Title

Provenance Controls on the Nd-Sr-O Isotopic Composition of Sandstones: Example from Late Mosozoic Great Valley Forearc Basin, California

Permalink

Authors

Linn, A.M.
DePaolo, D.J.

Publication Date

1992-04-01

LBL-33628
UC-401

Lawrence Berkeley Laboratory
UNIVERSITY OF CALIFORNIA

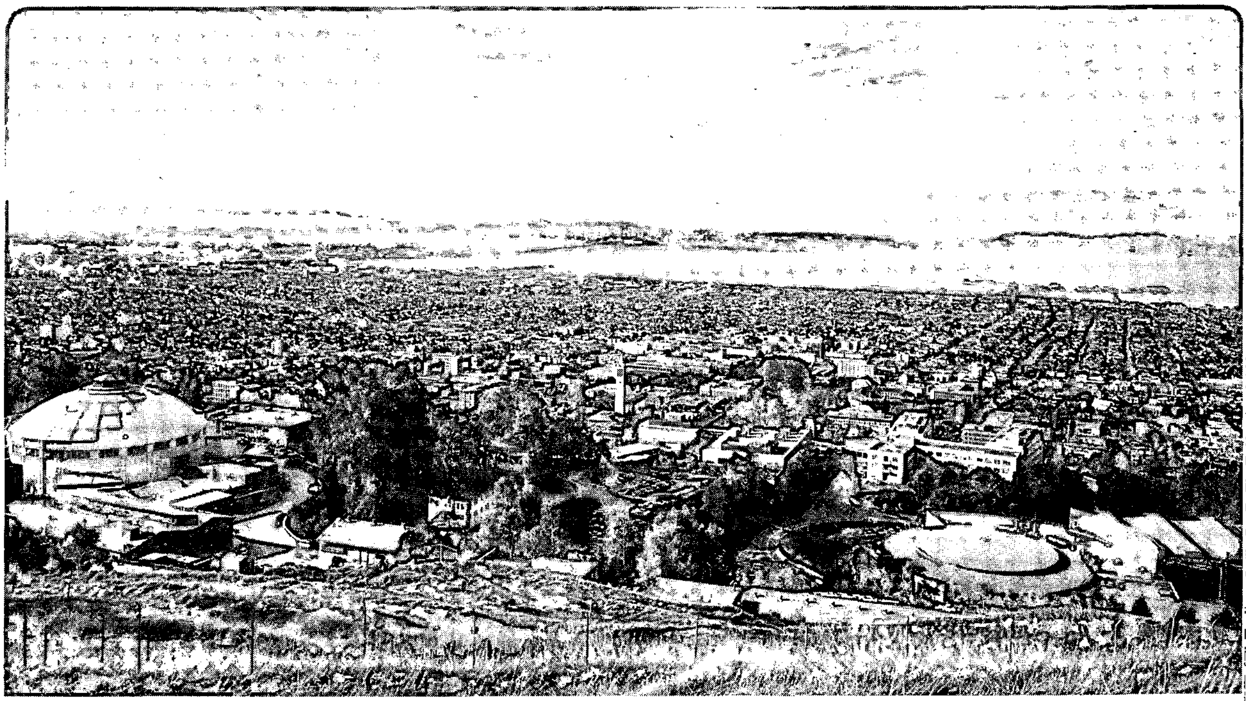
EARTH SCIENCES DIVISION

To be published as a chapter in *Processes Controlling the Composition of Clastic Sediments* (GSA Special Paper 284), Geological Society of America, Boulder, Colorado

Provenance Controls on the Nd-Sr-O Isotopic Composition of Sandstones: Example from Late Mesozoic Great Valley Forearc Basin, California

A.M. Linn and D.J. DePaolo

April 1992



Prepared for the U.S. Department of Energy under Contract Number DE-AC03-76SF00098

LBL-33628

Provenance Controls on the Nd-Sr-O Isotopic Composition of Sandstones: Example from Late Mesozoic Great Valley Forearc Basin, California

A. M. Linn and D. J. DePaolo

Department of Geology and Geophysics
University of California

and

Earth Sciences Division
Lawrence Berkeley Laboratory
University of California
Berkeley, California 94720

April 1992

This work was supported by the Director, Office of Energy Research, Office of Basic Energy Sciences, under U.S. Department of Energy Contract No. DE-AC03-76SF00098.

ABSTRACT

Measurements of Nd, Sr and O isotopes were made on whole-rock Great Valley sandstones in order to determine the extent to which isotopes preserve the composition and weathering history of tectonically active sources. Previous work has shown that the Nd isotopic composition of sedimentary rocks provides an accurate record of the source and can be used to evaluate the effect of weathering and diagenesis on Sr-O isotope systematics. Variations in the whole-rock Nd, Sr and O isotopic compositions are similar within and between petrofacies; this suggests that the isotopic composition is controlled by provenance. The Nd-Sr isotopic compositions are sensitive to the large west-east variation in crust formation ages of the igneous and metasedimentary components. The sandstones decrease from +7 to -5 in ε_{Nd} and increase from 0.7045 to 0.7073 in $^{87}Sr/^{86}Sr$ with decreasing stratigraphic age. These values nearly encompass the range observed in the plutonic rocks and suggest that the Nd-Sr isotopic composition of the arc is preserved in the sandstones. In contrast, $\delta^{18}O$ decreases from +19 to +9 with decreasing stratigraphic age; these values are significantly higher than $\delta^{18}O$ values in the volcanic/plutonic arc source. Isotopic and petrographic variations are correlated; with increasing proportions of sedimentary and metasedimentary lithic fragments, values of ε_{Nd} decrease, and values of $^{87}Sr/^{86}Sr$ and $\delta^{18}O$ increase. The sedimentary and metasedimentary components decrease the whole-rock value by 2 to 4 ε_{Nd} units and by 1 to 10‰, depending on the fraction and lithology of the recycled component. The metasedimentary component has a significantly lower Sr concentration than the igneous component and therefore has a smaller effect on the whole-rock $^{87}Sr/^{86}Sr$ value. Subtraction of the sedimentary-metasedimentary component from the whole-rock sandstone compositions yields igneous isotopic compositions which are directly comparable to the composition of the arc source. Calculated igneous compositions of the sandstones decrease in ε_{Nd} from +7 to -5 with decreasing stratigraphic age and indicate that sediment sources were located at the eastward-migrating volcanic

front. However, the calculated igneous components are +1 to +7 ‰ higher in $\delta^{18}O$ than the igneous source. The high $\delta^{18}O$ values are a measure of alteration of the source; increasing alteration in $\delta^{18}O$ is correlated with increasing values of CIA. This suggests that high mechanical erosion rates of tectonically active sources does not preclude significant chemical weathering.

INTRODUCTION

In recent years, the isotopic and geochemical analysis of sedimentary rocks has provided new insights into the evolution of crustal, tectonic and climatic processes through geologic time (e.g., Taylor and McLennan, 1985; Bowan, 1991). Nd isotopes are of particular interest in provenance studies because the Nd isotopic composition of the source rocks is preserved throughout the sedimentary cycle. Measurements of the Nd isotopic composition of siliciclastic rocks have been made to locate sediment source areas (e.g., Frost and Winston, 1987; Goldstein and Jacobsen, 1988; Nelson and DePaolo, 1988; Chen et al., 1990; McLennan et al., 1990; Linn et al., 1991, 1992), to reconstruct tectonic terranes (Miller and O'Nions, 1984; Clarke and Halliday, 1985), and to constrain models of crustal growth and sediment recycling (e.g., Hamilton et al., 1983; O'Nions et al., 1983; Allegre and Rousseau, 1985; Michard et al., 1985). Sr and O isotopes can also identify sedimentary sources (e.g, Clayton et al., 1978; Spiers et al., 1984; Yeh and Eslinger, 1986; Goldstein and Jacobsen, 1988; Nelson and DePaolo, 1988; Chen et al., 1990), but are more susceptible to modification by weathering, transport and diagenetic processes. Nevertheless, Sr and O isotopes provide additional ways to locate source areas and their analysis in association with Nd isotopes or other provenance indicators offers a means to test and refine provenance interpretations (see Heller et al., 1985 for an excellent example).

