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EARLY TO MIDDLE PALEOZOIC PHOSPHOGENIC PROVINCE IN TERRANES OF THE SOUTHERN CORDILLERA, WESTERN UNITED STATES

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ABSTRACT. Black shale and chert containing nodules, lenses, and/or streaks of phosphate occur in the Cordilleran miogeocline of California and Nevada, in the Roberts Mountains allochthon, Nevada, in metasedimentary rocks and roof pendants in the northern and eastern Sierra Nevada and El Paso Mountains, California, at San Marcos, northern Baja California, and in the Marathon Basin, west Texas. The phosphate is generally Ordovician, Ordovician-Silurian, Devonian, or Mississippian in age.

Analysis of the phosphates shows them to be nearly ideal fluorapatites. Four of the phosphates have rare earth abundances less than average shale and (except the Marathon Basin) relatively flat patterns with minor negative cerium anomalies. These resemble rare earth patterns of phosphorites formed in zones of coastal upwelling at latitudes less than 50°. Further, the association of the Cordilleran phosphates with black shale and chert is similar to both ancient and modern examples that formed in upwelling zones along margins of continents.

The several phosphate-bearing terranes now in the Cordillera once made up a coherent province adjacent to a single continent. Early to middle Paleozoic paleogeographic reconstructions show that the Cordilleran margin was one of the few continental margins with conditions favorable for forming phosphates in black shale and chert. The Cordilleran phosphates most likely formed adjacent to North America; subsequent displacements of phosphate-bearing terranes have been primarily along the margin.

INTRODUCTION

Paleozoic and Mesozoic terranes of the Cordillera of North America formed in a variety of tectonic settings, and many terranes subsequently were accreted to and/or translated along the continental margin (Helwig, 1974; Coney, Jones, and Monger, 1980). Several of the terranes have travelled thousands of kilometers, whereas others are little displaced. Paleomagnetic studies have shed light on the displacement histories of some terranes. Where such direct evidence is lacking, however, as in many of the Paleozoic terranes of the Cordillera, indirect techniques, such as reconstruction of the depositional environment, may constrain the displacement history.

We argue that the depositional environment of phosphatic rocks can be used to constrain the paleogeography of some early to middle Paleozoic terranes in California and Nevada. First, we describe the

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locations of the phosphate-bearing lower to middle Paleozoic rocks. Second, we compare the Cordilleran phosphates to lithologically similar phosphates known to have formed along the margins of continents in low latitudes. Third, we discuss the depositional environment of the Cordilleran phosphates and point out that the present western margin of North America was ideally located for phosphate formation during the Paleozoic. We conclude that the Cordilleran phosphates and their host rocks are not exotic to North America.

PALEOZOIC PHOSPHATIC ROCKS IN THE SOUTHERN CORDILLERA

Phosphate-bearing rock is known in the Cordilleran miogeocline (Mansfield, 1927, p. 57; Allison, 1966; Rogers and others, 1970; Sandberg and Gutschick, 1980) and the Roberts Mountains allochthon (Smith and Ketner, 1975; Coles and Snyder, 1985). Most other reports of lower to middle Paleozoic phosphate-bearing chert and shale (compiled in table 1) are only brief descriptions. We add to these several newly discovered occurrences of phosphate-bearing black chert and shale (fig. 1). The compilation covers a significant portion of the southern Cordillera of North America. Permian strata contain significant phosphate deposits in the western United States but are younger than the time interval considered in this paper.

The term phosphorite refers to rock having P_2O_5 greater than 19.5 wt percent (greater than 50 percent apatite; Pettijohn, 1975) and often appears in the literature as a general term. Rocks of lesser P_2O_5 content are here termed phosphate or phosphatic. We apply these terms to the Cordilleran deposits, because their overall P_2O_5 grade is unknown. Phosphate occurs in the Cordilleran rocks as nodules or lenses in a matrix of chert, argillite, or shale. The lenses display varying aspect ratios and in some cases grade into phosphatic laminae or streaks. The chert or argillite is host to stratiform barite at several phosphate localities.

The terms shallow or relatively shallow in this paper refer to oceanic settings shallower than abyssal depths.

Miogeocline.—Lower to middle Paleozoic strata of the southern Cordillera of the United States occur in several geologic settings (fig. 1). First, cratonic, or continental-platform, sedimentary rocks form thin (a few hundred meters thick) successions as far west as Utah, Arizona, southern Nevada, and extreme southeastern California. The second, or miogeoclinal, setting includes basal clastic rocks overlain by a thick (up to 10 km thick) relatively shallow-water carbonate and quartzite succession. This sequence occurs in western Utah, eastern Nevada, and southeastern California, to the west of the cratonic association (Stewart and Poole, 1974). Rogers and others (1970) and Sandberg and Gutschick (1980) reported occurrences of phosphate in the Nevada and Utah portions, respectively, of the Paleozoic miogeocline. Phosphate of Late Devonian age also occurs at Vaughn Gulch in the Inyo Mountains of California (C. Stevens, personal commun., 1983; table 1; figs. 1, 2B).

TABLE 1

Phosphate localities in the Southern Cordillera

Location	Host formation	Age	References* Geology	References* Phosphate
Miogeocline				
Inyo Mountains, Calif.	Squares Tunnel beds	Famennian (Late Devonian)	1–3	4
Roberts Mountains allochthon				
Toquima Range, Nev.	Vinini Fm?	Ordovician (?)	5–7	7
"	Dev-Miss beds	Famennian; Kinderhook§	7	7–9
Warm Springs, Nev.	correl. to Slaven Chert	Famennian§	10–12	this study
Roberts Creek Mountain, Nev.	Vinini Fm	Ordovician	13	this study
Edna Mountain, Nev.	Vinini Fm	Ordovician	14	15
Carlin-Pinon, Nev.	Woodruff Fm	Early Devonian	16	16
	"	Late Devonian	16	16 (possible)
Metamorphic roof pendants				
Northern Sierra Nevada, Calif.	Sierra Buttes Fm@	Late Devonian	17–20	17–21
"	upper Shoo Fly Complex	pre-Late Devonian	17,18,21,22	21
Saddlebag Lake, Calif.		Ordovician-Silurian (?)	23,24	24,25
Ritter Range, Calif.	Rush Creek sequence	early Paleozoic (?)	26	26
Bishop Creek, Calif.		Ord or Dev-Miss	27,28	4,28
El Paso Mountains, Calif.	Garlock assemblage	Camb-Ord; Dev	29–31	this study
San Marcos, Baja, Calif.		Ordovician	32–34	this study
Marathon region				
Marathon Basin, Tex.	Persimmon Gap Shale (Maravillas Chert)	Late Ordovician	35–37	36

*(1) Stevens and Ridley, 1974; (2) Stevens, 1984; (3) Stevens, 1986; (4) C. Stevens, personal commun., 1983; (5) Kay and Crawford, 1964; (6) McKee, 1976; (7) Coles and Snyder, 1985; (8) Shawe, Poole, and Brobst, 1969; (9) Rye, Shawe, and Poole, 1978; (10) Kleinhampl and Ziony, 1985; (11) Kleinhampl and Ziony, 1984, p. 216; (12) F. G. Poole *in* Stewart, 1980; (13) Murphy and others, 1978; (14) Erickson and Marsh, 1974; (15) R. Schweickert, personal commun., 1983; (16) Smith and Ketner, 1975; (17) Durrell and D'Allura, 1977; (18) D'Allura, Moores, and Robinson, 1977; (19) Brooks, Wood, and Garbutt, 1982; (20) Hanson and Schweickert, 1986; (21) Varga, 1982; (22) Hannah and Moores, 1986; (23) Brook, 1977; (24) Schweickert and Lahren, 1987; (25) Lahren, Schweickert, and Girty, 1984; (26) R. Strobel, personal commun., 1984; (27) Moore and Foster, 1980; (28) Frazier and others, 1986; (29) Carr, Poole, and Christiansen, 1984; (30) Dibblee, 1967; (31) Burchfiel and Davis, 1981; (32) Lothringer, 1984; (33) Gastil and Miller, 1984; (34) Gastil, 1985; (35) King, 1937; (36) McBride, 1970; (37) Ross, 1986

§Unpublished radiolarian age by B. Murchey, U.S. Geol. Survey, on fossil collection by K. Coles

