sequences of this age. It must arise from, hence offer some indication of the tectonic processes and the stage of crustal evolution during this time (Hutchinson, 1973). It is probably significant that base metal deposits of two other important families seem commonest in the two sedimentary lithofacies that characterize Helikian-Hadrynian successions. These are the massive, but probably sedimentary, exhalative (Oftedahl, 1958) family of lead-zinc ores such as the Sullivan and Faro-Dynasty deposits of western Canada and the Broken Hill deposits of New South Wales and secondly, the sedimentary copper deposits of Northern Michigan, the African Copper Belt, and possibly Australia (Johns, 1965).

Summary and Conclusions

The volcanogenic family of massive base metal sulfide ore deposits can be divided into three different types. These are identifiable by their base and precious metal assemblages, the petrochemistry and petrogenesis of their associated volcanic rocks, the nature and relative amounts of their associated volcaniclastic, epiclastic, and chemical sedimentary rocks and their ages, both absolute and relative. All these features are coproducts of differing tectonic environments in which the rocks and ores alike were deposited.

A zinc and copper-rich type containing both silver and gold appears oldest and characteristic of Archean rocks. It reappears in earliest Paleozoic stages of young orogenic belt evolution but is apparently rare or absent in younger Phanerozoic rocks. It is considered a primitive, poorly differentiated ancestor of the other two types. A lead-zinc-copper-silver-rich type appears in early Proterozoic time, reoccurs in somewhat younger stages of Phanerozoic orogenic belt evolution than the zinc-copper type, and continues into young Tertiary rocks. A cupreous pyrite type is first recognizable in early Paleozoic ophiolitic rocks, is most important in similar Mesozoic rocks, and also occurs in Tertiary rocks. These two types are considered to be later, more specialized evolutionary derivatives of the earlier, primitive zinc-copper type. They developed in response to the more diverse tectonic processes that evolved in later geologic time. The changes in the tectonic environment that govern the type of base metal deposit formed, result from broad crustal tectonic evolution, explaining the progression in absolute ages of the different types and also their relative age progression in Phanerozoic orogenic belts.

The analogy between organic evolution and mineral deposit evolution from primitive and less diversified forms in earlier time to more varied and specialized forms in later time is striking. As in paleontology, the evolutionary changes are "unidirectional"; primitive families and types of mineral deposits reappear in younger time when primitive tectonic environments are somehow reproduced, but later, more specialized mineral deposits are lacking in very ancient rocks. When the evolutionary succession and relationships of the broad spectrum of mineral deposits is more completely understood through additional broad metallogenic investigations, then mineral deposits, like fossils, may prove usable as gross-scale age indicators and correlation tools in very old, or very highly metamorphosed terrain. Insofar as they reflect the tectonic environments in which they were originally formed, mineral deposits may also provide a useful added tool in unravelling the complex tectonic history of highly deformed orogenic belts. Metallogenic studies also have obvious practical uses in helping to define the most favorable regions, rock types, and rock ages for mineral exploration for a wide variety of important ore deposits.

Finally, it is evident that metallogenic studies of this type are interdisciplinary, bridging several fields of geoscience specialization. Igneous and sedimentary petrology are fundamental to studies of the volcanic-sedimentary rocks in which the ore bodies are found; stratigraphy and structural geology are basic to unravelling their age relationships, correlations, and tectonic history. Mineral deposits geology is necessary to provide basic data about the nature and geological setting of the ore deposits. Geochemistry is essential to proper interpretation of the magmatic and metallic evolution that is involved, and tectonophysics is probably the broad base to which all the former must be related. Economic geologists who seek to use metallogenic studies in mineral exploration or for other applied or basic purposes must become familiar with these fields. Conversely, those in specialized fields of geology who ignore mineral deposits and metallogenesis may forego another tool that could be usefully applied in their own specializations.

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Department of Geology
University of Western Ontario
London, Ontario, Canada
January 18, April 10, 1973

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A WORKING TERMINOLOGY OF PYROCLASTIC DEPOSITS

JOHN V. WRIGHT[1], ALAN L. SMITH[1] and STEPHEN SELF[2]

[1] *Department of Geology, University of Puerto Rico, Mayaguez, PR 00708 (U.S.A.)*
[2] *Department of Geology, Arizona State University, Tempe, AZ 85281 (U.S.A.)*

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ABSTRACT

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A nomenclature for pyroclastic deposits is proposed which it is hoped will provide a working terminology. Three basic types of deposit may be distinguished: fall, flow and surge, and descriptions are given of their dominant characteristics. No unique classification for pyroclastic rocks can be made and at least two systems are required: (1) a genetic classification to interpret the genesis of a deposit, and (2) a lithological classification which may be solely descriptive, but also which may be used to discriminate on a lithological basis the mechanisms which produced a particular pyroclastic deposit. Genetic classification schemes are presented for various types of fall, flow and surge deposits. A lithological classification is given based on grain size limits and distribution, constituent fragments and degree and type of welding. A glossary of some other terms in use to describe pyroclastic deposits is also given.

INTRODUCTION

The purpose of this paper is to propose a nomenclature for pyroclastic deposits and rocks; this system, however, is not meant to be a review of all the terms now in use (for a comprehensive review the reader is referred to Fisher, 1966). Work on young pyroclastic deposits by us has shown that the existing terminology is of limited use for our studies. Accordingly, we propose a genetic classification relating deposits to eruptive mechanisms, together with a descriptive lithological classification.