In this paper, we investigate the relation between provenance and the Nd-Sr-O isotopic composition of Mesozoic Great Valley forearc sandstones of California. Great Valley sedimentary rocks are ideal for this purpose because the large west-east age and isotopic gradients in the Sierra Nevada arc source make it possible to do high-resolution provenance studies. Moreover, petrographic analysis indicates that a wide variety of lithologies are present in the sandstones (e.g., Dickinson and Rich, 1972; Ojakangas, 1968; Ingersoll, 1983); this provides constraints on the isotopic interpretations and allows us to test the relation between mineralogy and isotopic composition. Finally, the high erosion rates of active continental margin arcs (Dumitru, 1988; Linn et al., 1992) and the subsequent immaturity of the sandstones enhances the probability that the sediments will preserve the isotopic composition of the source. Linn et al. (1991, 1992) showed that the Nd isotopic and geochemical compositions of the Sierra Nevada arc are preserved in Great Valley sedimentary rocks and could be used to pinpoint source areas. In contrast, the Sr isotopic composition of the Upper Jurassic and Lower Cretaceous sandstones was modified by weathering, transport or diagenetic processes and was less useful in determining provenance (Linn et al., 1992). The ability to use Nd isotopes as provenance indicators allows us to evaluate the extent to which the oxygen isotopic composition of arc sources is retained in sedimentary rocks. At the same time, the susceptibility of oxygen isotopes to weathering and diagenetic processes allows us to place constraints on the sedimentary processes which can modify the chemical and isotopic composition of siliciclastic rocks.

ISOTOPIC COMPOSITIONS OF SILICLASTIC ROCKS

Nd isotopes

Sm and Nd are rare earth elements which have similar chemical properties, except that Nd has a slightly larger ionic radius. The major partitioning of Sm from Nd occurs during extraction of continental crust from the mantle (DePaolo and Wasserburg, 1976). Consequently the Nd isotopic composition of igneous sources varies between mantle and continental rocks and between crustal provinces of different ages. Isotopic variations in the source are most likely to be preserved in first-cycle sedimentary rocks; subsequent reworking of the sediments will yield an average age (denoted the crustal residence age) or isotopic composition of the source. Because the Sm/Nd value is similar in most igneous minerals and is preserved except under intense conditions of mechanical and chemical weathering of crustal rocks (e.g., Balashov et al., 1964; Fleet, 1984), the Nd isotopic composition of siliciclastic rocks is independent of grain size (Linn et al., 1992). Transport processes, however, may concentrate minerals or lithic fragments which differ in age and/or isotopic composition into different size fractions (e.g., Frost and Coombs, 1989; McLennan et al., 1989). Similarly, recrystallization of clay minerals or precipitation of cement during late-stage diagenesis may alter the Sm/Nd value of the rock (e.g., Balashov and Girin, 1969; Awwiller and Mack, 1991), although redistribution of the rare earth elements may be localized (Schieber, 1988). Finally, metamorphism can fractionate rare earth elements and form minerals which differ in Sm/Nd from the whole-rock (Humphris, 1984; Nelson and DePaolo, 1988). This effect is significant only if the age difference between crust formation and metamorphism is large (Nelson and DePaolo, 1988), or if the rare earth elements are concentrated in the metamorphic minerals, as is the case in quartzose sediments (Cullers et al., 1979; Frost and Winston, 1987).

Sr isotopes

Fine-grained sedimentary rocks commonly fall on linear arrays in $^{87}Sr/^{86}Sr$ - $^{87}Rb/^{86}Sr$ space; these "isochrons" yield the mineral, depositional or diagenetic age (e.g., Moorbath, 1969; Peterman et al., 1981; Heller et al., 1985; Faure, 1986). Isochron ages are valid only to the extent that the minerals or sedimentary rocks behave as a closed system with regard to Rb and Sr (Faure, 1986). However, the chemical properties of Rb, an alkali element, differ from Sr, an alkaline earth element. Sr is mobile relative to Rb, and chemical weathering, seawater exchange and diagenesis can remove Sr from plagioclase and partition it into calcium carbonate cement. Rb extracted from micas and potassium feldspar during weathering is partitioned into clay minerals (e.g., Clauer, 1982), but tends to remain relatively inert during diagenesis (Perry and Turekian, 1974). Transport processes enhance the parent-daughter fractionation by mechanical sorting of fine-grained clays and micas which have high Rb/Sr values from coarser framework minerals which have lower Rb/Sr values. The magnitude of these fractionations can be difficult to quantify. Nevertheless, analysis of immature, relatively unaltered sandstones or mineral separates can yield geologically meaningful Sr isotopic compositions and provenance ages (e.g., Heller et al., 1985; Faure, 1986; Linn et al., 1992).

O isotopes

In contrast with the Nd-Sr isotope systems, the oxygen isotopic composition is primarily a function of the temperature, rather than the age, of rock formation. Igneous rocks typically crystallize at high temperatures and have relatively uniform $\delta^{18}O$ values of +6 to +11 (Hoefs, 1987). Sedimentary rocks are high in $\delta^{18}O$, ranging from about +10 to +40‰ (Hoefs, 1987), because they contain detrital or authigenic minerals which formed at low temperatures. For example, clay minerals formed during chemical weathering are 10 to 30‰ higher than the parent rock, depending on the temperature and isotopic composition of the meteoric waters (Savin and Epstein, 1970a, b). Rock fragments and coarser-grained framework minerals are resistant to isotopic exchange during weathering (e.g., Lawrence and Taylor, 1972; Clayton et al., 1978), but may begin to equilibrate with formation fluids during diagenesis at temperatures exceeding 200 to 300 °C (e.g., O'Neil and Kharaka, 1976; Yeh and Savin, 1977; Land and Milliken, 1981; Longstaffe, 1983; Eslinger and Yeh, 1986). The fine size fraction is more susceptible to modification by burial metamorphism, and may begin to equilibrate with pore fluids upon burial (Yeh and Savin, 1977). Whether diagenetic reactions increase or decrease the $\delta^{18}O$ value of siliciclastic rocks depends on the origin and evolution of the formation fluids during the burial history of the sedimentary basin. For example, alteration of pelagic sediments by fluids derived from an underlying basalt may decrease the whole-rock $\delta^{18}O$ value (Smith, 1991), whereas recharge by meteoric fluids during uplift may increase the $\delta^{18}O$ value (e.g., Ayalon and Longstaffe, 1987; Tilley and Longstaffe, 1989). Because of the sensitivity of oxygen isotopes to alteration effects, only immature sandstones or quartz separates are likely to retain significant provenance information (Savin and Epstein, 1970a; Clayton et al., 1978; Eslinger and Yeh, 1986).

GEOLOGY OF SIERRA NEVADA ARC - GREAT VALLEY FOREARC

The Great Valley Group consists of nearly 15 km of Late Jurassic and Cretaceous sandstone, shale and conglomerate which is exposed in a homocline along the western margin of the central valley of California (Fig. 1) (Dickinson and Rich, 1972). The forearc basin tectonically overlies the Franciscan accretionary complex to the west, extends below the surficial sediments of the Central Valley, and nonconformably overlies the eroded western flank of the Sierra Nevada arc and southern margin of the Klamath Mountains (Ingersoll, 1979). The Sierra Nevada arc is compositionally zoned with large west-east gradients in age, mineralogy, and geochemical and isotopic composition. Plutonic rocks in the western part of the arc are relatively mafic and are high in concentrations of FeO, MgO and CaO, whereas plutonic rocks along the eastern margin are more felsic and are high in concentrations of K_2O, Rb, U and Th (Dodge et al., 1982; Ague and Brimhall, 1988). From west to east, plutonic rocks decrease in ε_{Nd} from +7 to -6 (Fig. 1a) (Farmer and DePaolo, 1983; Linn et al., 1991) and increase in $^{87}Sr/^{86}Sr$ from 0.704 to 0.708 (Kistler and Peterman, 1973). The isotopic gradients are inferred to have been similar in the coeval volcanic cover (Linn et al., 1992). The gradients are largest in the southern part of the arc due to involvement of mantle-derived magma with Proterozoic continental shelf sediments (DePaolo, 1981) which are estimated to be -10 to -14 in ε_{Nd} and about 0.717 in $^{87}Sr/^{86}Sr$ (Linn et al., 1992). West-east crustal assimilation patterns in $\delta^{18}O$ were apparently overprinted by secondary processes, resulting in oxygen isotopic compositions which vary between +7 and +10‰ in the plutonic rocks (Fig. 1b) (Masi et al., 1981). Volcanic sources which supplied sediment to the early forearc basin were likely +6 to +7 in $\delta^{18}O$, similar to average island arcs (Hoefs, 1987).