@Includes here the Grizzly and Elwell Formations

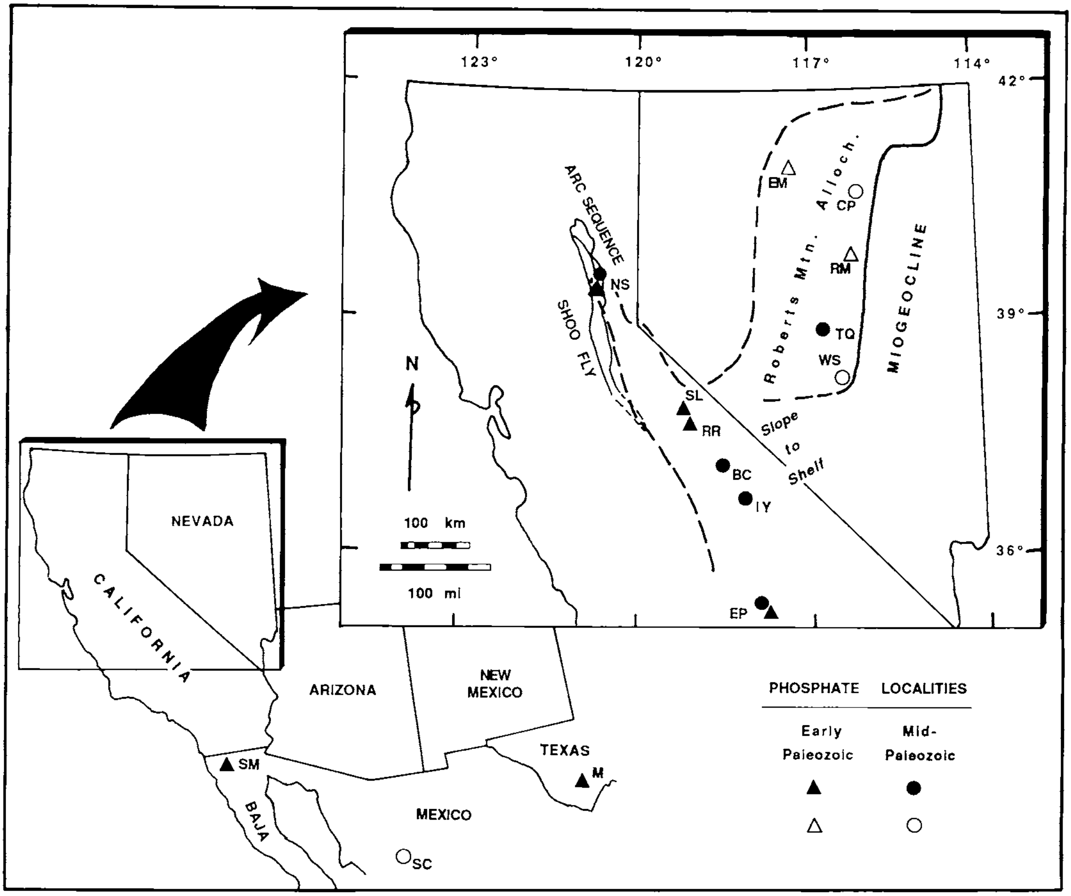


Fig. 1. Lower to middle Paleozoic phosphate in the southern Cordillera. Symbols indicate locations discussed in text: BC, Bishop Creek roof pendant; CP, Carlin-Pinon Range area; EM, Edna Mountain; EP, El Paso Mountains; IY, Inyo Mountains; M, Marathon region; NS, Northern Sierra Nevada; RM, Roberts Creek Mountain; RR, Ritter Range roof pendant; SC, Sierra de Cobachi; SL, Saddlebag Lake roof pendant; SM, San Marcos Rancho; TQ, Toquima Range; WS, Warm Springs. Solid symbols indicate samples analyzed for major or rare earth elements (tables 2, 4); open symbols are other phosphate locations. Triangles denote host rocks of Ordovician-Silurian age; circles are primarily Late Devonian but include Devonian and earliest Mississippian ages. The line of initial $^{87}Sr = 0.7060$ (dashed) indicates the approximate edge of North American Precambrian sialic basement (Kistler and others, 1981). The Cordilleran phosphates occur in both shelf (for example Rogers and others, 1970) and deeper oceanic environments, suggesting a regional phosphogenic province. Geologic provinces compiled from Burchfiel and Davis, 1981; King and Beikman, 1974; Poole, Sandberg, and Boucot, 1977, Schweickert, 1981; and Stewart, 1980.

Stevens (1986) suggested an off-shelf depositional environment of about 750 to 1000 m depth at Vaughn Gulch based on a reconstructed stratigraphic cross section.

Roberts Mountains allochthon.—The third geologic setting lies west of the first two and consists of oceanic rocks in a north–south trending belt (up to 140 km wide, Stewart, 1980) in Nevada known as the Roberts Mountains allochthon (Roberts Mountains terrane of Coney, Jones, and Monger, 1980). The allochthon comprises siliciclastic rocks, chert, and volcanic rocks that lie in thrust sheets above coeval carbonates of the miogeocline (Stewart, 1980). Rocks of the allochthon have been interpreted as slope and rise deposits (Churkin, 1974; Stewart and Poole, 1974; Stanley, Chamberlain, and Stewart, 1977; Wrucke, Churkin, and Heropoulos, 1978). The Devonian to Mississippian Antler orogeny juxtaposed the oceanic and miogeoclinal assemblages (Roberts and others, 1958; Stewart, 1980). Phosphate locations in the Roberts Mountains allochthon (table 1; figs. 1, 2A) include several new ones. Phosphate-bearing sedimentary rocks often also host stratiform barite (Toquima Range, Coles and Snyder, 1985; Papke, 1984, p. 18, p. 115; Warm Springs, Kleinhampl and Ziony, 1984; see also Churkin, 1974). "Spheres and elongate pods of white chert" in the bedded black chert of the Upper Devonian portion of the Woodruff Formation in the Carlin-Pinon area, Nevada (Smith and Ketner, 1975) are included in table 1 as a possible phosphate locality.

Paleozoic metamorphic rocks of California.—The fourth, and westernmost, geologic setting includes metamorphosed oceanic sedimentary rocks in roof pendants in the east-central Sierra Nevada, in the northern part of the Sierra Nevada metamorphic belt, in the eastern Klamath Mountains, and in the El Paso Mountains (all in California), as well as in northern Baja California (King and Beikman, 1976; also Burchfiel and Davis, 1981; Schweickert and Snyder, 1981; and references therein). Paleozoic phosphatic rocks are widely scattered in these sedimentary rocks (table 1; figs. 1, 2C, D, 3A, B). As noted above these rocks lie in terranes whose position relative to one another during the Paleozoic is unresolved.

The Shoo Fly Complex and the Grizzly, Sierra Buttes, and Elwell Formations are part of the Northern Sierra terrane of Silberling and others (1987). Phosphatic rock (fig. 2C) in the Shoo Fly, which is pre-Late Devonian in age, is restricted to the upper part of the Complex (Varga, 1982); other rocks in the Shoo Fly may represent different ages and depositional environments (see Hannah and Moores, 1986; Schweickert and others, 1984; and references therein). Our conclusions, therefore, do not necessarily apply to the Shoo Fly Complex as a whole. The relationship of the rocks of the Shoo Fly Complex to the Antler orogeny is the topic of considerable debate (Burchfiel and Davis, 1975; Speed, 1979; Bond and Devay, 1980; Schweickert and Snyder, 1981; Varga and Moores, 1981; Girty and Wardlaw, 1984).

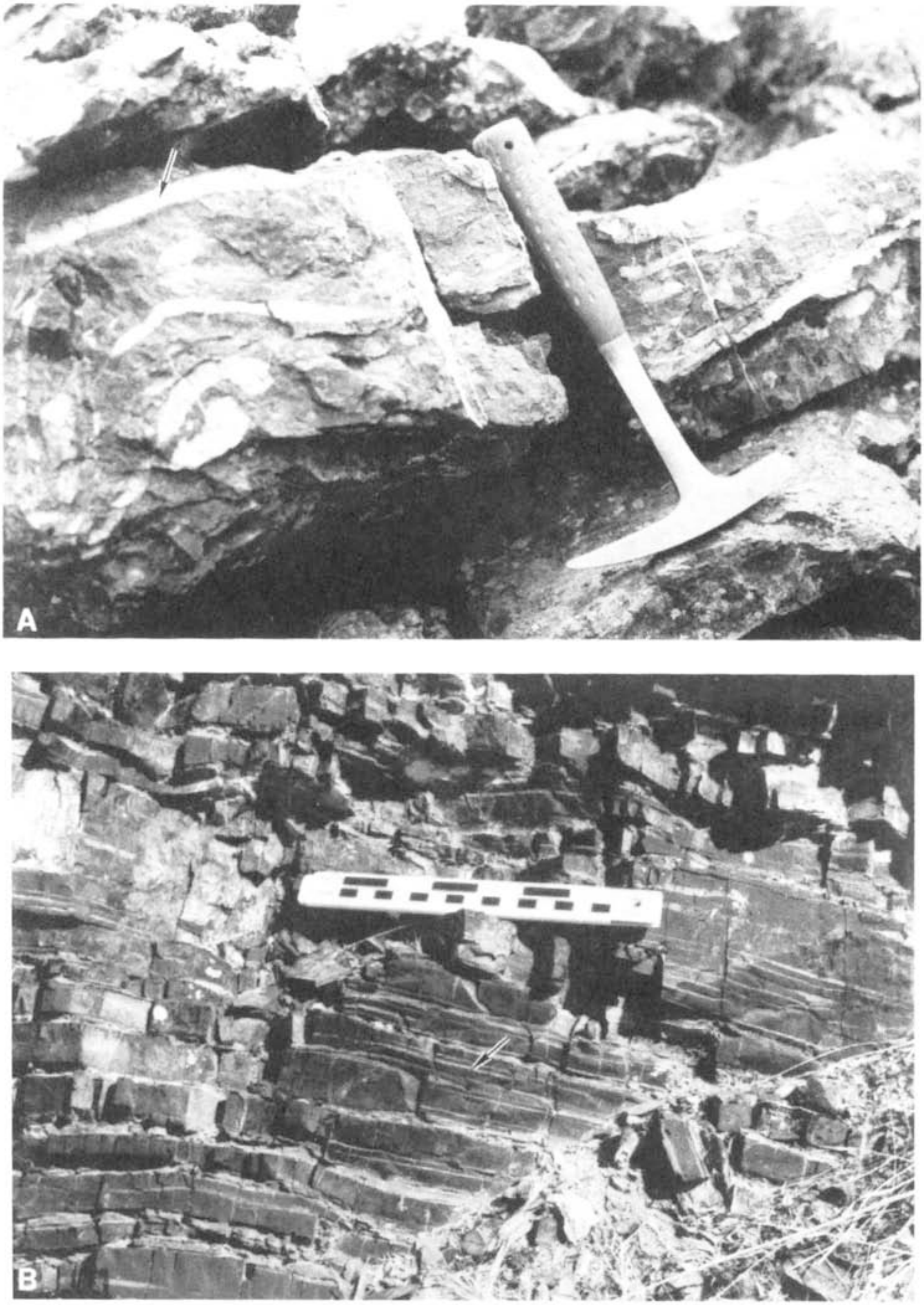


Fig. 2. Photographs of phosphate. (A) Devonian-Mississippian chert in the Toquima Range, central Nevada, containing phosphate lenses (arrow at upper left). (B) Upper Devonian chert of the Squares Tunnels beds, Inyo Mountains, California. Arrow at center indicates phosphate lens. Scale in inches and cm.