Pyroclastic deposits are those formed by the fragmentation of magma and rock by explosive volcanic activity. Three kinds of component are found in any pyroclastic deposit, namely juvenile vesiculated fragments, crystals and lithics:

Juvenile vesiculated fragments. This term includes highly vesiculated pumice and denser, less well vesiculated juvenile magmatic fragments. Indeed the

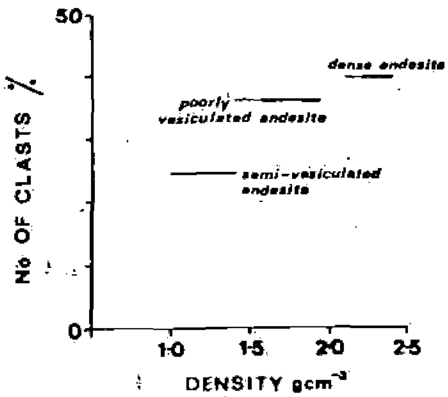

Fig. 1. Variation in density of juvenile fragments in a single pyroclastic deposit. The example is a block and ash deposit from Mt. Pelée, Martinique.

same pyroclastic deposit may contain a complete range in density from pumice to dense non-vesiculated juvenile fragments (termed *cognate lithics*, see below). This is illustrated in Fig. 1 with an example from Mt. Pelée, Martinique.

Pumice is the common product of explosive eruptions involving viscous magmas. In the literature the term has become inextricably linked to the larger grain sizes. Consequently we believe it necessary to emphasize grain size in order to avoid confusion. The terms *pumice block* or *bomb* (> 64 mm), *pumice lapilli* (< 64 mm to > 2 mm) and *ash* (< 2 mm) should be used when appropriate (see Table 5 in section "Lithological classification"). In addition the terms *glass shard* and *ash shard* for ash-sized particles which result from the fragmentation of pumice vesicle walls is in wide use and cannot be abandoned. Morphologically glass shards have various Y and cuspate shapes. The reader is referred to Heiken (1974) for scanning electron photomicrographs of various types of glass shards and ash particles.

Scoria is the term used to describe vesiculated fragments of basalt and basaltic andesite compositions. The larger fragments tend to have a ropey or breadcrust surface texture.

Crystals. During the explosive disruption of porphyritic magmas the crystals are released and can be regarded as forming a discrete juvenile component, often having their own behavior patterns during transport.

Lithics. This term is generally used to describe the dense components in a pyroclastic deposit. Lithics may be subdivided into non-vesiculated juvenile fragments (*cognate lithics*) and country rock which has been explosively ejected during eruption (*accessory lithics*) or, in the case of pyroclastic flows and surges, clasts picked up locally during transport (*accidental lithics*).

Following Sparks and Walker (1973) three basic types of pyroclastic deposits are distinguished:

(1) *Pyroclastic fall deposits.* These are produced when material is explosively ejected from the vent into the atmosphere producing an eruption column in the form of a convective plume. The plume expands and pyroclasts fall back under the influence of gravity, down wind, at varying distances from the source. The geometry and size of the deposits reflects the eruption column height, and wind velocity and direction (Eaton, 1964; Knox and Short, 1964; Shaw et al., 1974; Wilson, 1976; Blackburn et al., 1976; Wilson et al., 1978). Fall deposits show mantle bedding, maintaining a uniform thickness over restricted areas while draping all but the steepest topography (Fig. 2a). They are generally well sorted and sometimes show internal stratification due to variations in eruptive column conditions. Carbonized wood is generally lacking in these deposits, when it does occur it is invariably restricted to near vent deposits.

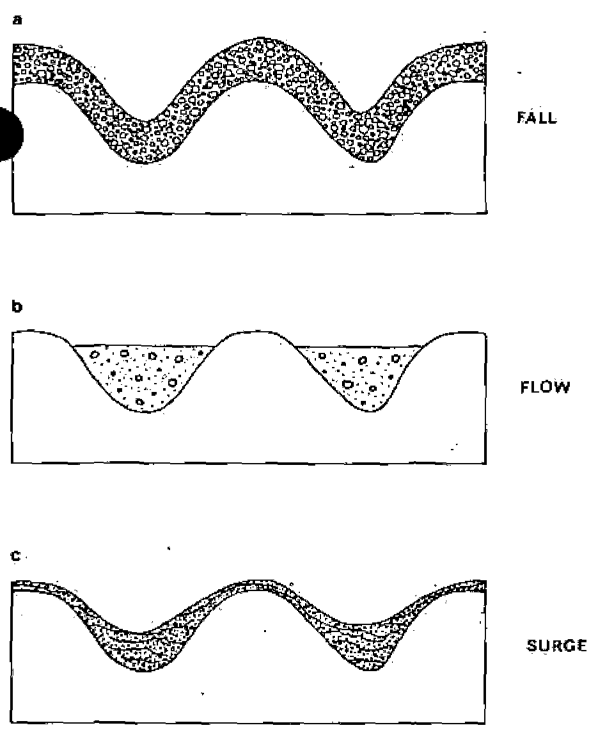

Fig. 2. Geometric relations of the three main types of pyroclastic deposit overlying the same topography.

(2) *Pyroclastic flow deposits.* Pyroclastic flows involve the lateral movement of pyroclasts as a gravity controlled, hot, high concentration gas/solid dispersion, which may in some instances be partly fluidized (Sparks, 1976).

Deposits are topographically controlled, filling valleys and depressions (Fig. 2b). They are poorly sorted and sometimes show coarse-tail grading (Smith, 1960a; Sparks, 1976). Poor sorting in flow deposits is attributed to high particle concentrations and not turbulence, with the dominant flow mechanisms probably being laminar and/or plug flow (Sparks, 1976; Wright, 1979; Wright and Walker, 1980; Wilson, 1980). Individual deposits generally lack internal stratification, although the superposition of a number of flow units (each flow unit being regarded as the deposit of a single pyroclastic flow) can give the appearance of stratification. They sometimes contain fossil fumarole pipes in which the fine ash fraction has been lost making the pipes enriched in crystals, lithics or vesicular fragments (Walker, 1971, 1972; Roobol and Smith, 1976). Carbonized wood may also be present.