The Sierra Nevada arc - Great Valley forearc system formed in Late Jurassic time following the Nevadan orogeny (Schweickert and Cowan, 1975; Saleeby, 1981). Late Jurassic and Early Cretaceous sediment derived from the Sierra Nevada arc and Klamath Mountains was deposited as submarine fans in a forearc slope setting which evolved into a deep forearc basin by Early Cretaceous time (Ingersoll, 1982). Throughout Cretaceous time, the forearc basin widened as a result of prograde westward accretion of the Franciscan complex and the eastward migration of arc magmatism (Dickinson 1973; Ingersoll, 1979). As inactive parts of the Sierra Nevada arc and Klamath mountains cooled and subsided below sea level, the subaerial arc and its basement supplied an increasing proportion of detritus to the forearc basin and became the only sediment source in Late Cretaceous time (Ingersoll, 1979). Subduction ended in Neogene time with the development of a transform margin along the California coast (Atwater 1970), and the Great Valley Group was tilted and uplifted above sea level (Ingersoll, 1979).

Variations in the relative mineral and lithic fragment abundances subdivide the Great Valley Group and record the evolution of sediment sources (e.g, Ojakangas, 1968; Dickinson and Rich, 1972; Ingersoll, 1983). The most comprehensive analysis of the sandstone composition is reported in Ingersoll (1983) and is summarized in Table 1. The lower part of the stratigraphic section is exposed in the Sacramento Valley and is divided into the Stony Creek, Lodoga and Boxer petrofacies. Upper Jurassic and lowermost Cretaceous Stony Creek sandstones are either composed of volcanic lithic fragments and plagioclase feldspar, or are high in chert and shale-argillite lithic fragments. Lower Cretaceous Lodoga sandstones are high in metamorphic lithic fragments, and middle Cretaceous Boxer sandstones are high in volcanic lithic fragments. The upper part of the section is exposed in the San Joaquin Valley and is divided into the Grabast, Los Gatos and Rumsey petrofacies. Middle Cretaceous Grabast and Los Gatos sandstones are high in metamorphic lithic fragments, and uppermost Cretaceous Rumsey sandstones are relatively quartzo-feldspathic with significant amounts of k-feldspar. The temporal change from lithic to quartzo-feldspathic sandstone has been interpreted as the unroofing sequence of the arc, and the relative increase in k-feldspar with decreasing stratigraphic age has been interpreted as indicating that source areas moved east during Cretaceous time (Ojakangas, 1968; Dickinson and Rich, 1972; Ingersoll, 1983).

PROVENANCE OF GREAT VALLEY GROUP

Isotopic Stratigraphy

Stratigraphic variations in the whole-rock Nd-Sr-O isotopic compositions of Great Valley strata are shown in Figure 2 and are reported in Table 2. Isotopic variations in the sandstones are large but somewhat systematic among the petrofacies, and are strikingly similar in all the isotope systems. Because the Nd-Sr isotopic systems are best correlated, we use them to illustrate the stratigraphic variations in the sandstones, then compare the overall trends to the oxygen isotopic composition. Stony Creek sandstones are divided into two groups; one with ε_{Nd} values between +1.0 and -1.5 and one with ε_{Nd} values between +7 and +4. Although sandstones that are low in ε_{Nd} are high in $^{87}Sr/^{86}Sr$, there are no groups with distinct Sr isotopic compositions within the petrofacies. Rather, $^{87}Sr/^{86}Sr$ decreases from 0.7063 to 0.7045 in Upper Jurassic to Lower Cretaceous strata. Lodoga sandstones decrease in ε_{Nd} from +7 at the base to about 0 in the middle, then increase to +3 at the top of the petrofacies. The $^{87}Sr/^{86}Sr$ value also reverses in isotopic composition with decreasing stratigraphic age, and increases from 0.7044 to 0.7062, then decreases to 0.7048. Boxer sandstones are highly variable and encompass nearly the entire range of Nd and Sr isotopic compositions observed in the Sacramento Valley sandstones. Boxer sandstones vary in ε_{Nd} from +5.4 to -1.6, and in $^{87}Sr/^{86}Sr$ from 0.7039 to 0.7066. Nd-Sr isotopic compositions in San Joaquin Valley petrofacies are less variable than in Sacramento Valley petrofacies. Grabast and Los Gatos sandstones decrease in ε_{Nd} from -0.7 to -4 and increase in $^{87}Sr/^{86}Sr$ from 0.7052 to 0.7067 with decreasing stratigraphic age. Rumsey sandstones are low in ε_{Nd} and range from -4 to -5, although the uppermost sandstone is higher in ε_{Nd}. The apparent decrease in $^{87}Sr/^{86}Sr$ from 0.7073 to 0.7060 with decreasing stratigraphic age is likely a sampling artifact; two sandstones which were low in ε_{Nd}, and presumably high in $^{87}Sr/^{86}Sr$, were calcite-cemented and were not measured for Sr isotopes.

Variations in the oxygen isotopic composition of Great Valley sandstones are similar to variations observed in the Nd-Sr isotopes (Fig. 2). In general, sandstones that are high in ε_{Nd} and low in $^{87}Sr/^{86}Sr$ are high in $\delta^{18}O$. This observation is somewhat surprising given the susceptibility of oxygen isotopes to modification by sedimentary processes. Stony Creek sandstones decrease in $\delta^{18}O$ from +19 to +13 with decreasing stratigraphic age, and like Sr isotopes, are not divided into distinct groups. There is a reversal in oxygen isotopic composition in Lodoga sandstones, which is analogous to the reversal in Nd-Sr isotopic compositions. With decreasing stratigraphic age, $\delta^{18}O$ increases from +12 to +17, then decreases to +14. Boxer sandstones are again the most variable petrofacies and range in $\delta^{18}O$ from +10 to +16. San Joaquin Valley sandstones are more variable in $\delta^{18}O$ than in ε_{Nd} and $^{87}Sr/^{86}Sr$, and the relation with Nd and Sr isotopes is less clear (Fig. 2). In contrast to the systematic stratigraphic changes in the Nd-Sr isotopic composition of Grabast and Los Gatos sandstones, $\delta^{18}O$ varies between +12 and +15.

Rumsey sandstones are low in $\delta^{18}O$ and range from +9 to +12, although the uppermost sandstone is again anomalous. The overall decrease in $\delta^{18}O$ with decreasing stratigraphic age in San Joaquin Valley sandstones is generally correlative with decreasing ε_{Nd} and increasing $^{87}Sr/^{86}Sr$. This trend is in the opposite sense as the Nd-Sr-O relation observed in Sacramento Valley sandstones.