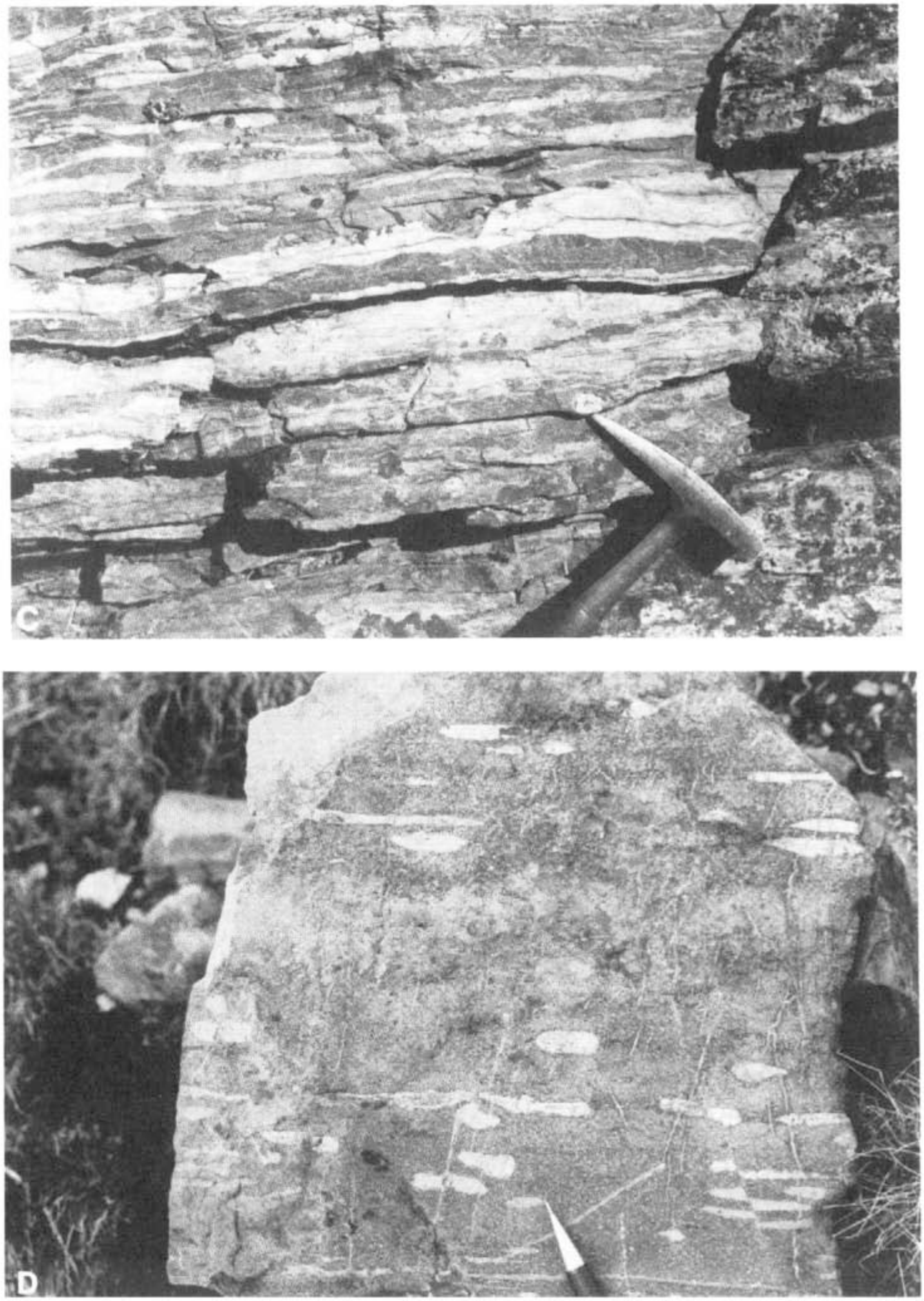


Fig. 2 (continued) (C) Pre-Upper Devonian chert of the Shoo Fly Complex, Lakes Basin region, Sierra Nevada, California. Note phosphate nodule by pick end of hammer. (D) Paleozoic siliceous hornfels of the Bishop Creek roof pendant, Sierra Nevada, California.

Phosphate-bearing rocks east of Saddlebag Lake in the eastern Sierra Nevada are lithologically and structurally similar to parts of the Roberts Mountains allochthon in Nevada according to Schweickert and Lahren (1984; 1987); see also Schweickert (1981); Stewart (1985). Nokleberg (1983), however, placed these rocks in the Owens and High Sierra terranes, which he considered distinct from the miogeocline and Roberts Mountains allochthon to the east and north (Nokleberg, 1983; Silberling and others, 1987).

Phosphate nodules occur (C. Stevens, personal commun., 1983; Frazier and others, 1986) within black pelitic hornfels (fig. 2D) along Coyote Ridge in the Bishop Creek roof pendant (also known as the Coyote Creek roof pendant; Frazier and others, 1986). Moore and Foster (1980) reported Middle Ordovician graptolites from the pelitic hornfels and correlated it with the Johnson Spring Formation of the miogeoclinal section of the Inyo Mountains. They suggested a shelf to slope environment of deposition for the rocks of the Bishop Creek pendant. Frazier and others (1986) collected additional fossils, however, and suggested that the phosphatic part of the section in the Bishop Creek pendant may actually correlate with the Devonian-Mississippian portion of the miogeoclinal section and that the Bishop Creek section originated along the Cordilleran margin south of its present location (Frazier and others, 1986). Nokleberg (1983) and Silberling and others (1987) placed the Bishop Creek pendant in the allochthonous Owens terrane.

Phosphate-bearing argillites are present within the eugeoclinal sequence (term of Carr, Poole, and Christiansen, 1984) of Paleozoic metasedimentary and metavolcanic rocks (Garlock assemblage, Carr, Poole, and Christiansen, 1984; Garlock Formation of Dibblee, 1967) in the El Paso Mountains (table 1; figs. 1, 3A, B; El Paso terrane of Silberling and others, 1987). Our reconnaissance revealed no phosphate in the other parts of the Paleozoic sequence. The origin of the Garlock assemblage is controversial; Burchfiel and Davis (1981) suggested these rocks are out of place with respect to Paleozoic facies in the southwestern United States, requiring large southward transport along a Mesozoic strike-slip fault. Carr, Poole, and Christiansen (1984), however, argued that significant north–south movement relative to other Paleozoic Cordilleran rocks is not required by the facies patterns.

An allochthon of quartzite, chert, phosphatic chert, and carbonate at Rancho San Marcos in northern Baja California occurs within Mesozoic flysch (Peninsular flysch terrane of Gastil, 1985) and is regarded as a giant olistolith emplaced by submarine gravity sliding (Lothringer, 1984; phosphate shown in fig. 3C). Mesozoic and Cenozoic strike-slip faulting has offset the San Marcos rocks from correlative strata to the north and east; possible sources for the allochthon on the continental margin have been discussed by Gastil and Miller (1984) and Gastil (1985).

Marathon Basin.—The Marathon region in west Texas exposes shale, chert, and limestone in a thin oceanic assemblage thrust northwestward onto the southern continental shelf of North America during middle Pennsylvanian through early Permian continent–continent collision (King, 1977; Hickman and others, 1985; Ross, 1986). Phosphate nodules occur within brown to black chert (fig. 3D) of the Upper Ordovician Persimmon Gap Shale (considered the upper member of the Ordovician Maravillas Chert by some workers; King, 1937; McBride, 1970; Ross, 1986) in the Marathon Basin. The unit is characterized in well logs by a prominent gamma ray kick (unpub. data, Unocal), which is consistent with the presence of uranium-bearing apatite.

COMPARISON OF CORDILLERAN AND OTHER KNOWN PHOSPHORITES

Composition.—X-ray diffractograms of the Cordilleran phosphates show that the lenses and nodules contain quartz and apatite; measurement of apatite lattice dimensions by the method of Gulbrandson (1970) indicates less than 1 percent by weight of carbonate (CO_3) in the apatite in samples from the Toquima Range, the Shoo Fly Complex, and the Sierra Buttes Formation. Microprobe analyses (table 2) of samples from several locations (table 1) show the phosphates to be nearly ideal fluorapatites. In particular, the CaO/P_2O_5 and F/P_2O_5 ratios of most samples are very similar to those of the fluorapatite standard (table 2). Significant substitution of CO_3 for PO_4 would be expected to raise the CaO/P_2O_5 ratio, even if CO_3 is not detected by microprobe analysis, but this does not seem to be the case. Neogene and Quaternary phosphorites usually contain CO_3 in the form of francolite (carbonate-fluorapatite, Kolodny, 1981). We suggest that later diagenesis or metamorphism of the Paleozoic phosphate removed the carbonate component from the Cordilleran phosphates, as observed for other deposits (Flicoteaux and Lucas, 1984; McArthur, 1985).

Patterns of deposition.—Quaternary marine phosphorites occur on continental margins, seamounts, and oceanic plateaus at low latitudes (see maps in Kolodny, 1981; Burnett, Roe, and Piper, 1983) and have been grouped into several types by McKelvey (1967). *East-coast type* phosphorite forms along east-facing continental margins and is associated with sandstone, limestone, and dolomite (McKelvey, 1967), or with black shale and chert (Sheldon, in press). The phosphate occurs as pellets, pavements, and sands. Examples include the deposits in the southeastern United States (primarily Miocene, Manheim, Pratt, and McFarlin, 1980; Riggs, 1984). Although Miocene water depths of such deposits are not precisely known, they probably were not significantly deeper than the present-day water depths of 1000 m or less for the Miocene Blake Plateau phosphorites of the southeastern United States (Manheim, Pratt, and McFarlin, 1980). Quaternary east-coast type phosphorite in 200 to 500 m water depth off East Australia (O'Brien and