(3) *Pyroclastic surge deposits.* Surges involve the lateral movement of pyroclasts as expanded, turbulent, low-concentration gas/solid dispersions. Deposits do mantle topography but tend to accumulate thickest in depressions (Fig. 2c). Characteristically they show unidirectional sedimentary bedforms (cross-stratification, dunes, planar lamination, antidunes, pinch and swell structures, and chute and pool structures) and individual laminae are generally well sorted (Fisher and Waters, 1970; Waters and Fisher, 1971; Crowe and Fisher, 1973; Schmincke et al., 1973; Sparks, 1976; Wohletz and Sheridan, 1979). They also can contain carbonized wood.

In detail it seems probable that there is a complete gradation in concentration from high-concentration, high-density pyroclastic flows to low-concentration, low-density surges. Indeed ash-cloud surge deposits represent the lateral equivalents of pyroclastic flows (Fisher, 1979; Fisher et al., 1980). Of course surges are a type of flow, but the term pyroclastic flow has traditionally been associated with the high-concentration flows and it is appropriate to classify the fundamentally different types of deposits produced by flows and surges separately.

PROPOSED CLASSIFICATION

There is no unique classification system for pyroclastic rocks and at best two quite different systems are needed. One to interpret the genesis of deposits which can then be related to a volcano's history, behavior pattern and eruptive mechanisms. The other system is lithological. Such a system is prima ly descriptive, describing major characteristics of a deposit such as grain size. However, these features are themselves often diagnostic of a particular proces and allow conclusions to be made on the deposit's genesis.

Genetic classification

The genesis of a pyroclastic deposit is partly deduced from its lithology but also from its overall geometry and field relations. In ancient volcanic terrains where rocks may have suffered much erosion and even tectonic deform;

tion the record may not be good enough to determine parameters such as geometry and grain size, and hence, a genetic classification cannot be rigorously applied. A genetic classification system can indeed only be rigorously applied to very young, Quaternary deposits.

For pyroclastic fall deposits the classification scheme of Walker (1973) is adopted (Fig. 3). This quantitative scheme relies on accurate mapping of the distribution of a fall deposit and detailed granulometric analysis (for methods see Walker, 1971; and Walker and Croasdale, 1971). The term phreatoplinian has been introduced by Self and Sparks (1978) for the salic analog of a surtseyan deposit (which could more strictly be termed phreatostrombolian). Ultraplinian has been recently introduced by Walker (1980) to describe the most widely dispersed plinian fall deposits.

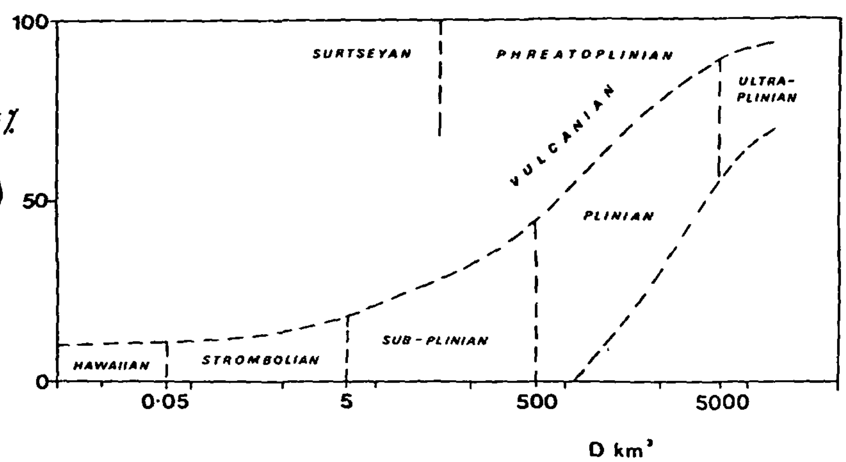

Fig. 3. Classification scheme of pyroclastic fall deposits after Walker (1973). Plot of F (weight percentage of deposit finer than 1 mm on the axis of dispersal where it is crossed by the 0.1 T_{max} isopach) against D (the area enclosed by the 0.01 T_{max} isopach). Ultraplinian has been introduced by Walker (1980) for the most widely dispersed plinian fall deposits. The field of vulcanian deposits is described in Fig. 4 and discussed in text.

Vulcanian pyroclastic fall eruptions (short-lived explosions) appear to be common on andesite and basaltic-andesite strato-volcanoes (Self et al., 1979) and they generally produce small-volume (< 1 km^3), fine-grained ash deposits with a wide dispersal (Fig. 4). However, during recently observed eruptions, for example Fuego 1974 (W.I. Rose, personal communication) and Ngaurohoe 1975, coarser-grained scoria fall deposits of more limited dispersal were produced during periods of more intense, maintained explosions. These occurred intermittently with short-lived explosions and hence two types of deposit were formed during eruptions which have overall been termed vulcanian. These coarser scoria fall deposits seem to have similar fragmentation and dispersal indices to those deposits termed violent strombolian by Walker (1973). The approximate field of vulcanian deposits is shown in Fig. 4 by a few examples.