The variation in isotopic composition of the petrofacies is correlative with the lithic fragment composition and thus reflects variations in provenance. Stony Creek sandstones which are high in volcanic lithic fragments are high in ε_{Nd} and low in $\delta^{18}O$, whereas sandstones which are high in chert and shale-argillite lithic fragments are low in ε_{Nd} and high in $\delta^{18}O$. There is also a correlation between high sedimentary lithic fragment content and high $^{87}Sr/^{86}Sr$, but the relation is less clear. Isotopic variations in the Lodoga, Grabast and Los Gatos petrofacies are related to the metasedimentary lithic fragment content. For any stratigraphic age, the sandstone with the highest fraction of metasedimentary lithic fragments is generally lowest in ε_{Nd} and highest in $^{87}Sr/^{86}Sr$ and $\delta^{18}O$. The sensitivity of whole-rock Nd-Sr isotopes to the metasedimentary component reflects the large difference in crust formation ages of the volcanic/plutonic and metasedimentary sources. Sedimentary and metasedimentary lithic fragments were partly derived from recycled Proterozoic continental crust, whereas the arc source has Nd-Sr model ages which are intermediate between Cretaceous mantle and the Proterozoic sediments which it intruded and assimilated (Fig. 3). The Sr model ages of the batholith are young relative to the Nd model ages. This indicates that arc and crustal sources lost Sr during chemical weathering and recycling and would lead to an underestimate of the age or fraction of the metasedimentary source. However, both Nd and Sr model ages of Great Valley sandstones are older than the depositional ages and indicate the presence of an older component (Table 2). Nd model ages (T_{DM}), which are defined as the age in which continental crust was extracted from a model depleted mantle, of the sandstones vary between 250 and 1190 Myr. This range is similar to the range in model ages of plutonic rocks in the Sierra Nevada batholith and suggests that the Nd isotopic compositions of the arc and its basement are preserved in the sedimentary rocks. In contrast, Sr model ages (denoted T_{UR} for crust extraction from a uniform mantle reservoir) of the sandstones range from 250 to 570 Myr. The Sr model ages of the sandstones are younger than both the Nd model ages of the sedimentary rocks and the Sr model ages of many of the plutonic rocks (Fig. 3). The old Sr model ages of some of the Sacramento Valley sandstones relative to model ages of the arc and the Upper Cretaceous sandstones is likely a result of Sr exchange by seawater interaction.

The whole-rock isotopic composition depends on the relative fraction, concentration and isotopic composition of the sedimentary, metasedimentary and igneous components. Assuming that the isotopic compositions and concentrations of the sedimentary and metasedimentary components are similar, the whole-rock Nd isotopic composition (ε_{WR}) of the sandstone is given by:

$$\varepsilon_{WR} = \frac{(f.sed)[Nd]_{sed}(\varepsilon_{Nd})_{sed} + (1-f.sed)[Nd]_{ig}(\varepsilon_{Nd})_{ig}}{(f.sed)[Nd]_{sed} + (1-f.sed)[Nd]_{ig}} ,$$

where *f.* is the fraction of metasedimentary lithic fragments (*sed*), and *ig* (=igneous) is the volcanic/plutonic arc component determined by petrography. Linn et al. (1992) estimated that the Great Valley metasedimentary component has a Nd concentration of 20 ppm, similar to the plutonic rocks, and is -8 to -14 in ε_{Nd}, which is significantly lower than the igneous rocks. Subtraction of this component yields volcanic/plutonic compositions that are 2 to 4 ε_{Nd} units higher than the whole-rock value in average Lodoga, Grabast and Los Gatos sandstones. This is a significant fraction of the overall isotopic variation and indicates that the Nd isotopic composition is a sensitive indicator of the pre-arc source. In contrast, the Sr concentration of the metasedimentary component is estimated to be about 100 ppm (Linn et al., 1992) which is only about 25% of the concentration of the igneous component due to Sr loss during sedimentary recycling. Consequently, the whole-rock $^{87}Sr/^{86}Sr$ value is controlled by the composition of the igneous source (Linn et al., 1992).

Location of Source Areas

In order to locate source areas, assumptions must be made regarding 1) the size and geometry of the drainage basin, 2) the erosion rate as a function of position within the drainage basin, and 3) the concentration and isotopic composition of sources contributing sediment to the forearc basin. Details of the first two parameters are described in Linn et al. (1992); the latter has been briefly discussed above. Subtraction of sedimentary and metasedimentary components from the whole-rock Nd-Sr isotopic compositions yields isotopic compositions that are representative of the arc source. The calculated igneous compositions are then compared with the isotopic gradient in the plutonic rocks, which varies only with geographic position, in order to locate the source area. Source locations are better constrained with Nd than Sr isotopes because the isotopic gradient in the plutonic rocks is larger and because Nd isotopes are less susceptible to modification by sedimentary processes.

With decreasing stratigraphic age, calculated igneous compositions of Great Valley sandstones decrease in ε_{Nd} from +7 to -5 and increase in $^{87}Sr/^{86}Sr$ from 0.704 to 0.707 (Linn et al., 1992). These values are the same as the range observed in the plutonic rocks of the Sierra Nevada arc and suggest that the Nd-Sr isotopic composition of the arc source is preserved in the sedimentary rocks. The decrease in ε_{Nd} and increase in $^{87}Sr/^{86}Sr$ with decreasing stratigraphic age is in the correct sense for the eastward-migration of source areas, in agreement with inferences drawn from sandstone petrography. Sediment source models show that San Joaquin Valley source areas moved from the central to the eastern part of the arc during Late Cretaceous time (Fig. 4a), and that Sacramento Valley source areas generally moved from the western to the central part of the arc during Early Cretaceous time (Fig. 4b). This conclusion does not depend on the values chosen for the metasedimentary component because the isotopic endmembers are Stony Creek and

Rumsey sandstones which are low in metasedimentary lithic fragment and therefore have whole-rock isotopic compositions that are directly comparable to the composition of the arc. The large range in ε_{Nd} suggests that source areas were localized at the migrating volcanic front; contributions of substantial amounts of sediment from previously active parts of the arc would yield a smaller range in ε_{Nd} because the values of the sources would be averaged. In general, the depositional age and calculated igneous isotopic composition of Great Valley sandstones are the same as the crystallization age and isotopic composition of the Sierra Nevada plutonic rocks. This indicates that the isotopic composition of the plutonic rocks is representative of the isotopic composition of the coeval volcanic cover and that erosion of volcanic sources was nearly contemporaneous with forearc deposition. Rapid erosion rates are also consistent with results from fission track dating (Dumitru, 1988), and with geomorphologic considerations which suggest that the volcanic front is a topographic high, which would quickly erode following cessation of magmatism (Linn et al., 1992).

Preservation of Source $\delta^{18}O$

The oxygen isotopic composition of sedimentary rocks is a function of the composition of the source rocks and the amount of alteration in $\delta^{18}O$ produced during weathering and diagenesis. In contrast to Nd-Sr isotopes, the whole-rock $\delta^{18}O$ values of the sandstones are significantly higher than the arc source and are more similar to Proterozoic shelf sources (Fig. 5). This suggests that oxygen isotopes are particularly sensitive to the presence of recycled detritus. The relative contributions of sedimentary, metasedimentary, volcanic and plutonic components to the whole-rock oxygen isotopic composition is determined in a manner analogous to the Nd mixing calculation above, except that the sources are:

1) Sedimentary rocks with $\delta^{18}O$ between +20 in average shale-argillite (Eslinger and Savin, 1973; Savin and Epstein, 1970b; Yeh and Savin, 1977), and +28 in chert (Blatt, 1987),

2) Metasedimentary rocks with $\delta^{18}O = +15$, similar to the value of average metamorphic rocks (Hoefs, 1987),

3) Western arc volcanic rocks with $\delta^{18}O = +6.5$, similar to average island arcs (Hoefs, 1987),

4) Central and eastern arc plutonic rocks with $\delta^{18}O$ between +7 and +8 (Masi et al., 1981).

The sedimentary and metasedimentary fractions are determined by petrography, and the $\delta^{18}O$ values of the volcanic and plutonic sources are determined by inference with Nd isotopes (see above). The correlation between $\delta^{18}O$ and the lithic fragment content is illustrated in Figure 6. Stony Creek sandstones which are high in chert and shale-argillite have the highest $\delta^{18}O$ values of the Great Valley Group, whereas Boxer and Rumsey sandstones which are high in volcanic or plutonic detritus are lowest in $\delta^{18}O$. Lodoga, Grabast and Los Gatos sandstones are high in metamorphic lithic fragments and have intermediate $\delta^{18}O$ values. The relation between petrography and $\delta^{18}O$ indicates that the oxygen isotopic composition of Great Valley sandstones is primarily controlled by provenance.