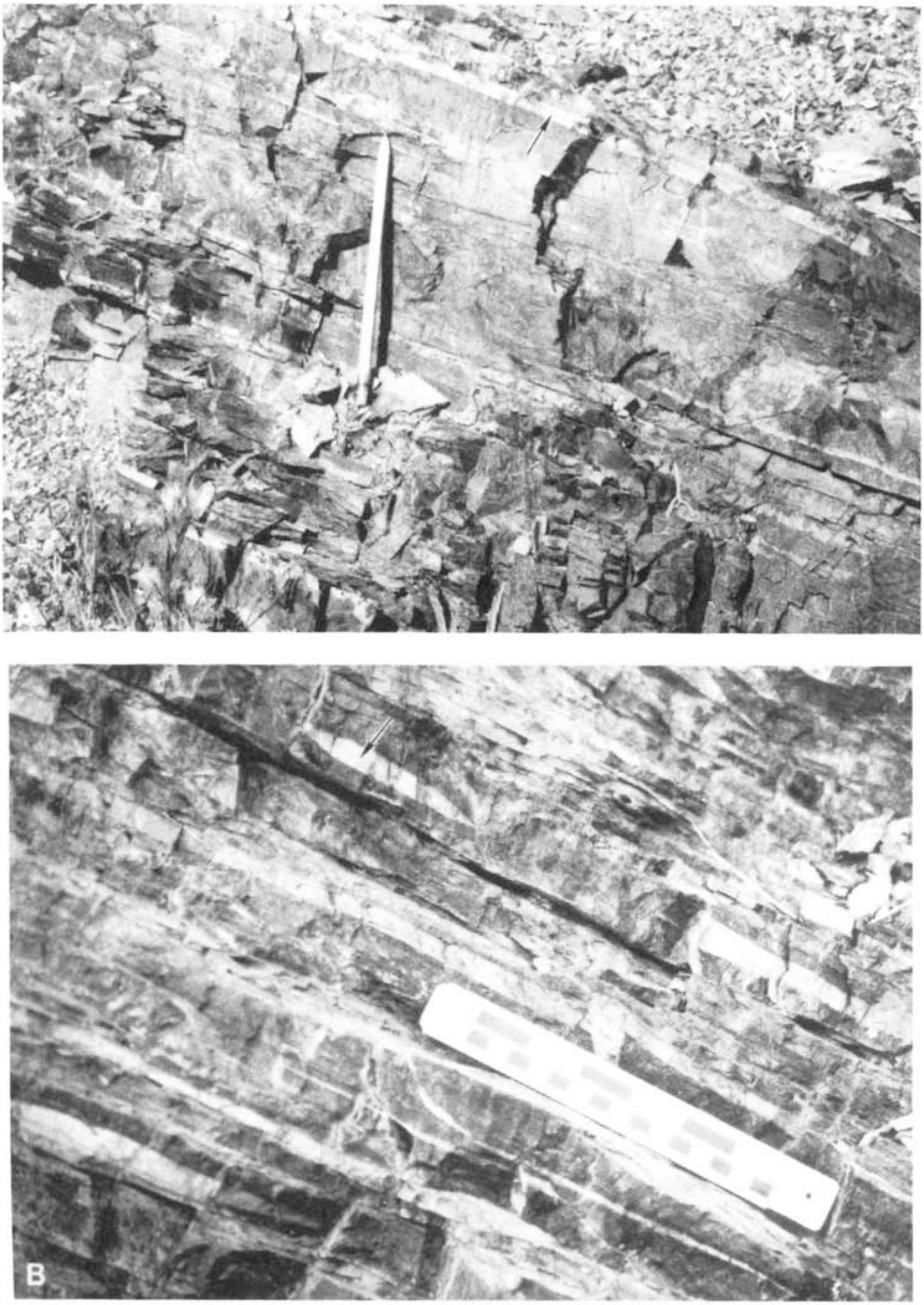


Fig. 3. Photographs of phosphate. (A) Ordovician eugeoclinal chert of the El Paso Mountains, California. Arrow at upper center indicates phosphate lens. (B) Devonian chert, also in the El Paso Mountains. Phosphate lens (arrow) is at upper center. Scale in inches and cm.

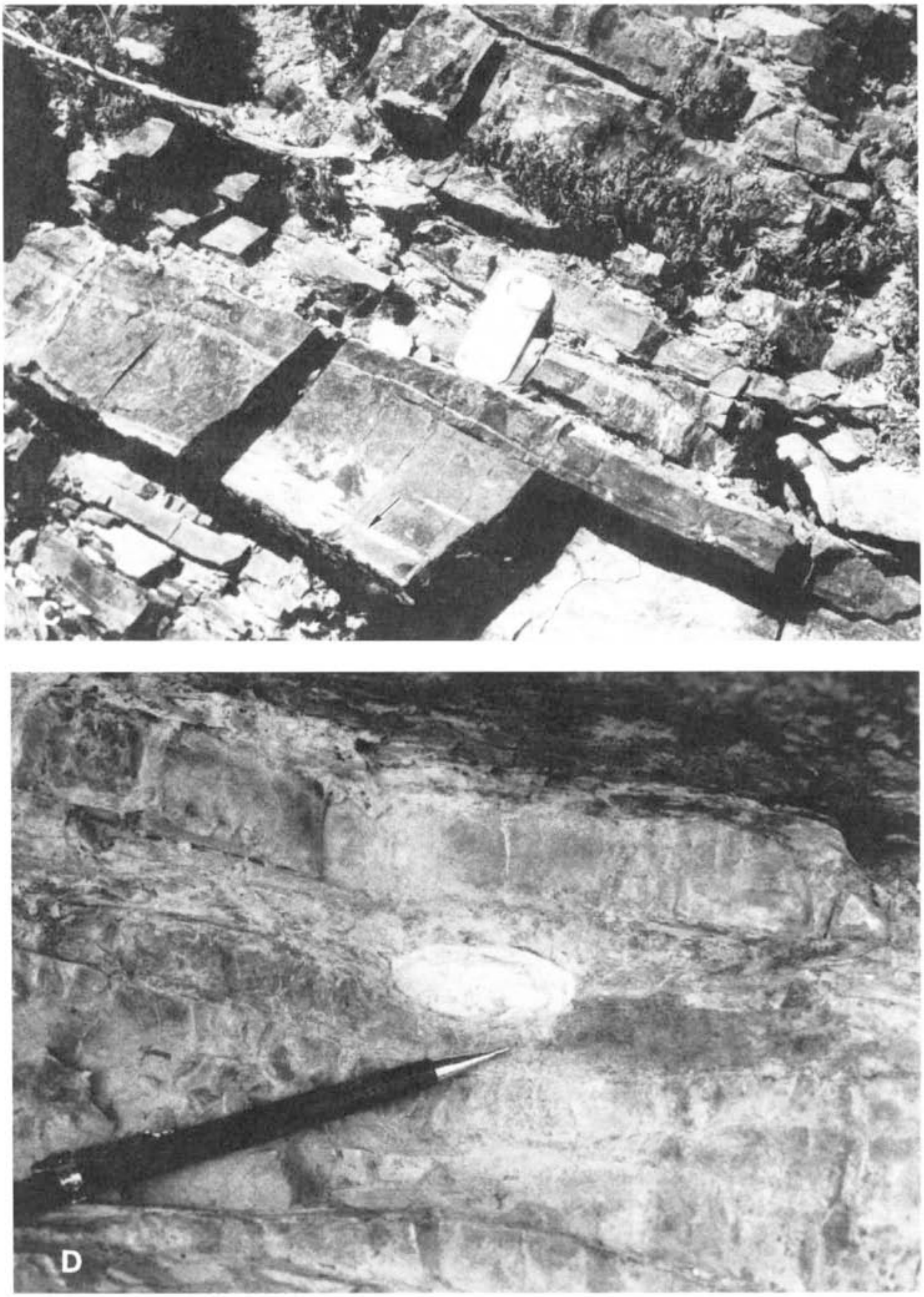


Fig. 3 (continued) (C) Ordovican chert of the San Marcos allochthon, Baja California. Phosphate lens is at lower center (arrow). View is 15 × 20 cm. (D) Phosphate nodule (at pencil point) in Ordovician Persimmon Gap Shale, Marathon Basin, Texas.

TABLE 2

Phosphate analyses by microprobe
Sample locality (number of spot analyses)

Oxide Wt %	Fluor-Apatite Std (2)	Toquima Rg, NV 83-VE-321 (4)	Inyo Mtns 83-INY-305 (2)	Shoo Fly PO-1 (4)	Sierra Buttes JD-1 (4)	Saddlebag 8-13-83-1 (2)	N Ritter (2)	Bishop Ck 8-15-83-1 (4)	El Paso Ord. 1213	El Paso Dev. 1209 (5)	El Paso Dev. 1207 (1)	San Marcos BP-1 (2)	Marathon Basin 3-9-83-1
CaO	55.16	54.54	52.94	52.04	52.40	55.72	55.77	54.63		55.91	52.80	54.78	
P205	40.63	40.43	39.86	39.96	40.07	40.18	37.42	41.48	TOO	40.59	40.49	39.89	TOO
F	4.44	4.26	6.33	3.56	4.33	3.78	4.06	3.83		2.60	3.37	3.43	
Na2O	0.25	0.14	0.08	<0.05	0.13	<0.05	<0.05	<0.05	FINE-	<0.05	<0.05	<0.05	FINE-
MgO	<0.05	<0.05	<0.05	<0.05	<0.05	<0.05	<0.05	<0.05	GRAINED	<0.05	<0.05	<0.05	GRAINED
SO3	0.28	0.11	<0.1	<0.1	0.13	<0.1	<0.1	<0.1		<0.1	<0.1	<0.1	
SrO	0.43	0.38	<0.3	<0.3	0.36	<0.3	<0.3	<0.3	FOR	<0.3	0.66	0.46	FOR
MnO	<0.1	<0.1	<0.1	N.M.	<0.1	<0.1	<0.1	<0.1		<0.1	<0.1	<0.1	
									ANALYSIS				ANALYSIS
K2O	<0.05	<0.05	0.22	<0.05	<0.05	<0.05	<0.05	<0.05		<0.05	<0.05	<0.05	
SiO2	0.13	0.12	3.46	1.71	<0.05	0.93	5.60	0.47		1.03	0.15	0.17	
FeO*	<0.2	<0.2	<0.2	N.M.	0.79	<0.2	<0.2	0.18		<0.2	<0.2	<0.2	
TOTAL	101.32	99.98	102.89	97.27	98.21	100.61	102.85	100.59		100.13	97.47	98.73	
−F = 0	1.87	1.79	2.67	1.50	1.82	1.59	1.71	1.61		1.09	1.42	1.44	
TOTAL	99.45	98.19	100.22	95.77	96.39	99.02	101.14	98.98		99.04	96.05	97.29	
CaO/P205	1.36	1.35	1.33	1.30	1.31	1.39	1.49	1.32		1.38	1.30	1.37	
F/P205	0.11	0.11	0.16	0.09	0.11	0.09	0.11	0.09		0.06	0.08	0.09	

Data for each sample are average of <10 μm spot analyses, the number of which are given in parentheses, of phosphate in a single field sample. Measurement error (one standard deviation): CaO and P205, ≤0.5 wt %; F, 1.2 wt % (Sierra Buttes), ≤0.25 wt % for others; SiO2, ≤0.1 wt %.

Some samples were too finely intergrown with quartz to permit analysis for apatite alone; this may also account for the low totals.

* = Includes total Fe as FeO.

N.M. = Not measured.

others, 1986) is in a zone of very limited upwelling and organic productivity (O'Brien and Veeh, 1983).

West-coast and *Tethyan type* phosphorites form along west-facing continental margins in zones of coastal upwelling and high organic productivity, and along the margins of narrow, east-west trending, low latitude (Tethyan) seaways. The phosphate occurs as nodules and pellets associated with black shale and chert (McKelvey, 1967). Quaternary west-coast type deposits occur on the continental margins of Peru/Chile and Namibia (Burnett, Roe, and Piper, 1983). Recent phosphorites off Peru/Chile formed at the margins of the oxygen-minimum zone where it impinges on the continental shelf and slope, at depths of 500 m or less (Burnett, 1980; Burnett, Roe, and Piper, 1983). These phosphorites have formed in or adjacent to laminated, anaerobic sediments (Burnett, 1980; Burnett, Roe, and Piper, 1983).

Seamount type phosphorites form on seamounts in open oceans, typically far from continents, and are associated with volcaniclastic rocks and limestone, commonly as a replacement product (McKelvey, 1967). The source of phosphorus may be insular guano deposits in a few cases but commonly is unknown (Baturin, 1982; Burnett, Roe, and Piper, 1983). Examples include deposits on the Musician seamounts (McKelvey, 1967; Burnett, Roe, and Piper, 1983). Seamount type phosphorites now occur at depths as great as 4000 m (mostly around 2000 m, Sheldon, 1980); possibly these have subsided on old seamounts.

A few *moderately-high-latitude* phosphorites of Miocene age may reflect upwelling of the Antarctic circumpolar current near bathymetric highs (Sheldon, in press). The phosphate is not associated with chert. An example is the Miocene phosphate on Chatham Rise at depths of 200 to 500 m (Cullen, 1980).