TABLE 1

Genetic classification of pyroclastic flows

Essential fragment	Eruptive mechanism		Pyroclastic flow	Deposit
Vesiculated				
			pumice flow	ignimbrite; pumice and ash deposit
	Eruption column collapse		scoria flow	scoria and ash deposit
Decreasing average density of juvenile clasts				
			semi-vesicular andesite flow	semi-vesicular andesite and ash deposit
		explosive	block and ash flow; nuée ardente	block and ash deposit
	Lava/dome collapse			
Non-vesiculated		gravitational	block and ash flow; nuée ardente	block and ash deposit

Historic examples	Comments
Valley of Ten Thousand Smokes ignimbrite, Katamai, Alaska, 1912 (Fenner, 1920; Curtis, 1968)	Large-volume deposits formed by continuous collapse of a plinian eruption column as envisaged by Sparks et al. (1978). Salic in composition
Komagatake, Japan, 1929 Aramaki and Yamasaki, 1963)	Small-volume deposits probably formed by interrupted column collapse as in the case of scoria flows described below. Intermediate to salic in composition
Soufrière, St. Vincent, 1902 (Anderson and Flett, 1903; Roobol and Smith, 1975); Hibok-Hibok, Philippines, 1950-51 (Macdonald and Alcarez, 1956); Mount Lamington, Papua, 1951 (Macdonald, 1972); Mayon, Philippines, 1968 (Moore and Melson, 1969); Ngauruhoe, North Zealand, 1975 (Nairn and Self, 1978)	Small-volume deposits probably formed by interrupted eruption column collapse produced by short explosions (see Nairn and Self, 1978). Basalt to andesite composition. Some deposits do contain large unvesiculated blocks, e.g. those of Ngauruhoe 1975
Asama, 1783 (Aramaki, 1956, 1957)	Small-volume deposits composed of semi-vesiculated angular andesite clasts
Mt. Pelée, Martinique, 1902 (LaCroix, 1904; Roobol and Smith, 1975; Fisher et al., 1980); Bezymianny, Kamchatka, 1956 (Gorshkov, 1959); Santiaguito, Guatemala, 1973 (Rose et al., 1977)	Small-volume deposits, usually andesitic or dacitic in composition. Produced both by explosive collapse of an actively growing dome or lava flow and by the collapse of a vertical eruptive column as recognized in the early eruptions (e.g. May 8 and 20) of Mt. Pelée 1902 (Fisher et al., 1980)
Merapi, Indonesia, 1930 (Van Bemmelen, 1949); Santiaguito, Guatemala, 1967 (Stoiber and Rose, 1969)	Small-volume deposits usually andesitic or dacitic in composition. These are the hot-avalanche deposits of Francis et al. (1974)

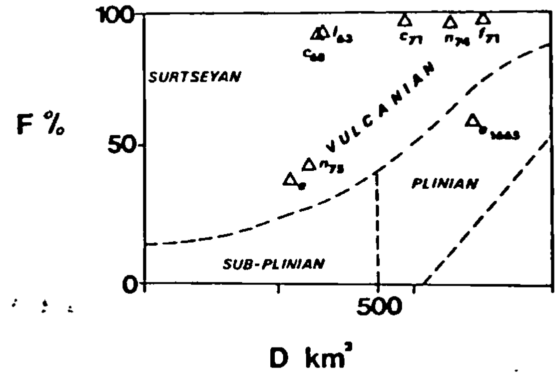

Fig. 4. Fragmentation F plotted against dispersal D (terms defined in Fig. 3) to show the field of vulcanian deposits which is discussed in text; $c_{68,71}$: eruptions of Cerro Negro, Guatemala, in 1968 and 1971; e: an old undated fall deposit of Mt. Egmont, New Zealand; e_{1665}: fall deposit of the 1665 eruption of Mt. Egmont; f_{71}: eruption of Fuego, Guatemala, in 1971; i_{63}: eruption of Irazú, Costa Rica in 1963; $n_{74,75}$: eruption of Ngauruhoe, New Zealand, in 1974 and 1975.

For pyroclastic flows the proposed classification is presented in Table 1 and our system is compared with other classifications in Table 2. For pyroclastic surges the proposed classification is presented in Table 3. A brief description of both pyroclastic flow and surge deposits is given in Table 4.

With regard to the classification of pyroclastic flows and their deposits several points are worth further discussion:

(1) The scheme is based on a limited number of observations of actual eruptions, which have invariably produced small volume pyroclastic flows composed of relatively dense material. The mechanism of continuous column collapse as envisaged by Sparks and Wilson (1976) and Sparks et al. (1978) for the formation of ignimbrites has as of yet not been observed. Field evidence in support of continuous collapse has however been described from an ignimbrite in Mexico, with the recognition that an extremely coarse, lithic-rich airfall deposit termed a *co-ignimbrite lag-fall deposit* (Wright and Walker, 1977) shows the same compositional zoning as the ignimbrite. This deposit consists of pyroclasts that are too large and heavy to be carried away in the pumice flows and is therefore thought to indicate the site of the eruptive column collapse.

(2) Another type of deposit formed by the eruption of a pyroclastic flow is the *vitric air-fall ash deposit* (*co-ignimbrite ash-fall* of Sparks and Walker, 1977). This consists of the fine vitric ash lost during the eruption and transport of a pyroclastic flow; this forms the upper part of the turbulent cloud seen in the observed historic pyroclastic flows (for photographs see Moore and Melson, 1969; Nairn and Self, 1978). Studies have shown that ignimbrites and other types of pyroclastic flow deposits are crystal enriched due to loss of this vitric component (Hay, 1959; Lipman, 1967; Walker, 1972; Sparks