MODIFICATION OF Sr AND O ISOTOPES

Although broad compositional variations in the sources are recorded in the Sr-O isotopic compositions of the sandstones, the isotopic compositions are shifted toward higher values than would be predicted for the source rocks. For example, $^{87}Sr/^{86}Sr$ values of the igneous component in Upper Cretaceous sandstones correspond to $^{87}Sr/^{86}Sr$ values of the plutonic rocks, but Lower Cretaceous and Upper Jurassic sandstones are significantly higher in $^{87}Sr/^{86}Sr$ than plutonic rocks of the same age (Linn et al., 1992). With regard to oxygen isotopes, Rumsey sandstones are 1 to 4 permil higher in $\delta^{18}O$ than the estimated compositions of the sources, and Grabast, Boxer and Lodoga sandstones are up to 6 permil higher than their sources (Fig. 6). The volcaniclastic variety of the Stony Creek petrofacies ($\delta^{18}O \approx +15$) is nearly 9 permil higher than the source rocks. These observations suggest that the Sr-O isotopic compositions have been modified by sedimentary processes. In order to evaluate the extent to which the isotopes have been affected, we calculate the alteration factor Δ, which is the difference between the measured isotopic composition ($\delta^{18}O_{sand}$) and the composition calculated for mixtures of the sedimentary, metasedimentary and igneous components ($\delta^{18}O_{source}$), as determined by sandstone petrography. The oxygen isotopic values of the components are given above. The Sr isotopic composition of the igneous sources ($^{87}Sr/^{86}Sr_{calc.}$) are determined by analogy with Nd isotopes. Figure 7 shows that alteration of $\delta^{18}O$ varies between +1 and +5 in Upper Cretaceous sandstones, and varies between +3 and +8 in Lower Cretaceous and Upper Jurassic sandstones. The average shift in $\delta^{18}O$ of the sandstones is $\Delta = +4‰$. The Sr alteration factor behaves similarly; it increases from $\Delta = 0$ to 0.001 at the top of the section, to $\Delta = 0.002$ at the base of the section (Fig. 7). The alteration factors indicate that typically 30 to 40% of the Sr and O in the sandstones is secondary.

Modification of both Sr and O isotopes increases with increasing stratigraphic age. However, it is difficult to determine when the alteration occurred because the sandstones which were the most deeply buried and diagenetically modified were derived from mafic sources which are more susceptible to chemical weathering. The chemical index of alteration (CIA) provides a means of evaluating the effect of weathering on $\delta^{18}O$ and $^{87}Sr/^{86}Sr$. The CIA measures the degradation of feldspar to clay minerals and is determined from the abundances of K_2O, CaO and Na_2O relative to Al_2O_3 (Nesbitt and Young, 1982; Taylor and McLennan, 1985). CIA varies from 50 in unweathered continental crust to 100 in clay minerals, and is generally less than 65 in Great Valley sandstones (Table 2). This value is consistent with the high mechanical erosion rates observed in tectonically active settings (e.g., Blatt et al., 1980) and with the CIA in modern continental margin arc sand (McLennan et al., 1990). CIA does not vary with stratigraphic age in the samples we analyzed; this suggests that diagenetic processes did not significantly modify the major element abundances. Figure 8 shows that increasing CIA is correlated with increasing $\delta^{18}O$ and with increasing alteration. The range in $\delta^{18}O$ is large compared with the range in CIA and suggests that the oxygen isotopic composition is a more sensitive indicator of chemical weathering than even mobile major elements. The Δ values indicate that the sandstones contain about 35% weathering products, whereas the CIA values are consistent with only about 15% alteration. However, the CIA measures only feldspar alteration, and may be expected to indicate lower amounts of alteration than whole-rock oxygen isotopes. The $\delta^{18}O$ data suggest that high erosion rates of tectonically active sources do not preclude chemical weathering.

CIA is not correlated with the Sr alteration factor; this suggests that weathering did not modify the $^{87}Sr/^{86}Sr$ values of the sandstones. The most likely way to increase the $^{87}Sr/^{86}Sr$ value, and therefore Δ, is to exchange Sr in plagioclase with Sr in seawater during deposition and/or submarine eruption of the earliest volcanic sources. $^{87}Sr/^{86}Sr$ values in seawater ranged from 0.7070 to 0.7075 during Early Cretaceous time (Burke et al., 1982), and were significantly higher than $^{87}Sr/^{86}Sr$ values of the arc.

Because alteration increases with increasing stratigraphic age, and therefore with burial depth, it is necessary to evaluate the effect of burial metamorphism on $\delta^{18}O$. We consider 2 endmember models, which are illustrated in Figure 9. If the stratigraphic section behaves like an open system, an infinite reservoir of seawater will modify the isotopic composition of the sandstones toward higher $\delta^{18}O$ values at low temperatures and toward lower $\delta^{18}O$ values at high temperatures. The magnitude of the displacement depends on the fraction and mineralogy of the clay, and on the temperature gradient, because the clay-water fractionation decreases with increasing temperature. Great Valley sandstones which contain 30% authigenic chlorite should decrease from about +11 to +7‰ with increasing depth of burial. This is in the opposite sense as the trend observed in the sandstones (boxes), which *increase* in $\delta^{18}O$ with increasing depth of burial (Fig. 9). If the section behaves like a closed system, water/rock ratios are low and the isotopic compositions of the pore fluids are modified. As a result, increasing burial metamorphism could not significantly alter the whole-rock values. This is likely the case for the Great Valley Group because the section is inferred to have been buried nearly 15 km (Dumitru, 1988) which would result in very low porosity and permeability.

SUMMARY AND CONCLUSIONS

The whole-rock isotopic composition of sedimentary rocks is a function of 1) the composition of the source rocks, and 2) shifts in the isotopic values produced by weathering, transport and diagenetic processes. The Nd isotopic composition is generally preserved throughout the sedimentary cycle, and can be used to determine the provenance, and to evaluate provenance interpretations of Sr and O isotopes which are more susceptible to weathering and diagenetic processes. Nd-Sr-O isotopic variations are similar in Great Valley sandstones and are correlated with the lithic fragment content. Metasedimentary sources decrease ε_{Nd} and increase $^{87}Sr/^{86}Sr$ in the sandstones. O isotopes are particularly sensitive to the sedimentary-metasedimentary source because recycled components contain clay minerals which form at low temperatures and which are consequently higher in $\delta^{18}O$ than the igneous source. This suggests that even the oxygen isotopic composition is mainly controlled by provenance. However, $^{87}Sr/^{86}Sr$ and $\delta^{18}O$ values are higher than would be expected for the sources determined by Nd isotopes, and are highest in those sandstones which were derived from relatively mafic volcanic sources. The high $^{87}Sr/^{86}Sr$ values are likely a result of Sr-exchange with seawater either in the volcanic edifice or during deposition. However, alteration in $\delta^{18}O$ is correlated with CIA; this suggests that the oxygen isotopic composition was modified by low-temperature weathering processes, rather than diagenesis. This indicates that chemical weathering is important in tectonically active orogenic belts, particularly in those composed of mafic volcanic rocks. Modification of oxygen isotopes may also have been enhanced by the climatic conditions in the near-equatorial position of the Sierra Nevada arc during Cretaceous time. However, the intense chemical weathering regime associated with low latitudes was apparently not sufficient to disturb the Nd-Sr systematics or the major element ratios.

ACKNOWLEDGMENTS

Acknowledgment is made to the Donors of the Petroleum Research Fund, Grants PRF 19452-AC2 and PRF 22585-AC2-C, administered by the American Chemical Society, for partial support of this research. This work was also supported by the director, Office of Basic Energy Sciences, Engineering and Geosciences Division of the U.S. Department of Energy under Contract No. DE-AC03-76SF00098.

REFERENCES

Ague, J.J., and Brimhall, G.H., 1988, Regional variations of major chemical constituents across the central Sierra Nevada batholith: Geological Society of America Bulletin, v. 100, p. 891-911.

Allegre, C.J., and Rousseau, D., 1984, The growth of the continent through geological time studied by Nd isotope analysis of shales: Earth and Planetary Science Letters, v. 67, p. 19-34.

Atwater, T., 1970, Implications of plate tectonics for the Cenozoic tectonic evolution of western North America: Geological Society of America Bulletin, v. 81, p. 3513-3535.