The geographic distribution of Quaternary phosphorite deposits is confined to about 40° latitude or less (Sheldon, 1964; Kolodny, 1981; Burnett, Roe, and Piper, 1983). Paleolatitudinal reconstructions of Phanerozoic phosphorite deposits show that, except for the rare moderately-high-latitude deposits that lack chert, they all formed within about 50° latitude of the equator (Sheldon, 1964; Cook and McElhinny, 1979). Both the Quaternary and Phanerozoic cases have peaks at 10° to 30° latitude (Sheldon, 1964; Cook and McElhinny, 1979). Thus, the sites of formation of Quaternary and Phanerozoic phosphorites are restricted geographically, with respect to latitude and, for the west-coast type, with respect to the oxygen-minimum zone.

Rare earth patterns in marine sediments.—Rare earth element (REE) distributions in marine sedimentary rocks may provide information about depositional setting (Shimizu and Masuda, 1977; Jenkyns and Winterer, 1982). Dissolved REE enter the oceans largely via river water and leaching of aeolian particles from continents. Settling sedimentary or biogenic particles scavenge the REE from seawater and transport them to the sediment column (Kolodny, 1981; Elderfield and Greaves, 1982; de Baar, Bacon, and Brewer, 1985). Cerium (Ce) is removed from seawater more rapidly than the other REE, because it oxidizes to a tetravalent state less soluble in seawater than the other, trivalent, REE (Elderfield and Greaves, 1982; de Baar, Bacon, and Brewer, 1985). Thus the REE pattern of seawater in the deep ocean is characterized by a negative Ce anomaly (fig. 4, table 3; Høgdahl, Melsom, and Bowen, 1968; de Baar, Bacon, and Brewer, 1983; 1985). Both the shallow mixed layer of Atlantic Ocean water (de Baar, Bacon, and Brewer, 1983) and seawater near continents (result of Høgdahl, noted in Piper, 1974), on the other hand, do not have negative Ce anomalies and may even have positive Ce anomalies (de Baar, Bacon, and Brewer, 1983).

Shimizu and Masuda (1977) observed that deep-sea cherts formed within large oceanic basins have substantial negative shale-normalized Ce anomalies, relative to the REE as a group. They interpreted this anomaly as the result of silica precipitation from open ocean seawater which was itself depleted in Ce (North Atlantic Deep Water in fig. 4). REE patterns from most ancient chert sequences exposed on land do not have negative Ce anomalies. These cherts have therefore been interpreted as continental margin, marginal basin, or epeiric sea deposits (Shimizu and Masuda, 1977; Jenkyns and Winterer, 1982) precipitated from seawater that lacks a negative Ce anomaly. While the smaller-scale distribution of Ce in seawater and modern marine sediment is not yet well documented, the first order geographic variation of the Ce anomaly suggests it may be useful as an indicator of proximity to a continent.

Measurement and normalization of rare earth abundance.—A number of methods exist for measurement of REE abundances. Isotope dilution is the most precise and most sensitive technique. Measurements using neutron activation analysis can be contaminated by U or Th. Plasma emission spectroscopy is a fast technique, although it requires care in sample preparation. Not all investigators report measurement error or methods of sample preparation. While the precision of measurements of individual elements varies, useful comparisons of overall REE abundance, anomalies of single elements, and depletion of light versus heavy REE are possible.

Several normalizations of the Ce anomaly appear in the literature (these are defined and compared in app. 1); we quantify the Ce anomaly as follows:

1. All data are normalized to average shale values of Piper (1974):

$$Ce_{SN} = Ce(ppm)/Ce(ppm)_{SHALE} \quad (1)$$

where Ce_{SN} = shale-normalized cerium concentration.

2. The value for Ce_{SN} expected if the REE pattern were linear between La_{SN} and Nd_{SN} on the log-linear plot (fig. 4; table 3) is called Ce^{*}_{SN}:

$$\log Ce^{*}_{SN} = (2 \cdot \log La_{SN} + \log Nd_{SN})/3 \quad (2)$$

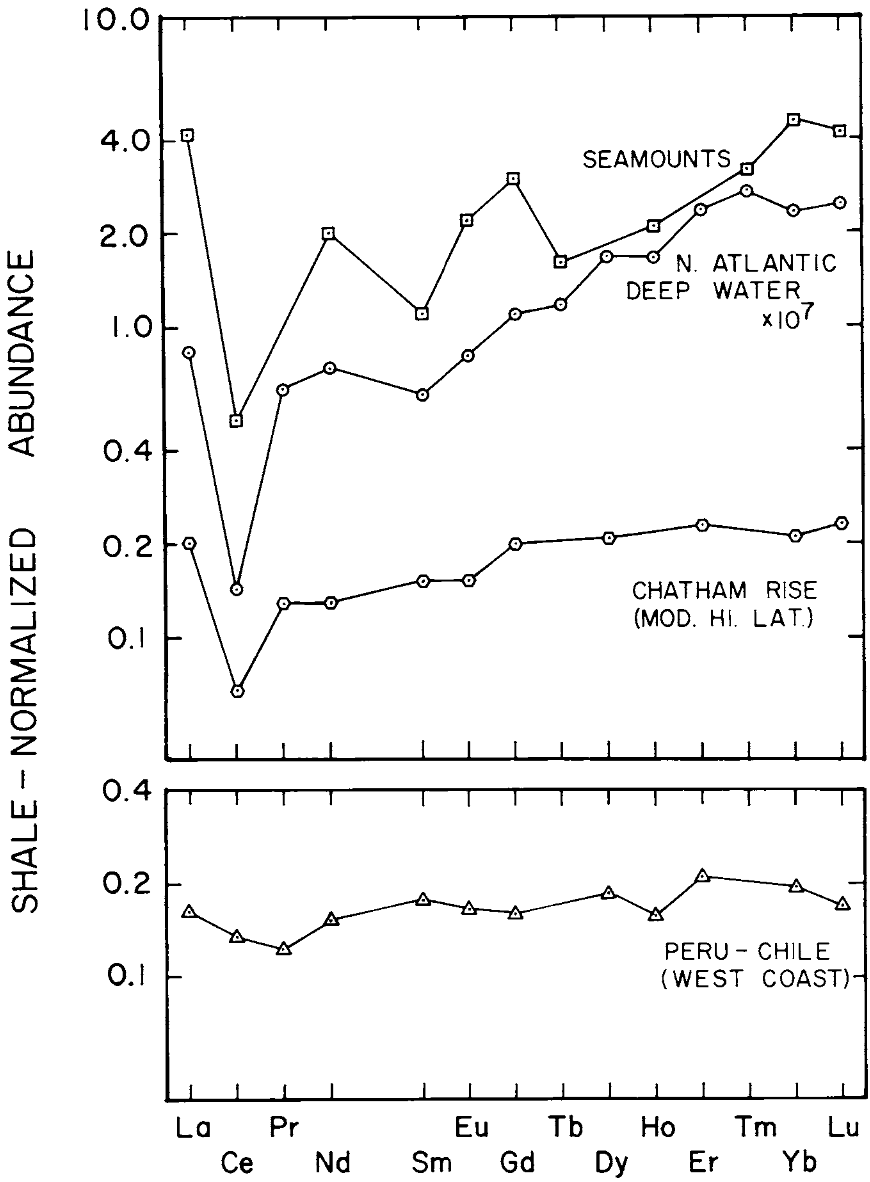


Fig. 4. Abundances of rare earth elements in phosphorites and North Atlantic Deep Water (NADW × 10^7). Values are normalized to average shale of Piper (1974) and plotted against atomic number. Note the overall trends and the abundance of cerium relative to the trend. Cerium anomalies for these curves are given in table 3.

3. The observed value of Ce_{SN} divided by Ce^*_{SN} is the Ce anomaly:

$$\text{Ce anomaly} = Ce_{SN}/Ce^*_{SN} \quad (3)$$

4. Another useful value is the unnormalized ratio of Ce to La:

$$\text{Ce (in ppm)/La (in ppm)} = \text{Ce/La ratio} \quad (4)$$

Readers accustomed to chondrite-normalized REE plots should note that the slope from light to heavy REE will be more negative (that is, rotated clockwise) on a shale-normalized plot than on a chondrite-normalized plot. Anomalies of Ce (or other REE, except Eu) do not differ substantially between shale-normalized and chrondrite-normalized plots as average shale (Piper, 1974) has no Ce anomaly relative to chondrites (Masuda, Nakamura, and Tanakall, 1973). Burnett, Roe, and Piper, (1983) noted that the normalizing value of Piper (1974) for Tb may be too large, giving normalized Tb values that are too low and causing apparent negative Tb anomalies that are not real (fig. 5).

Rare earths in Neogene and Quaternary phosphorites.—Abundances of rare earth elements in phosphorites vary with the type of phosphorite deposit (fig. 4; table 3). Note in the examples in figure 4 and table 3 that the negative Ce anomaly is most pronounced in the seamount phosphorite (bulk sample, 0.15) and North Atlantic Deep Water (0.18). The Chatham Rise example (moderately-high-latitude type phosphorite; Miocene) is less Ce-depleted (francolite, 0.38), and the Peru-Chile deposit (west-coast type; Quaternary) is undepleted (francolite, 0.85). Another difference is that the overall trend of the examples near continents is flatter than the light REE depleted pattern of the open

TABLE 3

Ce in Neogene and Quaternary phosphorites

	Ce anomaly (Ce_{SN}/Ce^*_{SN})	*Ce/La ratio* (Ce in ppm/ La in ppm)
North Atlantic deep water	0.18	0.35
Phosphate from Pacific seamounts	0.15	0.24
Phosphate from Chatham Rise (Mod. Hi. Lat.)	0.38	0.67
Phosphate from Peru-Chile (West Coast)	0.85	1.67
Average shale	1.00	2.02
Leedey chondrite	1.00	2.58

Sources and methods used:

North Atlantic deep water: Høgdahl, Melsom, and Bowen (1968); neutron activation (avg of 12 water samples).

Pacific seamounts: Burnett, Roe, and Piper (1983); neutron activation (avg of suite of bulk samples).

Chatham Rise: McArthur and Walsh (1984); plasma emission spectrometry (avg of HCl leaches of three samples).

Peru/Chile: McArthur and Walsh (1984); plasma emission spectrometry (avg of HCl leaches of five samples).

Average shale: Piper (1974); modified from other published analyses.

Leedey chondrite: Masuda, Nakamura, and Tanaka (1973); isotope dilution.