TABLE 2

Comparison of various classifications of pyroclastic flows

This paper: based on eruptive mechanism and characteristics of the deposits			Murai (1961): based on type eruptions and characteristics of the deposits	Williams and McBirney (1979): based on site of and type eruptions	Smith (1979): based on volume of the deposits
Pyroclastic flow	*Deposit*		Valles-type	large-volume ash-flow tuffs (100—1000 km^3)	
Pumice flow	ignimbrite; pumice and ash	(large volume)	VTTS-type* Krakatoa-type	VTTS-type Krakatoa-type	intermediate-volume ash-flow tuffs (1—100 km^3)
		(small volume)	St. Vincent-type	St. Vincent-type	**
Scoria flow	scoria and ash	(small volume)	St. Vincent-type	St. Vincent-type	
Semi-vesicular andesite flow	semi-vesicular andesite and ash		intermediate-type	Asama-type	
Block and ash flow;[1] nuée ardente	block and ash		Sakurajima-type	Peléan-type	small-volume ash-flow tuffs (0.001—1.0 km^3)
Block and ash;[2] nuée ardente	block and ash		Pelée-type	Peléan-type	small-volume ash-flow tuffs (0.001—1.0 km^3)
Block and ash;[3] nuée ardente	block and ash		Merapi-type	Merapi-type	small-volume ash-flow tuffs (0.001—1.0 km^3)

[1] Produced by collapse of vertical eruptive column accompanying lava/dome collapse.
[2] Produced by explosive collapse of an actively growing lava flow or dome.
[3] Produced by gravitational collapse of an actively growing lava flow or dome.
*VTTS = Valley of Ten Thousand Smokes.
**Overlap between boundary of intermediate- and small-volume ash-flow tuffs required to compare Smith's classification with other classifications is shown by dashed line.

TABLE 3

Genetic classification of pyroclastic surges

Essential fragment	Eruptive mechanism	Type of surge	Historic examples	Comments
Vesiculated—(non-vesiculated)	collapse of a phreatomagmatic eruption column	base surge	Taal, Philippines, 1965, 1966 (Moore et al., 1966; Waters and Fisher, 1971); Capelinhos, Faial, Azores, 1957—1958 (Machado et al., 1962; Waters and Fisher, 1971)	Base surges result from the explosive interaction of magmatic material and water and are consequently cool. They are often associated with maar volcanoes and tuff rings. The historic examples described are both basaltic but older phonolitic base surges are known from the Laacher See area, Germany (Schmincke et al., 1973) and rhyolitic ones from the Minoan eruption of Santorini (Bond and Sparks, 1976)
Vesiculated—(non-vesiculated)	accompanying pyroclastic flows erupted by mechanisms given in Table 1	ground surge	associated with examples of pyroclastic flows given in Table 1	Ground surge, although originally introduced by Sparks and Walker (1973) to encompass all pyroclastic surges, is here used to describe those surges found at the base of pyroclastic flow deposits, as well as those produced without any accompanying pyroclastic flow
Vesiculated (non-vesiculated)	also associated with air-fall deposits by collapse of an eruption column but without generation of pyroclastic flow		Askja, Iceland 1875 (Sparks and Wright, 1979); Vesuvius A.D. 79 (R.S.J. Sparks, personal communication)	
Vesiculated—non-vesiculated	—	ash-cloud surge	Mt. Pelée, Martinique, 1902* (Fisher et al., 1980)	Ash-cloud surges (Fisher, 1979) are the turbulent, low-density flows derived from the overriding gas-ash cloud of pyroclastic flows. These may in some cases become detached from the parent pyroclastic flow and move independently

*Ground surges were also produced by the 1902-03 eruptions of Mt. Pelée, e.g. August 30, 1902 (Fisher et al., 1980).

and Walker, 1977). Vitric air-fall ash deposits may be very extensive and have volumes comparable with those of the parent pyroclastic flow deposits (Sparks and Walker, 1977).

(3) The most obvious surficial manifestation of a pyroclastic flow is the overriding ash cloud. It was this feature that influenced LaCroix (1903) to call the pyroclastic flows generated by Mt. Pelée in 1902 nuées ardentes. It is, however, obvious from his work (LaCroix, 1904, p. 350) that he meant the term to refer to the complete phenomenon, that is, to both the overriding ash cloud and the basal avalanche or underflow. In a later publication LaCroix (1930) expanded the use of the term to include all types of pyroclastic flows. This use of the term nuée ardente for pyroclastic flows produced by different mechanisms has persisted to the present day with the result that the exact meaning of the term is now rather ambiguous. This has been especially so during the last few years with the recognition that the ash cloud produces distinct deposits (surge deposits) from the underflow, pyroclastic flow sensu stricto. In view of these problems the authors feel that either the term should be avoided altogether, the pyroclastic flows produced by "Peléan-type" eruptions being called block and ash flows (Perret, 1937); or nuée ardente should be restricted the original definition and only be used for those pyroclastic flows produced by the collapse of an actively growing lava flow or dome (Rose et al., 1977; Smith and Roobol, 1980). Both definitions are used in Tables 1 and 2.

(4) The deposits from pumice flows are termed ignimbrites or pumice and ash deposits. Such deposits are found in oceanic islands (Iceland, Azores), island arcs (Lesser Antilles), continental margins (Andes) and continental interiors (Western United States). They may be subdivided on the basis of lateral extent and volume into large-volume and small-volume ignimbrites (or pumice and ash deposits). The former are restricted in their occurrence to continental margins and interiors, and large islands (e.g. New Zealand); generally they result from eruptions of salic calc-alkaline magmas and tend to form large ignimbrite sheets (Smith, 1979). The latter are characteristically developed on island arc volcanoes and oceanic islands but can also be found in a continental setting as well (Francis et al., 1974; Sparks, 1975; Wright, 1979). Thus modern island arcs are characterized by small-volume ignimbrites while those produced at the active margins and interiors of continents are often of large volume and cover wide areas. This distinction should be borne in mind in interpreting the volcano-tectonic setting of deposits from pumice flows in ancient environments.