Awwiller, D.N., and Mack, L.E., 1991, Diagenetic modification of Sm-Nd model ages in Tertiary sandstones and shales, Texas Gulf Coast: Geology, v. 19, p. 311-314.

Ayalon, A., and Longstaffe, F.J., 1988, Oxygen isotope studies of diagenesis and pore-water evolution in the western Canada sedimentary basin: evidence from the Upper Cretaceous Basal Belly River sandstone, Alberta: Journal of Sedimentary Petrology, v. 58, p. 489-505.

Balashov, Y.A., and Girin, Y.P., 1969, On the reserve of mobile rare earth elements in sedimentary rocks: Geochemistry International, v. 6, p. 649-659.

Balashov, Y.A., Ronov, A.B., Migdisov, A.A., and Turanskaya, N.V., 1964, The effect of climate and facies environment on the fractionation of the rare earths during sedimentation: Geochemistry International, v. 1, p. 951-969.

Barbarin, B., Dodge, F.C.W., Kistler, R.W., and Bateman, P.C., 1989, Mafic inclusions, aggregates, and dikes in granitoid rocks, central Sierra Nevada batholith, California - Analytical data: U.S. Geological Survey Bulletin 1899, 28 p.

Blatt, H., 1987, Oxygen isotopes and the origin of quartz: Journal of Sedimentary Petrology, v. 57, p. 373-377.

Blatt, H., Middleton, G., and Murray, R., 1980, Origin of sedimentary rocks: Englewood Cliffs, New Jersey, Prentice-Hall, p. 17-42.

Bowan, R., 1991, Isotopes and climates: New York, Elsevier, 483 p.

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

Chen, C.-H., Jahn, B.M., Lee, T., Chen, C.H., and Cornichet, J., 1990, Sm-Nd isotopic geochemistry of sediments from Taiwan and implications for the tectonic evolution of SE China: Chemical Geology, v. 88, p. 317-332.

Clarke, D.B., and Halliday, A.N., 1985, Sm/Nd isotopic investigation of the age and origin of the Meguma Zone metasedimentary rocks: Canadian Journal of Earth Sciences, v. 22, p. 102-107.

Clauer, N., 1982, The rubidium-strontium method applied to sediments: certitudes and uncertainties, in Odin, G.S., ed., Numerical Dating in Stratigraphy: New York, John Wiley & Sons, p. 245-276.

Clayton, R.N., Jackson, M.L., and Sridhar, K., 1978, Resistance of quartz silt to isotopic exchange under burial and intense weathering conditions: Geochimica et Cosmochimica Acta, v. 42, p. 1517-1522.

Cullers, R., Chaudhuri, S., Kilbane, N., and Koch, R., 1979, Rare-earths in size fractions and sedimentary rocks of Pennsylvanian-Permian age from the mid-continent of the USA: Geochimica et Cosmochimica Acta, v. 43, p. 1285-1301.

DePaolo, D.J., 1981, A neodymium and strontium isotopic study of the Mesozoic calc-alkalic granitic batholiths of the Sierra Nevada and Peninsular Ranges, California: Journal of Geophysical Research, v. 86, p. 10470-10488.

DePaolo, D.J., and Wasserburg, G.J., 1976, Nd isotopic variations and petrogenetic models: Geophysical Research Letters, v. 3, p. 249-252.

DePaolo, D.J., Linn, A.M., and Schubert, G., 1991, The continental crustal age distribution: Methods for determining mantle separation ages from Sm-Nd isotopic data and application to the southwestern United States: Journal of Geophysical Research, v. 96, p. 2071-2088.

Dickinson, W.R., 1973, Widths of modern arc-trench gaps proportional to past duration of igneous activity in associated magmatic arcs: Journal of Geophysical Research, v. 78, p. 3376-3389.

Dickinson, W.R., and Rich, E.I., 1972, Petrologic intervals and petrofacies in the Great Valley sequence, Sacramento Valley, California: Geological Society of America Bulletin, v. 83, p. 3007-3024.

Dodge, F.C.W., Millard, H.T., Jr., and Elsheimer, H.N., 1982, Compositional variations and abundances of selected elements in granitoid rocks and constituent minerals, central Sierra Nevada batholith, California: U.S. Geological Survey Professional Paper 1248, 24 p.

Dumitru, T.A., 1988, Subnormal geothermal gradients in the Great Valley forearc basin, California, during Franciscan subduction: A fission track study: Tectonics, v. 7, p. 1201-1221.

Eslinger, E.V., and Savin, S.M., 1973, Oxygen isotope geothermometry of the burial metamorphic rocks of the Precambrian Belt Supergroup, Glacier National Park, Montana: Geological Society of America Bulletin, v. 84, p. 2549-2560.

Eslinger, E.V., and Yeh, H.-W., 1986, Oxygen and hydrogen isotope geochemistry of Cretaceous bentonites and shales from the Disturbed Belt, Montana: Geochimica et Cosmochimica Acta, v. 50, p. 59-68.

Farmer, G.L., and DePaolo, D.J., 1983, Origin of Mesozoic and Tertiary granite in the western United States and implications for pre-Mesozoic crustal structure, 1. Nd and

Sr isotopic studies in the geocline of the northern Great Basin: Journal of Geophysical Research, v. 88, p. 3379-3401.

Fauer, G., 1986, Principles of isotope geology, 2nd edition: New York, John Wiley & Sons, 589 p.

Fleet, A.J., 1984, Aqueous and sedimentary geochemistry of the rare earth elements: *in*, Henderson, P., ed., Rare earth element geochemistry, New York, Elsevier, p. 343-373.

Frost, C.D. and Coombs, D.S., 1989, Nd isotope character of New Zealand sediments: implications for terrane concepts and crustal evolution: American Journal of Science, v. 289, p. 744-770.

Frost, C.D. and Winston, D., 1987, Nd isotope systematics of coarse- and fine-grained sediments: Examples from the middle Proterozoic Belt-Purcell Supergroup: Journal of Geology, v. 95, p. 309-327.

Goldstein, S.L., and Jacobsen, S.B., 1988, Nd and Sr isotopic systematics of river water suspended material: Implications for crustal evolution: Earth and Planetary Science Letters,v. 87, p. 249-265.

Hamilton, P.J., O'Nions, R.K., Bridgwater, D., and Nutman, A., 1983, Sm-Nd studies of Archaean metasediments and metavolcanics from West Greenland and their implications for the Earth's early history: Earth and Planetary Science Letters, v. 62, p. 263-272.

Heller, P.L., Peterman, Z.E., O'Neil, J.R., and Shafiquillah, M., 1985, Isotopic provenance of sandstones from the Eocene Tyee Formation, Oregon Coast Range: Geological Society of America Bulletin, v. 96, p. 770-780.

Hoefs, J., 1987, Stable isotope geochemistry, 3rd edition: New York, Springer-Verlag, 241 p.

Humphris, S.E., 1984, The mobility of the rare earth elements in the crust: *in*, Henderson, P., ed., Rare earth element geochemistry, Amsterdam, Elsevier, p. 317-342.

Ingersoll, R.V., 1979, Evolution of the Late Cretaceous forearc basin, northern and central California: Geological Society of America Bulletin, v. 90, p. 813-826.

Ingersoll, R.V., 1982, Initiation and evolution of the Great Valley forearc basin of northern and central California, U.S.A.: Geological Society of London Special Publication 10, p. 459-467.

Ingersoll, R.V., 1983, Petrofacies and provenance of Late Mesozoic forearc basin, northern and central California: American Association of Petroleum Geologists Bulletin, v. 67, p. 1125-1142.

Kistler, R.W., and Peterman, Z.E., 1973, Variations in Sr, Rb, K, Na, and initial Sr^{87}/Sr^{86} in Mesozoic granitic rocks and intruded wall rocks in central California: Geological Society of America Bulletin, v. 84, p. 3489-3512.

Kistler, R.W., Chappell, B.W., Peck, D.L., and Bateman, P.C., 1986, Isotopic variation in the Tuolumne intrusive suite, central Sierra Nevada, California: Contributions to Mineralogy and Petrology, v. 94, p. 205-220.