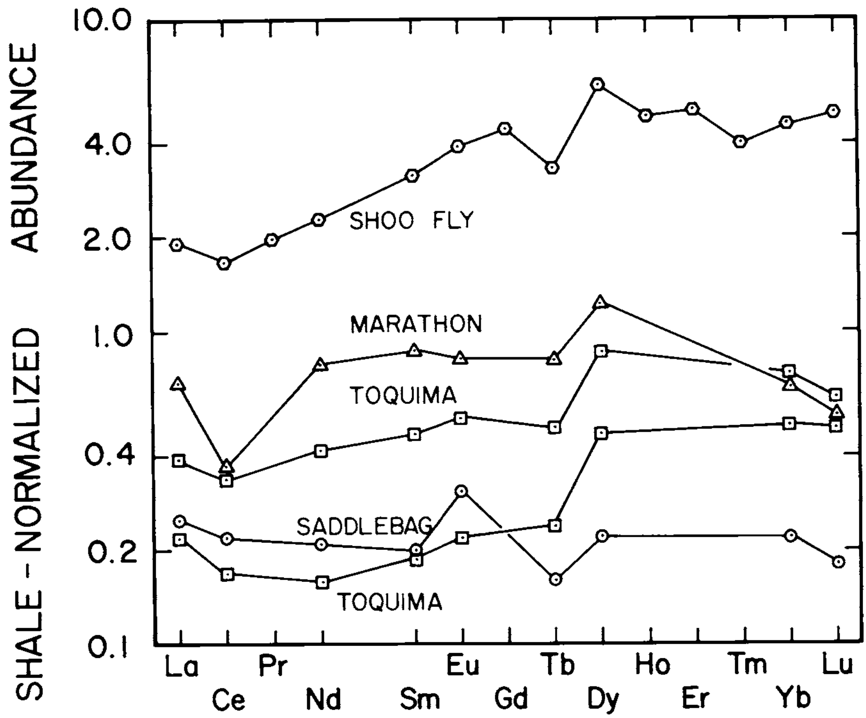


Fig. 5. Abundances of rare earth elements in Cordilleran phosphates given in tables 4 and 5 plotted as in figure 4. Except for the Marathon and the possibly enriched Shoo Fly samples, the trends most closely resemble the Peru-Chile curve of figure 4. See text.

ocean examples. This is in spite of the possible presence of lithogenic minerals in the seamount sample; any such minerals do not introduce enough continental REE to mask the oceanic signal. Taken together, the data suggest that proximity to a continent affects the REE pattern of francolite in Quaternary phosphorite deposits (Burnett, Roe, and Piper, 1983). O'Brien and Veeh (1983) proposed the use of REE patterns to distinguish between upwelling phosphorites (that is, west-coast and Tethyan types) having shale-like REE patterns without negative Ce anomalies and phosphorites from non-upwelling areas (east-coast type) having patterns more like open ocean water in that they have Ce anomalies and may be somewhat depleted in light REE.

Alteration of rare earth patterns in phosphorites.—Data from pre-Quaternary phosphorites and conodonts (McArthur and Walsh, 1984; Wright, Seymour, and Shaw, 1984: Wright, Miller, and Holser, 1987) indicate additional possible controls on REE patterns in phosphate minerals. First, oxidation-reduction conditions during deposition influence the uptake of Ce (de Baar, Bacon, and Brewer, 1985: McArthur and Walsh, 1984; Wright, Seymour, and Shaw, 1984), the rate of which differs in pelletal and nodular phosphorites (Burnett, Roe, and Piper, 1983; McArthur and Walsh, 1984). Second, weathering leaches REE, acting preferentially on light REE (McArthur and Walsh, 1984). Enrichment of REE in francolite is common during diagenesis and may also occur during metamorphism (McArthur and Walsh, 1984). Third, variation of the REE content of the oceans may have occurred over time, but whether this variation was spatially uniform is still uncertain (Wright, Seymour, and Shaw, 1984; Wright, Miller, and Holser, 1987).

Deep sea and continental REE patterns may alter to resemble one another, but some generalizations about environment are still possible. McArthur and Walsh (1984) observed that a deep sea francolite REE trend could become shale-like during diagenetic enrichment if shale was present and non-oxidizing conditions prevailed. Similarly, depletion by weathering could make a shale REE trend more like a deep sea signal. Based on their data McArthur and Walsh also suggested that francolite with a strong negative Ce anomaly indicates a deep seawater source, directly or indirectly, of REE. Positive or zero Ce anomalies may indicate diagenetic uptake of REE from adjacent rocks under non-oxidizing conditions (McArthur and Walsh, 1984) or a shale-like (for example, continental) source of REE (see also Altschuler, Berman, and Cuttitta, 1967; Kolodny, 1981). It appears that a shale-like REE pattern in francolite lacking a Ce anomaly would either be primary or require a nearby shale source of secondary REE. Either explanation could account for the lithogenic minerals and associated francolites in Peru-Chile, which differ in REE abundances but have similar REE patterns (compare Burnett, Roe, and Piper, 1983 with McArthur and Walsh, 1984). Analysis of bulk phosphate may thus reveal the continental REE signature.

Because of the complications noted above, we limit our interpretation of REE patterns as follows (compare O'Brien and Veeh, 1983). Phosphatic rocks that lack a negative Ce anomaly and have an overall shale-like REE trend more likely formed in a continental-margin setting than a deep-sea setting. Phosphatic rocks having a negative Ce anomaly and depleted in light rare earths relative to shale are interpreted as most likely to be of deep-sea origin. Although this limited interpretation gives only a general indication of depositional setting, we suggest that REE may prove useful in study of accreted terranes, whose origin often is uncertain. We caution that future work will require a larger set of REE patterns from known settings, understanding of alteration effects, and perhaps should search for a mineral whose REE are less mobile than francolite.

Rare earth patterns in Cordilleran phosphates.—Analyses of Lower to middle Paleozoic bulk phosphate nodules from California and Nevada yield patterns that are flat to slightly depleted in light REE and lack Ce anomalies (fig. 5; tables 4, 5), permissive of a continental margin origin. Although two samples have poorly known Nd values, the overall trends

TABLE 4

REE analyses of Cordilleran phosphates

	Paleozoic samples:						
Element (in ppm)	Marathon 3-9-81-1	Saddlebag 8-13-83-1	Toquima 83-VE-321a	Toquima 83-VE-321b	Shoo Fly	Average Shale (Piper, 1974)	Detection Limit
U	4.5	0.8	24.4	20.2			0.1
Th	1.2	1.0	1.8	(0.4)			0.2
Sc	6.2	1.8	4.0	3.4			0.01
Y*					200		1.0
La	28.1	10.3	16.1	9.0	78.7	41	0.1
Ce	31	18	28	14	139	83	1.0
Pr*					20	10.1	2.0
Nd	30	(8)	16	(6)	87	38	3.0
Sm	6.49	1.49	3.50	1.43	23.5	7.5	0.1
Eu	1.31	0.49	0.85	0.35	6.24	1.61	0.05
Gd*					28	6.35	2.0
Tb	1.0	(0.2)	0.6	(0.3)	4.1	1.23	0.1
Dy	6.7	(1.2)	4.8	2.6	33.6	5.5	0.5
Ho*					6.5	1.34	0.5
Er*					19.0	3.75	0.5
Tm*					2.5	0.63	0.5
Yb	2.38	0.79	2.56	1.75	16.1	3.53	0.05
Lu	0.33	0.11	0.37	0.30	3.04	0.61	0.01

*Analyses by DC Plasma.

Analyses of other elements by instrumental neutron activation analysis and, expect for Shoo Fly sample, corrected by uranium fission. Values in parentheses have analytical error >20 percent. Error estimates are ±5 percent for values >100 x detection limit; ±10 percent for values 10 x to 100 x detection limit; ±detection limit for values <10 x detection limit.

TABLE 5
Ce in Cordilleran phosphates

Sample, host rock, location	*Ce anomaly* (Ce_{SN}/Ce^*_{SN})	*Ce/La ratio* (Ce in ppm/ La in ppm)
3-9-81-1; Persimmon Gap Shale (Upper Ordovician); Marathon Basin, Tex.	0.51	1.10
8-13-83-1; Chert in Saddlebag Lake Roof Pendant (Ordovician/Silurian?); Sierra Nevada, Calif.	0.92	1.75
83-VE-321a; Chert of Robert Mtns. Allochthon (Upper Devonian); Toquima Range, Nev.	0.85	1.74
83-VE-321b; Chert of Roberts Mtns. Allochthon (Upper Devonian); Toquima Range, Nev.	0.85	1.56
Shoo Fly Complex (pre-Late Devonian age); Northern Sierra Nevada, Calif.	0.82	1.77

and presence or lack of a Ce anomaly are nonetheless evident. One sample (Shoo Fly) was not corrected for U fission, but REE measured by plasma spectroscopy, and hence unaffected by U abundance, fit smoothly into the pattern. The reader should bear in mind that normalized values of Tb may be too small, and that significant errors (10 percent or more) apply in most cases (table 4). Hence the Cordilleran patterns may not exactly match one another (fig. 5) or patterns from known settings (fig. 4).

The REE abundances in the Shoo Fly sample are as much as an order of magnitude higher than the other samples. Enrichment of REE during metamorphism may be responsible (McArthur and Walsh, 1984). As noted above, CO_3 loss from apatite may also be due to metamorphism, and we cannot rule out the possibility that some samples may have lost REE at the same time. The two Toquima Range samples do not show significantly different Ce anomalies, given the measurement errors. Their relative abundances and trends, however, are very similar to those of two Miocene samples of McArthur and Walsh (1984), in which greater weathering in one sample removed REE and depleted light REE relative to heavy REE. Alternatively, the content of lithogenic minerals may vary between the Toquima samples.

Note that in all the samples in figure 5, except the possibly enriched Shoo Fly sample, the overall trend is relatively flat, more like the Peru-Chile and the Chatham Rise trends than the seamount trend in figure 4. The California/Nevada samples lack a strong negative Ce anomaly (values of 0.82–0.92 in table 5), more like the Peru-Chile average (0.85 in table 3) than the seamount (0.15) or Chatham Rise (0.38) phosphorite. The Marathon Basin sample, however, has a moderate negative Ce anomaly (0.51), which may reflect a contrasting setting away from strong upwelling, on the southern margin of Paleozoic North America, or a contrasting diagenetic history. We cannot rule out the possibility that the primary REE signal is not preserved in the Cordilleran phosphates. Nevertheless, the patterns reflect continental input, either as primary signal in apatite, in REE added to apatite during alteration, or in coexisting detrital minerals. Our REE results *do not require* a deep, open oceanic origin for the Cordilleran phosphates.