(5) Concerning the classification scheme of Williams and McBirney (1979, and Table 2), the Valley of Ten Thousand Smokes type and the Valles type are included as a category of pyroclastic flows discharged from fissures. Many workers have suggested that ignimbrites are erupted from fissures or ring fractures (Smith, 1960a,b; Van Bemmelen, 1961; Rittman, 1962; Smith and Bailey, 1968; Macdonald, 1972); this type of eruption has been regarded as the way of accounting for the more voluminous deposits.

The fissure hypothesis originated from the observation of lines of fumaroles

TABLE 4

Summary descriptions of types of pyroclastic flow and surge deposits

Deposit	Description
Ignimbrite; pumice and ash	Unsorted ash deposits containing variable amounts of rounded salic pumice lapilli and blocks up to 1 m in diameter. In flow units pumice fragments can be reversely graded while the lithic clasts can show normal grading; ungraded flow units are as common. A fine-grained basal layer is found at the bottom of flow units. They sometimes contain fossil fumarole pipes and carbonized wood. The coarser smaller-volume deposits usually form valley infills while the larger-volume deposits may form large ignimbrite sheets. Sometimes they may show one or more zones of welding
Scoria and ash	Topographically controlled, unsorted ash deposits containing basalt to andesite vesicular lapilli and scoriaceous ropey surfaced clasts up t 1 m in diameter. They may in some circumstances contain large non-vesicular cognate lithic clasts. Fine-grained basal layers are found at the bottom of flow units. Fossil fumarole pipes and carbonized woo may also be present. The presence of levées, channels and steep flow fronts indicate a high yield strength during transport of the mo ing pyroclastic flow
Semi-vesicular andesite and ash	Topographically controlled unsorted ash deposits containing intermediate vesicular (between pumice and non-vesicular juvenile clasts) andesite lapilli, blocks and bombs. Fine-grained basal layers, fossil fumarole pipes and carbonized wood all may be present
Block and ash	Topographically controlled, unsorted ash deposits containing large, generally non-vesicular, jointed, cognate lithic blocks which can exceed 5 m in diameter. The deposits are generally reversely graded. Fine-grained basal layers are again present. Again they may contain fossil fumarole pipes and carbonized wood. Surface manifestations include the presence of levées, steep flow fronts and the presence of large surface blocks, all of which again indicate a high yield strength during transport of the flow
Base surge	Stratified and laminated deposits containing juvenile vesiculated fragments ranging from pumice to non-vesiculated cognate lithic clasts, ash and crystals with occassional accessory lithics (larger ballistic ones may show bomb sags near-vent) and deposits produced in some phreatic eruptions which are composed totally of accessory lithics. Juvenile fragments are usually less than 10 cm in diameter du to the high fragmentation caused by the water/magma interaction. Deposits show unidirectional bedforms. Generally they are associated with maar volcanoes and tuff rings. When basaltic in composition they are usually altered to palagonite
Ground surge	Generally less than 1 m thick, composed of ash, juvenile vesiculated fragments, crystals and lithics in varying proportions depending on constituents in the eruption column. Typically enriched in denser components (less well vesiculated juvenile fragments, crystals and lithics) compared to accompanying pyroclastic flow. Again they show unidirectional bedforms; carbonized wood and small fumarole pipes may be present

TABLE 4 (continued)

Deposit	Description
Ash-cloud surge	Stratified deposits found at the top of and as lateral equivalents of flow units of pyroclastic flows. They show unidirectional bedforms, pinch and swell structures and may occur as descrete separated lenses. Grain size and proportions of components depend on the parent pyroclastic flow. Can contain small fumarole pipes

in the Valley of Ten Thousand Smokes ignimbrite, 1912 eruption of Katmai. This led Fenner (1920) to conclude that the ignimbrite erupted from fissures in the floor of the valley. More recently, Curtis (1968), from detailed mapping of the closure of isopach maps of pyroclastic fall deposits produced in the same eruption, located the central vent of Novarupta as the source. Now Wright et al. (1979) have suggested that, at least initially, the Bandelier Tuffs from Valles caldera (Valles "type example" of Williams and McBirney, 1979) were erupted from a central vent.

Korringa (1973) and Ekren and Byers (1976) have documented large welded ignimbrites erupted from linear vent systems in Nevada. However, the field evidence for such fissure eruptions is meagre in many cases. The authors believe that ignimbrites are generally erupted from central vents.

(6) The group of rocks known variously as foam lavas, froth flows, tuff lavas and tuffolavas (Vlodavetz, 1963; Shirinian, 1962, 1963; McCall, 1964; Cook, 1966; Macdonald, 1972) have been interpreted as intermediate between lavas and pyroclastic flows (Panto, 1963) or as part of a gradational sequence from lava to ignimbrite (McCall, 1964). According to M.R. Sheridan (personal communication) the classic Russian examples from Armenia are ignimbrites, while the so-called froth flows of Vulsini volcano (Locardi and Mittempergher, 1965, 1967) can be interpreted as other types of pyroclastic rocks (Sparks, 1975). A study of rocks corresponding to published accounts of tuffolavas by one of the authors on Pantelleria (Wright, 1979, 1980) agrees with the view of Sparks. In the opinion of the authors tuffolavas, etc., do not exist as a separate category, but represent misinterpretations of other kinds of volcanic rock.

Lithological classification

The main bases of lithological classification are:

(1) The grain size limits of the pyroclasts and the overall size distribution of the deposit.

(2) The constituent fragments of the deposit.

(3) The degree and type of welding.

Both (1) and (2) can be used to help discriminate the genesis of a particular pyroclastic deposit in the older geological record.