Land, L.S., and Milliken, K.L., 1981, Feldspar diagenesis in the Frio Formation, Brazoria County, Texas Gulf Coast: Geology, v. 9, p. 314-318.

Lawrence, J.R., and Taylor, Jr., H.P., 1972, Hydrogen and oxygen isotope systematics in weathering profiles: Geochimica et Cosmochimica Acta, v. 36, p. 1377-1393.

Linn, A.M., DePaolo, D.J., and Ingersoll, R.V., 1991, Nd-Sr isotopic provenance analysis of Upper Cretaceous Great Valley fore-arc sandstones: Geology, v. 19, p. 803-806.

Linn, A.M., DePaolo, D.J., and Ingersoll, R.V., 1992, Nd-Sr isotopic, geochemical and petrographic stratigraphy and paleotectonic analysis: Mesozoic Great Valley forearc sedimentary rocks of California: Geological Society of America Bulletin, in press.

Longstaffe, F.J., 1983, Diagenesis 4. Stable isotope studies of diagenesis in clastic rocks: Geoscience Canada, v. 10, p. 43-58.

Masi, U., O'Neil, J.R., and Kistler, R.W., 1981, Stable isotope systematics in Mesozoic granites of central and northern California and southwestern Oregon: Contributions to Mineralogy and Petrology, v. 76, p. 116-126.

McLennan, S.M., McCulloch, M.T., Taylor, S.R., and Maynard, J.B., 1989, Effects of sedimentary sorting on neodymium isotopes in deep-sea turbidites: Nature, v. 337, p. 547-549.

McLennan, S.M., Taylor, S.R., McCulloch, M.T., and Maynard, J.B., 1990, Geochemical and Nd-Sr isotopic composition of deep-sea turbidites: crustal evolution and plate tectonic associations: Geochimica et Cosmochimica Acta, v. 54, p. 2015-2050.

Michard, A., Gurriet, P., Soudant, M., and Albarede, F., 1985, Nd isotopes in French Phanerozoic shales: external vs. internal aspects of crustal evolution: Geochimica et Cosmochimica Acta, v. 49, p. 601-610.

Miller, R.G., and O'Nions, R.K., 1984, The provenance and crustal residence ages of British sediments in relation to paleogeographic reconstructions: Earth and Planetary Science Letters, v. 68, p. 459-470.

Moorbath, S., 1969, Evidence for the age of deposition of the Torridonian sediments of northwest Scotland: Journal of Geology, v. 5, p. 154-170.

Nelson, B.K., and DePaolo, D.J., 1988, Comparison of isotopic and petrographic provenance indicators in sediments from Tertiary continental basins of New Mexico: Journal of Sedimentary Petrology, v. 58, p. 348-357.

Nesbitt, H.W., and Young, G.M., 1982, Early Proterozoic climates and plate motions inferred from major element chemistry of lutites: Nature, v. 299, p. 715-717.

Ojakangas, R.W., 1968, Cretaceous sedimentation, Sacramento Valley, California: Geological Society of America Bulletin, v. 79, p. 973-1008.

O'Neil, J.R., and Kharaka, Y.K., 1976, Hydrogen and oxygen isotope exchange reactions between clay minerals and water: Geochimica et Cosmochimica Acta, v. 40, p. 241-246.

O'Nions, R.K., Hamilton, P.J., and Hooker, P.J., 1983, A Nd isotopic investigation of sediments related to crustal development in the British Isles: Earth and Planetary Science Letters, v. 63, p. 229-240.

Peterman, Z.E., Coleman, R.G., and Bunker, C.M., 1981, Provenance of Eocene graywackes of the Flournoy Formation near Agness, Oregon - A geochemical approach: Geology, v. 9, p. 81-86.

Perry, E.A., and Turekian, K.K., 1974, The effect of diagenesis on the redistribution of strontium isotopes in shales: Geochimica et Cosmochimica Acta, v. 38, p. 929-935.

Saleeby, J., 1981, Ocean floor accretion and volcanoplutonic arc evolution of the Mesozoic Sierra Nevada, *in* Ernst, W.G., ed., The geotectonic development of California: Englewood Cliffs, New Jersey, Prentice-Hall, p. 132-181.

Savin, S.M., and Epstein, S., 1970a, The oxygen and hydrogen isotope geochemistry of clay minerals: Geochimica et Cosmochimica Acta, v. 34, p. 25-42.

Savin, S.M., and Epstein, S., 1970b, The oxygen isotopic compositions of coarse grained sedimentary rocks and minerals: Geochimica et Cosmochimica Acta, v. 34, p. 323-329.

Schieber, J., 1988, Redistribution of rare-earth elements during diagenesis of carbonate rocks from the mid-Proterozoic Newland Formation, Montana, U.S.A.: Chemical Geology, v. 69, p. 111-126.

Schweickert, R.A., and Cowan, D.S., 1975, Early Mesozoic tectonic evolution of the western Sierra Nevada, California: Geological Society of America Bulletin, v. 86, p. 1329-1336.

Smith, B.M., 1991, Geologic implications of the oxygen isotope profile of the Toa Baja drill hole, Puerto Rico: Geophysical Research Letters, v. 18, p. 549-552.

Spiers, G.A., Dudas, M.J., Muehlenbachs, K., and Turchenek, L.W., 1984, Mineralogy and oxygen isotope geochemistry of clays from surficial deposits in the Athabasca Tar Sands area: Canadian Journal of Earth Sciences, v. 21, p. 53-60.

Suchecki, R.K., and Land, L.S., 1983, Isotopic geochemistry of burial-metamorphosed volcanogenic sediments, Great Valley sequence, northern California: Geochimica et Cosmochimica Acta, v. 47, p. 1487-1499.

Taylor, S.R., and McLennan, S.M., 1985, The continental crust: Its composition and evolution: Oxford, Blackwell Scientific, 312 p.

Tilley, B.J., and Longstaffe, F.J., 1989, Diagenesis and isotopic evolution of porewaters in the Alberta Deep Basin: the Falher Member and Cadomin Formation: Geochimica et Cosmochimica Acta, v. 53, p. 2529-2546.

Yeh, H.-W., and Eslinger, E.V., 1986, Oxygen isotopes and the extent of diagenesis of clay minerals during sedimentation and burial in the sea: Clays and Clay Minerals, v. 34, p. 403-406.

Yeh, H.-W., and Savin, S.M., 1977, Mechanism of burial metamorphism of argillaceous sediments: 3. O-isotope evidence: Geological Society of America Bulletin, v. 88, p. 1321-1330.

FIGURE CAPTIONS

Fig. 1 a) Present configuration of Sierra Nevada arc - Great Valley forearc system showing source lithologies and sandstone sample localities in the Sacramento and San Joaquin valleys. Mesozoic and Paleozoic metasedimentary and metavolcanic rocks along the northwest margin of the Sierra Nevada batholith are the Foothills Metamorphic belt. Heavy lines are contours of $^{143}Nd/^{144}Nd$ of Sierra Nevada plutonic rocks, expressed as ε_{Nd}, which decrease with increasing crustal assimilation toward the Proterozoic continental margin (PC). From Linn et al. (1992). Fig. 1b) Oxygen isotopic composition, expressed as contours of $\delta^{18}O$, of Mesozoic plutonic rocks in the Sierra Nevada batholith and Klamath Mountains. Modified from Masi et al. (1981).

Fig. 2 Correspondence in Nd-Sr-O isotopic composition with stratigraphic age in whole-rock Great Valley sandstones. San Joaquin Valley sandstones are divided into Rumsey (open boxes), and Los Gatos - Grabast petrofacies (open circles), and Sacramento Valley sandstones are divided into Boxer (filled triangles), Lodoga (filled circles) and Stony Creek petrofacies (filled boxes). Time scale and stratigraphic relations after Ingersoll (1983) and Linn et al. (1992). Nd-Sr isotopic data from Linn et al. (1992).