DISCUSSION

Depositional environment and water depth of Cordilleran phosphates.— We suggest that the geographically widespread Paleozoic phosphates of the southern Cordillera, western United States formed in relatively shallow water associated with upwelling and high rates of marine productivity. Deposition of Quaternary phosphorite is restricted to water depths generally less than 1000 m (Bromley, 1967; Kolodny, 1981; Burnett, Roe, and Piper, 1983). The formation of organic-rich chert with phosphate is commonly ascribed to a detritus-starved low-oxygen environment in zones of upwelling of deep water (McKelvey, 1967). Does this simple picture adequately account for the Cordilleran phosphate deposits?

Not all ancient phosphorites resemble the Quaternary Peru-Chile example (Bentor, 1980; Sheldon, 1980), and not all phosphorites reliably indicate upwelling (Burnett, Roe, and Piper, 1983). Nevertheless, phosphorites associated with organic-rich shale and chert and having shale-like REE lacking a Ce anomaly are good candidates for deposits formed in zones of upwelling (O'Brien and Veeh, 1983; Burnett, Roe, and Piper, 1983). The lower to middle Paleozoic phosphate of the southern Cordillera, western United States closely resembles that of the modern Peru-Chile deposits in terms of REE patterns and lithologic associations and probably formed in a similar environment.

Were upwelling conditions favorable for formation of organic-rich phosphate more widespread in the Paleozoic than in the Quaternary? Widespread black shale (Berry and Wilde, 1978) and substantial phosphorite (Piper and Codispoti, 1975; Cook and McElhinny, 1979; P. Cook, personal commun., 1986), of both mid-Ordovician and Late Devonian age, occur in the world-wide geologic record. As previously noted, stratiform barite deposits, which are known along the length of the North American Cordillera (Noll, 1981; Pregger, 1981; Papke, 1984; Turner, 1984; 1985; Ketner, 1986), are often associated with the Cordilleran phosphates. Although the formation of stratiform barite is a complex process (Papke, 1984), Miller, Brobst, and Beck (1977) felt that low-oxygen bottom conditions are necessary and considered impingement of the oceanic oxygen minimum layer on the continental margin a plausible cause of such conditions. The oxygen-minimum zone probably occupied a wider range of depths at times in the Paleozoic than at present (as deep as 2000 m in the past according to Fischer and Arthur, 1977; Berry and Wilde, 1978), and it probably impinged on the sediment surface over wider geographic areas. Nonetheless, temporal changes in ocean oxygenation were probably not the sole factor controlling the deposition of organic-rich chert and phosphate. The setting of the Peru-Chile deposits suggests that strong upwelling is required as well. An additional factor restricting lower to middle Paleozoic phosphate deposition is the requirement of low sedimentation rates. We argue that upwelling and favorable sites of deposition were geographically restricted during the early to middle Paleozoic as during the Quaternary, though preservation of organic-rich sediment was likely over a greater range of depths, possibly as deep as several thousand meters. The environment suited for deposition of upwelling coastal type phosphorite was limited in extent in the Paleozoic, though it was probably more widespread than in the Quaternary.

Paleontologic and stratigraphic depth estimates are consistent with our estimates for phosphate deposition. *Nereites* assemblage trace fossils are characteristic of the Ordovician and Devonian strata in the Roberts Mountains allochthon (Chamberlain, 1977; 1979; Stanley, Chamberlain, and Stewart, 1977). The trace fossil assemblage is indicative of a bathyal to abyssal environment, corresponding to continental slope and rise deposition (Chamberlain, 1979, and references therein). As noted above, Stevens (1986) reconstructed a regional stratigraphic cross section in the Inyo Mountains, California and inferred water depths of 750 to 1000 m (that is, bathyal) for the Upper Devonian phosphatic chert.

Vesicles in pillow lavas in the Roberts Mountains allochthon suggest the pillows formed in deep water. Moore (1965) derived curves of the size of vesicles in pillow lavas versus depth from study of very young tholeiites of constant chemistry and volatile content. The absence of significant vesicles in tholeiitic pillow lavas in Ordovician rocks of the Roberts Mountains allochthon led Wrucke, Churkin, and Heropoulos (1978) to estimate that most of the pillows were erupted in 4000 or more meters water depth. This conflicts with depth estimates based on the presence of phosphatic rocks. We note, however, that none of the localities of deep-water Ordovician pillow lavas of Wrucke, Churkin, and Heropoulos (1978) coincides in both age and location with places where we have identified phosphatic chert and shale. Wrucke, Churkin, and Heropoulos suggested that Ordovician strata in the Roberts Mountains allochthon are divisible into easterly slope deposits and westerly abyssal deposits. If Wrucke, Churkin, and Heropoulos are correct, our work suggests that the phosphate-bearing rocks are part of the easterly slope deposits and predicts that the more westerly abyssal rocks, whose pillow lavas lack vesicles, should lack phosphate-bearing rocks.

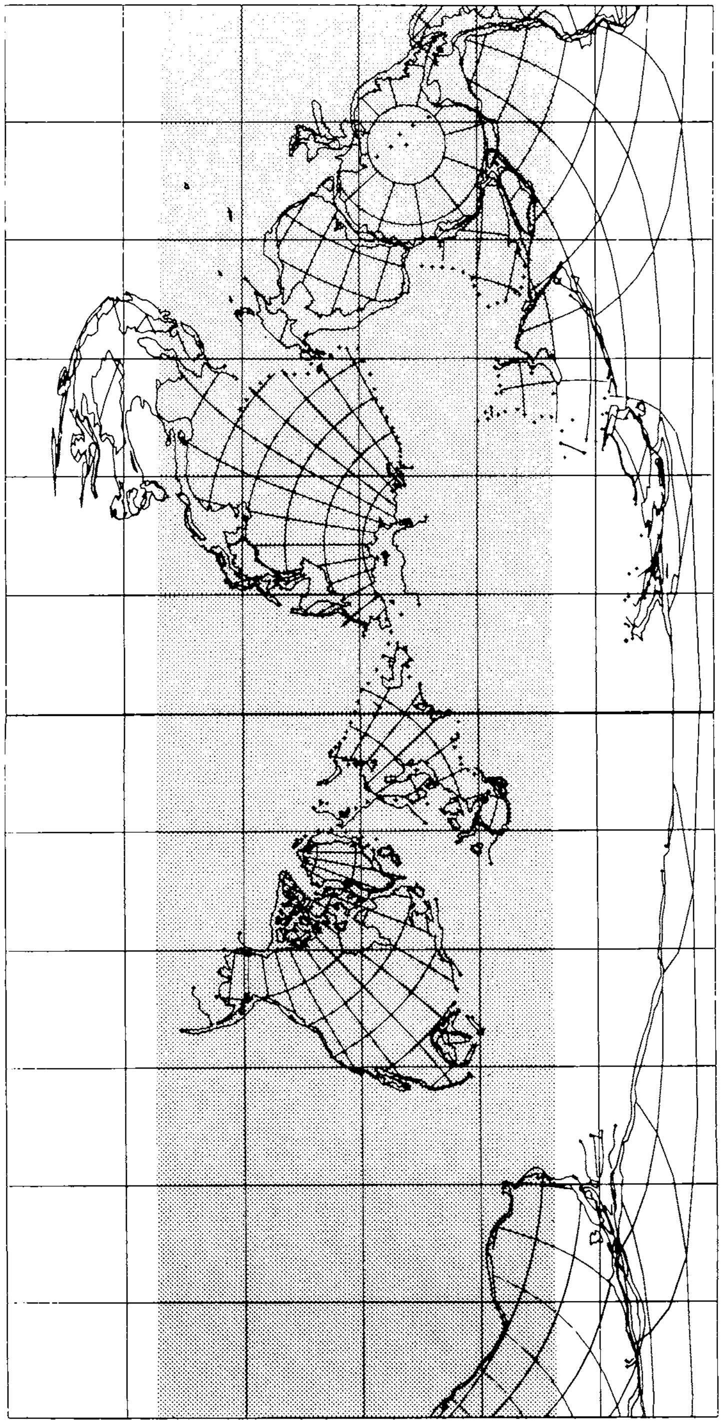
Paleogeographic setting.—Lower- to middle-Paleozoic phosphatic rocks in the southern part of the Cordillera, western United States closely resemble continental margin phosphate deposits in terms of both lithologic associations and REE patterns. By analogy with modern phosphorites, they were probably deposited within 50° of the equator and in a few thousand meters, or less, water depth. Where were the phosphate-bearing rocks originally located?

Figure 6 shows the paleocontinental maps of Smith, Hurley, and Briden (1981) for the Late Ordovician and the Late Devonian, the approximate periods of phosphate deposition in the southern Cordillera of the western United States. According to these maps, only a few continental margins occupied latitudes favorable for chert-hosted phosphate formation (shaded bands on fig. 6). Significantly, the Cordilleran margin of North America was ideally situated for phosphorite deposition throughout this extended span of time (see also Irving, 1979; Scotese, 1984; Miller and Kent, 1986, in support of this observation). Models of Late Devonian paleogeography of North America (fig. 81 of Ettensohn and Barron, 1981; fig. 12 of Parrish, 1982) indicate that trade-wind-driven upwelling probably occurred along the Cordilleran margin. The Transcontinental arch exposed carbonate and was not shedding significant siliciclastic detritus to the Cordilleran margin (Ettensohn and Barron, 1981), permitting local starved basin conditions. The paleocontinental maps are consistent with the hypothesis that the phosphate-bearing rocks in the southern Cordillera formed adjacent to North America in the early to middle Paleozoic (Varga, 1982; Coles and Snyder, 1985). Transport from a distant continental margin is not required by the phosphate, but transport and rearrangement along the margin probably occurred.

The phosphogenic province preserved in the southwestern United States extended along much of western North America. Pregger (1981) and Noll (1981) described Upper Ordovician to Upper Devonian black shale, chert, and stratiform barite in the northern Sierra de Cobachi region of Sonora, Mexico. Phosphate is present in the sequence (Pregger, personal commun., 1984). Cambrian and Ordovician rocks of the Misty Creek Embayment and Silurian rocks of the Selwyn Basin, in the miogeocline in British Columbia and in Yukon and Northwest Territories, Canada (Monger and Berg, 1987; Cecile and Norford, 1985), contain phosphate in shale and limestone, with associated stratiform barite (Christie, 1979; Cecile and Norford, 1979; Goodfellow and Jonasson, 1984). The uppermost Devonian and lowermost Mississippian black shale of the Exshaw Formation in southwest Alberta, Canada has several thin beds of phosphorite (Christie, 1979; Macqueen and Sandberg, 1970).

The presence of phosphates within the autochthonous miogeocline supports the original proximity of the allochthonous phosphate-bearing terranes to North America. The miogeoclinal sites that tie phosphate deposition to cratonic North America include, besides the Canada localities noted above, the Inyo Mountains (fig. 1; C. Stevens, personal commun., 1983) and central and eastern Nevada (Rogers and others, 1970; note that sites of Sandberg and Gutschick, 1980, in Utah are distinctly younger). The phosphogenic province spanned both the continental margin and a number of terranes.