Grain size. With regard to the grain size limits of pyroclastic fragments, the system of Fisher (1961) is adopted (Table 5).

TABLE 5

Grain size limits for pyroclastic fragments after Fisher (1961)

Grain size (mm)	Pyroclastic fragments	
	coarse	blocks and bombs
256	fine	
64	lapilli	
2	coarse	ash
1/16	fine	

Concerning the overall grain size distribution, granulometric analysis of non-welded and unlithified pyroclastic deposits can be an important discriminant in deducing their mechanism of formation (Murai, 1961; Walker, 1971; Sheridan, 1971; Sparks, 1976). Pyroclastic flow deposits generally show extremely poor sorting, while pyroclastic fall deposits are better sorted. A convenient way of representing grain size data is by Md_ϕ/σ_ϕ plots (Fig. 5)*. Using such a plot Walker (1971) was able to define separate field for the de-

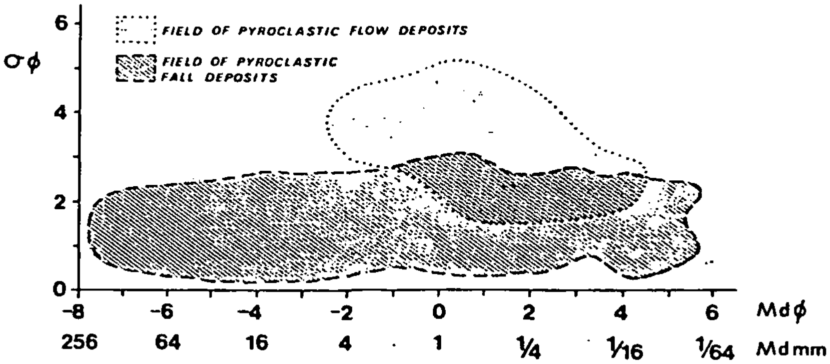

Fig. 5. Md_ϕ/σ_ϕ plot used to show the fields of pyroclastic fall and flow fields after Walker (1971). Dotted line is the 1% contour of the pyroclastic flow field from Walker (1973). Dashed line is the 1% contour for the field of pyroclastic fall deposits modified by Sparks and Wright (1979) after Walker (1973).

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*Cumulative curves of the grain size distribution are drawn on arithmetic probability paper and the Inman (1952) parameters of median diameter Md_ϕ ($= \phi_{50}$) and σ_ϕ ($= \phi_{84} - \phi_{16}/2$), which represents the graphical standard deviation and a measure of sorting, are derived. For discussion of the validity of these statistics when applied to pyroclastic deposits see Sparks (1976) and Bond and Sparks (1976).

posits produced by pyroclastic flows and falls. Pyroclastic flow deposits usually have σ_ϕ values greater than 2.0, while in general pyroclastic fall deposits have values less than 2.0. Based on the supposed dominance of ash sized particles in pyroclastic flows Smith (1960a,b) and Ross and Smith (1961) introduced the term ash-flow tuff. This term has now become almost synonymous with the older term ignimbrite. The authors feel that, despite some vagueness of the original description of Marshall (1935), the definition of ignimbrite by Marshall (p. 38, and reproduced in Chapin and Elston, 1979, p. 2) as applied here, for deposits solely from pumice flows, is valid and its use should be continued, especially in view of the fact that in most small volume pumice flows the modal grain size is in the lapilli or bomb size range (Smith and Roobol, 1980). It is worth pointing out here that ash-flow tuff has been used to describe block and ash-flow deposits (included in small volume ash-flow tuffs of Smith, 1979, and Table 2) although these deposits only contain between 15 and 40% ash (A.L. Smith and M.J. Roobol, unpublished data). The authors feel that the use of ash-flow tuff for such deposits is unteneable.

Md_ϕ/σ_ϕ plots of pyroclastic surge deposits tend to overlap both pyroclastic flow and fall fields (Roobol and Smith, 1974; Sparks, 1976). Individual laminae of deposits can be well sorted while channel samples through several laminae are often more poorly sorted (see Bond and Sparks, 1976). Ground surge deposits seem in general to be better sorted than base surge deposits (Handler et al., 1980); this may be due to cohesion of water-saturated fine ash to larger particles in the base surge cloud. Ash-cloud surge deposits formed from block and ash flows of the 1902 eruption of Mt. Pelée can be extremely coarse; some contain blocks up to 70 cm and are more poorly sorted than many pyroclastic flow deposits (Fisher et al., 1980).

In certain welded and lithified rocks grain size analyses can be attempted using thin sections and photographs. Such analysis can be of value in deciding the mechanism of formation of deposit, while for ancient, pyroclastic rocks even a qualitative assessment of the size distribution is of value.

Constituent fragments. A summary of the dominant components in a pyroclastic deposit provides a qualitative lithological description as well as providing some information as to the genesis (Table 6).

Welding. This post-depositional process involves the sintering together of hot vesicular fragments and glass shards under a compactional load (Smith, 1960a,b; Ross and Smith, 1961; Ragan and Sheridan, 1972; Sparks and Wright, 1979). Welding has only generally been described from ignimbrites; no descriptions in the literature are known to the authors of welding occurring in other denser-clast pyroclastic flows, however certain semi-vesicular andesite and ash deposits from Mt. Pelée, Martinique are welded (A.L. Smith and M.J. Roobol, unpublished data). Welded air-fall tuffs are known in the literature and their characteristics and the criteria for distinguishing them from welded ignimbrites have been documented by Sparks and Wright (1979) and Wright (1980).

TABLE 6

Summary of the components in pyroclastic deposits

A. Pyroclastic flows and surges

Type of flow or surge	Essential components: vesicular	Essential components: non-vesicular	Other components
Pumice flow/surge	pumice	crystals	Accessory and accidental lithics
Scoria flow/surge	scoria	crystals	cognate, accessory and accidental
Lava debris flow/surge Nuée ardente	poor-moderate vesicular clasts	cognate lithics and crystals	accidental lithics

B. Pyroclastic falls

Grain size (mm)	Type of fall	Essential components*: vesicular	Essential components*: non-vesicular	Other components
>64	agglomerate	pumice/scoria		cognate and accessory lithics
	breccia		cognate and/or accessory lithics	
>2	lapilli deposit	pumice/scoria	cognate and/or accessory lithics	crystals
<2	ash deposit	pumice/scoria	crystals and/or cognate and/or accessory lithics	

*Depending on type of deposit.

Features of these tuffs indicate post-emplacement compaction and welding over a relatively wide area so that they cannot be simply ascribed to the flattening of spatter lumps on impact (see *agglutinate* in the glossary below). Welded pyroclastic surge deposits are known to occur in the products of the 1875 eruption of Askja, Iceland (Sparks and Wright, 1979) from Mayor Island, New Zealand (S. Self, unpublished data), and in the Bandelier Tuffs, New Mexico (R.V. Fisher, S. Self and J.V. Wright, unpublished data).

Those ignimbrites in which welding is fully developed show three characteristic zones of dense, partial and incipient, and no welding (Smith, 1960b). In the zones of dense welding glass shards and larger, commonly glassy, flattened pumice fragments (*fiamme*) define a planar foliation or eutaxitic texture. Incipiently welded tuffs are here distinguished from partially welded tuffs on the basis that although coherent, they show no textural evidence of deformation of the constituent fragments. The term *sillar* (Fenner, 1948) is used here for incipiently welded tuffs. Coherence in sillars is generally due either to the sintering of clasts at their points of contact, or to vapor phase crystallization of minerals in pore spaces (see Sheridan, 1970) rather than to compactional welding.

Ignimbrites which are totally non-welded or only a small proportion of which is welded can be termed low-grade ignimbrites, in contrast to high-grade ignimbrites which show well-developed zones of dense welding and in which the proportion of welding is high.

GLOSSARY

In order to broaden the scope of this classification some terms introduced earlier and a number of other terms commonly used to describe pyroclastic deposits are described below:

Accretionary lapilli. Lapilli-sized pellets of ash, commonly exhibiting a concentric internal structure which have been recorded from pyroclastic fall, base- and ground-surge deposits; also from fossil fumarole pipes in pyroclastic flow deposits. They are believed to form by the accretion of fine ash around a nucleus of water or some solid particle. This may occur in the downwind eruption cloud, where a common mechanism of formation is by rain flushing (Moore and Peck, 1962; Walker, 1971) or in the vertical eruption column of phreatomagmatic eruptions (Self and Sparks, 1978), or by gases (in fossil fumarole pipes) streaming up through pyroclastic flow deposits (Walker, 1971).

Agglutinate. Coherent deposit of lava spatter and poorly vesiculated juvenile pyroclasts associated with strombolian and hawaiian eruptions (Macdonald, 1972).

Fiamme. Flattened glassy juvenile clasts in welded pyroclastic deposits (Zavaritsky, 1947). In most cases fiamme result from the deformation of original pumice clasts. However, Gibson and Tazieff (1967), Schmincke and Swanson (1967) and Self (1973) have suggested that some fiamme represent the flattening of non-vesiculated juvenile clasts.

Fused tuff. At the contact with lava flows and dikes older pumiceous pyroclastic deposits may become fused and resemble welded tuffs (Christiansen and Lipman, 1966; Schmincke, 1967).

Palagonite tuff. The original glass in deposits of basaltic phreatomagmatic eruptions (surtseyan and base surge) commonly alters by hydration to palagonite (Hay and Iijima, 1968). Such deposits may be lithified and could be confused with welded tuffs.

Rheomorphic welded tuff. Rheomorphism ("flow-change") of a welded tuff is the secondary mass flowage of the tuff as a coherent viscous fluid. Rheomorphic welded tuffs are characterized by stretched fiamme which define a lineation parallel to the flowage direction. Flowage is capable of producing folding and other internal flow structures similar to those found in salic lava extrusions. The term rheoignimbrite was first introduced by Rittmann (1958) to describe welded ignimbrites showing secondary flowage features. Such features in the ignimbrites of Gran Canaria were thought by Schmincke and Swanson (1967) to indicate primary laminar flow of the pyroclastic flows. However, Wolff and Wright (1981) reinterpret these as secondary flow features and document rheomorphism in both welded ignimbrites and welded air-fall tuffs.

Tephra. Thorarinsson (1954) proposed the term tephra as a collective term for all pyroclastic material transported from the vent through the air. In this original definition the term strickly applied only to pyroclastic fall deposits, however it was later re-defined (Thorarinsson, 1974) to include both pyroclastic fall and flow material.

Tuff. This is used here loosely as a collective term for all consolidated pyroclastic rocks.

Vitrophyre. This term is used to describe a densely welded tuff which in handspecimen has a glassy appearance.

Volcanic mudflows (lahars). These consist of volcanic debris mobilized by water. They are often generated during eruptions, as in the May 1902 eruptions of Mt. Pelée, Martinique and Soufrière, St. Vincent (Roobol and Smith, 1975), but can also be formed without any associated volcanic activity. They are a common feature of all recent large volcanoes and are sometimes hard to distinguish in the field from pyroclastic flow deposits. Excellent descriptions are given of lahars by Neall (1974) and from Mount Rainier by Crandell (1971).

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