Geologic and paleomagnetic studies are crucial to understanding the history of terrane motions, but the presence of phosphate may support particular scenarios. Controversy exists, for example, whether the Northern Sierra terrane was exotic to or close to North America or other Cordilleran terranes in the middle Paleozoic. The Northern



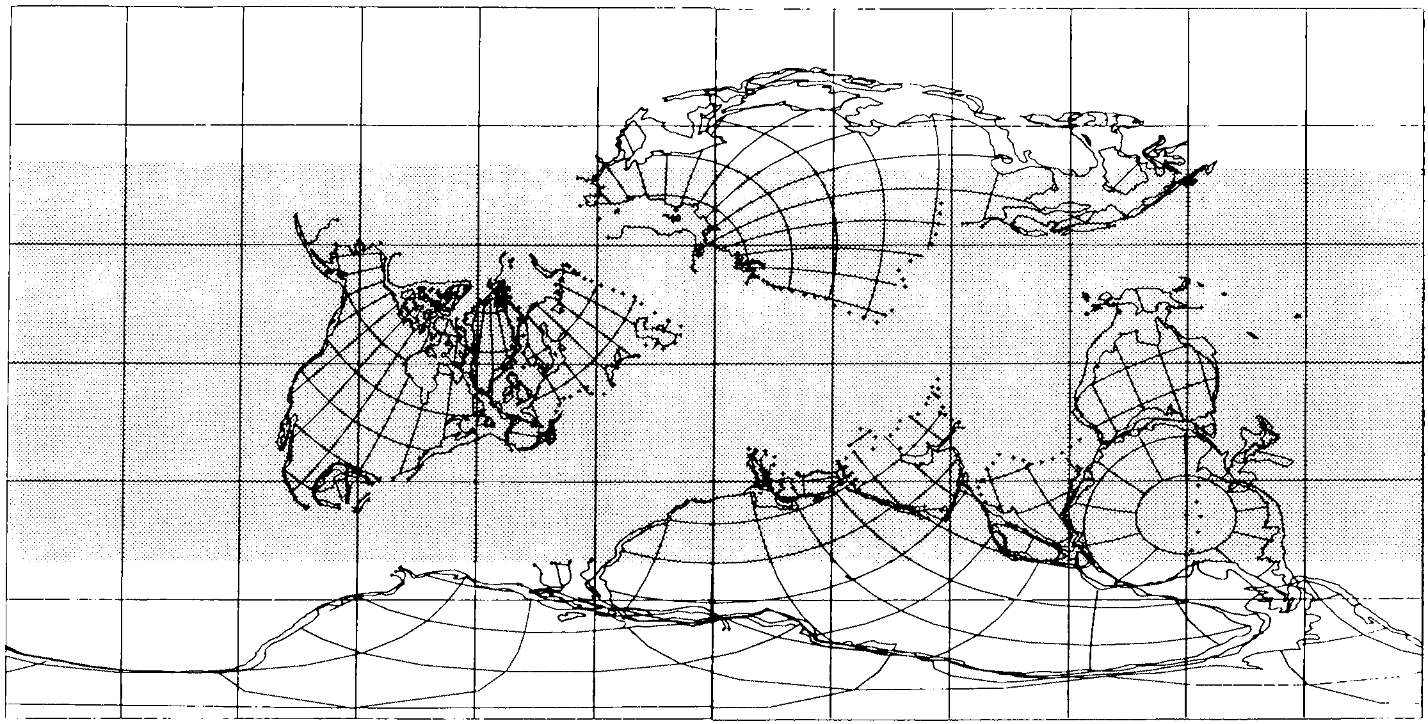


Fig. 6(A) Late Ordovician paleocontinental reconstruction of Smith, Hurley, and Briden (1981). Only a few continental margins face a large ocean within a 50° band around the equator; one of these is the Cordilleran margin of North America (Irving, 1979; Scotese, 1984; Miller and Kent, 1986). (B) Late Devonian paleocontinental reconstruction of Smith, Hurley, and Briden (1981).

Sierra terrane was part of the phosphogenic province over the span of time before and after middle Paleozoic deformation (Varga, 1982; deformation prior to the late Devonian, see Schweickert and Girty, 1981; Varga and Moores, 1981; Girty and others, 1984; Schweickert and others, 1984; Hanson and Schweickert, 1986). We have argued that the phosphogenic province was most likely adjacent to western North America. The presence of phosphatic cherts supports, though it does not prove, previous arguments (Churkin, 1974; Burchfiel and Davis, 1975; Schweickert and Snyder, 1981; Varga and Moores, 1981; Varga, 1982; Girty and others, 1984; Schweickert and Lahren, 1987) that the lower to middle Paleozoic strata in the northern Sierra Nevada and adjacent roof pendants are correlative with or related to rocks of the Roberts Mountains allochthon in Nevada (contrast this with the views of Dickinson, 1977; Speed, 1979; Bond and Devay, 1980).

Stratigraphic links among phosphatic rocks of the Cordillera.—Sedimentary rocks, described in published reports, tie the terranes bearing lower to middle Paleozoic phosphate to one another and to a continental source of detritus.

Several horizons of pure quartz arenite are known in the Ordovician strata of the Cordilleran miogeocline (Stewart and Poole, 1974; Ross, 1977) and in terranes to the west. Ketner (1966; 1968) noted that the quartz arenite of the miogeocline is extremely pure, medium-grained, and has well-rounded, well-sorted, frosted grains. The thickest sheet of quartz arenite is as much as several hundred meters thick and occurs across the miogeocline from California to Idaho and in the southern Canadian Cordillera. Similar quartz arenite of approximately correlative age occurs in beds from a few meters to 100 or more meters thick in Ordovician rocks of the Roberts Mountains allochthon (Ketner, 1966; Churkin and Kay, 1967; Stewart, 1980; Miller and Larue, 1983). Ordovician cherts in the allochthon contain well-rounded quartz grains (Suczek, 1981). The similarity in age and nature of the quartz arenites in the miogeocline and in the Roberts Mountains allochthon has been regarded as evidence that they are part of the same lithosome and that the Roberts Mountains allochthon is not exotic with regard to North America (Churkin, 1974; Miller and Larue, 1983; but see Ketner, 1966). Age equivalent Ordovician quartz arenite is present in the Inyo Mountains section (Stevens, 1986), in the Saddlebag Lake roof pendant (possibly age equivalent, Schweickert and Lahren, 1984; 1987), in lower Paleozoic units in the El Paso Mountains (Carr, Poole, and Christiansen, 1984), and at San Marcos (Lothringer, 1984).

Several Paleozoic sandstones of the Cordillera have yielded Precambrian U/Pb provenance ages from detrital zircons. Girty and Wardlaw (1985) reported early Proterozoic ages from detrital zircons in the Ordovician Valmy Formation and Silurian Elder Sandstone in the Roberts Mountains allochthon, and in subarkosic sandstones and quartz wacke in the Shoo Fly Complex. Girty and Wardlaw (1985) argued that these data, together with the petrography of the sandstones (see also

Bond and Devay, 1980), are consistent with a common continental provenance (which could have been North America) for the sandstones containing old detrital zircons.

Ordovician conodonts recovered from olistoliths in the upper part of the Shoo Fly Complex are indicative of North American Midcontinent faunal zones (Hannah and Moores, 1986). This conodont assemblage is consistent with an origin proximal to North America (Hannah and Moores, 1986).

SUMMARY AND PALEOGEOGRAPHIC MODEL

Phosphate-bearing black shale and chert of early to middle Paleozoic age are more widespread in the North American Cordillera than previously recognized. Deposits are known from the Yukon through southwestern United States and from Sonora and west Texas. The phosphate is present as apatite and contains REE having a continental rather than a deep-sea signature. Sedimentary rocks associated with phosphatic shale and chert probably were deposited in proximity to some continental area and in a water depth somewhat shallower than abyssal ocean floor. The lithologic associations and REE chemistry of the phosphate itself suggest the rocks were deposited in a continental-margin zone of upwelling.

We infer that a west-coast-type phosphogenic province spanned most of the western margin of North America at several times during the early to middle Paleozoic. We speculate that it was perhaps hundreds of kilometers in width and extended the length of the western North American margin (about 4000 km). Deposition of phosphatic chert and shale within this province was confined to areas lacking significant detrital input and experiencing upwelling of nutrient-rich deep water. This model does not specify geography, bathymetry, or tectonics of the province, and the presence of phosphate in a Cordilleran terrane only suggests that it was part of the Cordilleran province at the time the phosphate formed. A prediction of this model is that deeper-water sedimentary rocks deposited on abyssal oceanic basement and far-travelled terranes exotic to North America (other than those from other phosphogenic provinces) should lack organic-rich phosphatic chert and shale.

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Appendix I

Computation of cerium anomaly

Rare earth element (REE) abundances may be normalized to average shale values (Piper, 1974) or chondritic values (Leedy values of Masuda, Nakamura, and Tanaka, 1973, divided by 1.2, C. Langmuir, personal commun., 1986). Normalized REE abundances are usually plotted on a logarithmic scale against atomic number on a linear scale (Masuda-Coryell plot in the case of chondritic normalization, Shimizu and Masuda, 1977). Cerium abundance commonly deviates from the smooth pattern of the other REE. Usually its normalized value is compared to an expected value on a linear join between La and Nd (or sometimes La and Pr, as in de Baar, Bacon, and Brewer, 1983). Methods of calculation of the expected Ce value and the Ce anomaly vary in the literature. Normalization with respect to shale and a La-Nd join is used in the examples given below; the original sources vary in the method of normalization used.

One method of calculation (de Baar, Bacon, and Brewer, 1985) uses a *linear-linear* join between La and Nd to compute the expected value of Ce, called Ce^{*}_{SN} (in the shale normalized example):

$$Ce^{*}_{SN} = (2 \cdot La_{SN} + Nd_{SN})/3 \qquad (A1)$$

and

$$Ce\ anomaly = Ce_{SN}/Ce^{*}_{SN}. \qquad (A2)$$

A second method of calculation, given by Wright, Seymour, and Shaw (1984), computes Ce^{*}_{SN} as above in eq (A1), and

$$Ce\ anomaly = \log\,(Ce_{SN}/Ce^{*}_{SN}) \qquad (A3)$$

which is simply the log of the Ce anomaly in the first method. Anomalies described as negative or positive are given corresponding signs in this method.

A third method (used by Shimizu and Masuda, 1977) is that described in eqs (1) to (3) of this paper. The expected value of Ce is computed on a *log-linear* join. The calculated numerical anomaly values are easily compared with the commonly utilized log-linear REE plot.

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