Fig. 3 Comparison of Nd and Sr model ages of Great Valley sandstones and their sediment sources (data from Linn et al. [1992]). Diagonal line represents equal Nd and Sr model ages; model ages to the right of the line indicate Rb-loss and model ages to the left of the line indicate Sr-loss (e.g., Nelson and DePaolo, 1988). Model ages of Sierra Nevada arc source (SNB) and Proterozoic crust (PZ) calculated by methods of DePaolo et al. (1991) and Farmer and DePaolo

(1983), from Nd-Sr isotopic data reported in DePaolo (1981), Kistler et al. (1986) and Barbarin et al. (1989).

Fig. 4 a) Depositional age and calculated igneous isotopic composition of San Joaquin Valley sandstones (open symbols), compared with crystallization age and isotopic composition of plutonic rocks (boxes) in an idealized west-east transect across the southern part of the Sierra Nevada batholith . K_w, K_c and K_e are western, central and eastern Cretaceous belts, respectively, from Saleeby (1981). Fig. 6b) Calculated Nd isotopic composition of the igneous component in Sacramento Valley sandstones (filled symbols) compared with age and geographic position of locus of magmatism (boxes) in west-east transect across the northern part of the arc. J_w is western Jurassic belt from Saleeby (1981). Symbols are the same as in Fig. 2. From Linn et al. (1992).

Fig. 5 Sr-O isotopic systematics of Sierra Nevada (■) and Klamath Mountain (▲) plutonic rocks, modified from DePaolo (1981). Simple mixing of Cretaceous mantle magma with Proterozoic continental sediments produces the Sr-O isotopic compositions observed in most Sierra Nevada plutonic rocks. San Joaquin (□) and Sacramento Valley (D) sandstones (this study) are similar in $^{87}Sr/^{86}Sr$ to the plutonic rocks, but are higher in $\delta^{18}O$. Isotopic data for the plutonic rocks is from Masi et al. (1981).

Fig. 6 Correlation in whole-rock fraction of sedimentary and metasedimentary lithic fragments and $\delta^{18}O$. Fields encompass the petrofacies; Stony Creek sandstones are divided into volcaniclastic and high-chert varieties. Symbols are the same as in Fig. 2.

Fig. 7 Degree of alteration (Δ) in Sr and O isotopic composition as a function of stratigraphic age. Alteration in Sr isotopic composition is the difference between the calculated value of the igneous fraction in the sandstones ($^{87}Sr/^{86}Sr_{source}$) and the value of the plutonic rocks in source areas inferred from Nd isotopes ($^{87}Sr/^{86}Sr_{calc.}$). Alteration in O isotopic composition is the difference between the measured whole-rock value ($\delta^{18}O_{sand}$) and the value calculated for mixtures of sedimentary, metamorphic and igneous components ($\delta^{18}O_{source}$), as determined by sandstone petrography.

Fig. 8 Correlation in $\delta^{18}O$ and chemical index of alteration in Great Valley sandstones. (CIA) defined in molecular proportions as $[Al_2O_3/(Al_2O_3+CaO+Na_2O+K_2O)] * 100$ Shaded region indicates range of compositions observed in unweathered to moderately weathered rocks.

Fig. 9 Model to evaluate the effect of burial metamorphism on the whole-rock oxygen isotopic composition of sandstones. $\Delta_{clay-water} = \delta^{18}O_{clay} - \delta^{18}O_{seawater}$. In an open system, the arrows denote the magnitude and direction of shift in calculated $\delta^{18}O$ values resulting from replacement of 30% of the bulk sandstone by chlorite. In a closed system, the calculated isotopic compositions are not modified by burial metamorphism. Chlorite and illite curves are calculated from equations in Longstaffe (1983) and Eslinger and Savin (1973). Stratigraphic depth and geothermal gradient from Dumitru (1988). Magnitude of pore fluid enrichment in closed system from Suchecki and Land (1983).

TABLE 1. MEAN PETROGRAPIC PARAMETERS OF GREAT VALLEY GROUP

	QFL %			LmLvLs %					
Petrofacies	Q	F	L	Lm	Lv	Ls	P/F	Lv/L	Qp/Q
San Joaquin Valley									
Rumsey	40	36	24	31	60	9	0.55	0.60	0.03
Los Gatos	38	40	23	54	38	8	0.64	0.40	0.04
Grabast	30	30	41	57	37	6	0.68	0.37	0.09
Sacramento Valley									
Boxer	32	32	36	19	73	8	0.75	0.73	0.09
Lodoga	37	21	42	33	42	25	0.86	0.42	0.21
Stony Creek	25	22	53	12	60	28	0.92	0.60	0.37

Note: Modified from Ingersoll (1982), where Q = monocrystalline + polycrystalline (Qp) quartz; F = plagioclase (P) + potassium feldspar; L = metamorphic (Lm) + volcanic (Lv) + sedimentary (Ls) lithic fragments.

Table 2. ISOTOPIC AND PETROGRAPHIC COMPOSITION OF GREAT VALLEY SANDSTONES

Sample	Age (Myr)	Petrofacies	ε_{Nd}*	T_{DM}	$^{87}Sr/^{86}Sr$*	T_{UR}	$\delta^{18}O$	%Ls+Lsm	CIA
San Joaquin Valley									
87GV-9	73	Rumsey	-3.2	1.09	0.70597	0.36	15.7		56
87GV-8	75	Rumsey	-4.9						
87GV-19	77	Rumsey	-4.3	1.19	0.70671	0.26	9.8	6	53
87GV-2	80	Rumsey	-5.0						
87GV-18	80	Rumsey	-4.0	1.16	0.70881	0.29	9.5	12	52
87GV-16	81	Rumsey	-4.3	1.19	0.70726	0.29	12.3	13	56
87GV-21	84	Grabast	-1.8	0.98	0.70703	0.39	12.4	22	53
87GV-15	86	Grabast	-4.0	1.17	0.70671	0.37	14.8		58
87GV-13	88	Grabast	-3.8	1.16	0.70795	0.37	13.0		55
87GV-10	90	Grabast	-1.4	0.90	0.70532	0.27	14.3	27	60
87GV-33	92	Grabast	-2.3	1.03	0.70715	0.39	12.9	22	54
87GV-31	94	Grabast	-0.7	0.89	0.70628	0.32	12.5	23	54
87GV-27	97	Grabast	-1.6	0.97	0.70938	0.32	12.9	27	59
Sacramento Valley									
89GV-48	86	Boxer	2.9						
89GV-44	88	Boxer	-1.6	0.96	0.70655	0.38	13.6	13	56
89GV-42	90	Boxer	5.4	0.35	0.70424	0.50	10.0	13	49
89GV-40	94	Boxer	3.2	0.55	0.70563	0.41	12.4	13	53
89GV-52	95	Boxer	1.4	0.71	0.70746	0.25	14.6		60
89GV-36	96	Boxer	1.0	0.75	0.70753	0.25	16.1		61
89GV-54	97	Boxer	-0.5	0.89	0.70805	0.37	13.3		58
89GV-53	103	Lodoga	2.9	0.59	0.70568	0.35	14.3		67
89GV-55	108	Lodoga	0.6	0.79	0.70675	0.46	14.7	24	59
89GV-56	113	Lodoga	-0.2	0.86	0.70770	0.39	15.9	33	63
89GV-57	115	Lodoga	0.0	0.85	0.70753	0.40	16.9		61
89GV-59	117	Lodoga	4.5	0.46	0.70560	0.29	15.8	33	63
89GV-58	119	Lodoga	6.9	0.25	0.70491	0.57	13.8		59
89GV-61	121	Lodoga	4.3	0.48	0.70499	0.56	11.9	10	56
89GV-64	136	Stony Creek	5.2	0.41	0.70514	0.54	13.5		58
89GV-60	138	Stony Creek	3.9	0.53	0.70597	0.62	15.5	13	58
89GV-65	144	Stony Creek	0.6	0.82	0.70668	0.50			
89GV-63	147	Stony Creek	1.1	0.78	0.70670	0.50	18.7	43	63
87GV-R8	147	Stony Creek	1.0	0.78	0.70651	0.53	18.8	31	63
87GV-R11	148	Stony Creek	6.9	0.27	0.70526	1.94	15.3	8	54
89GV-66	151	Stony Creek	-1.6	1.02	0.70873	0.37	17.3		64

* From Linn et al. (1992)

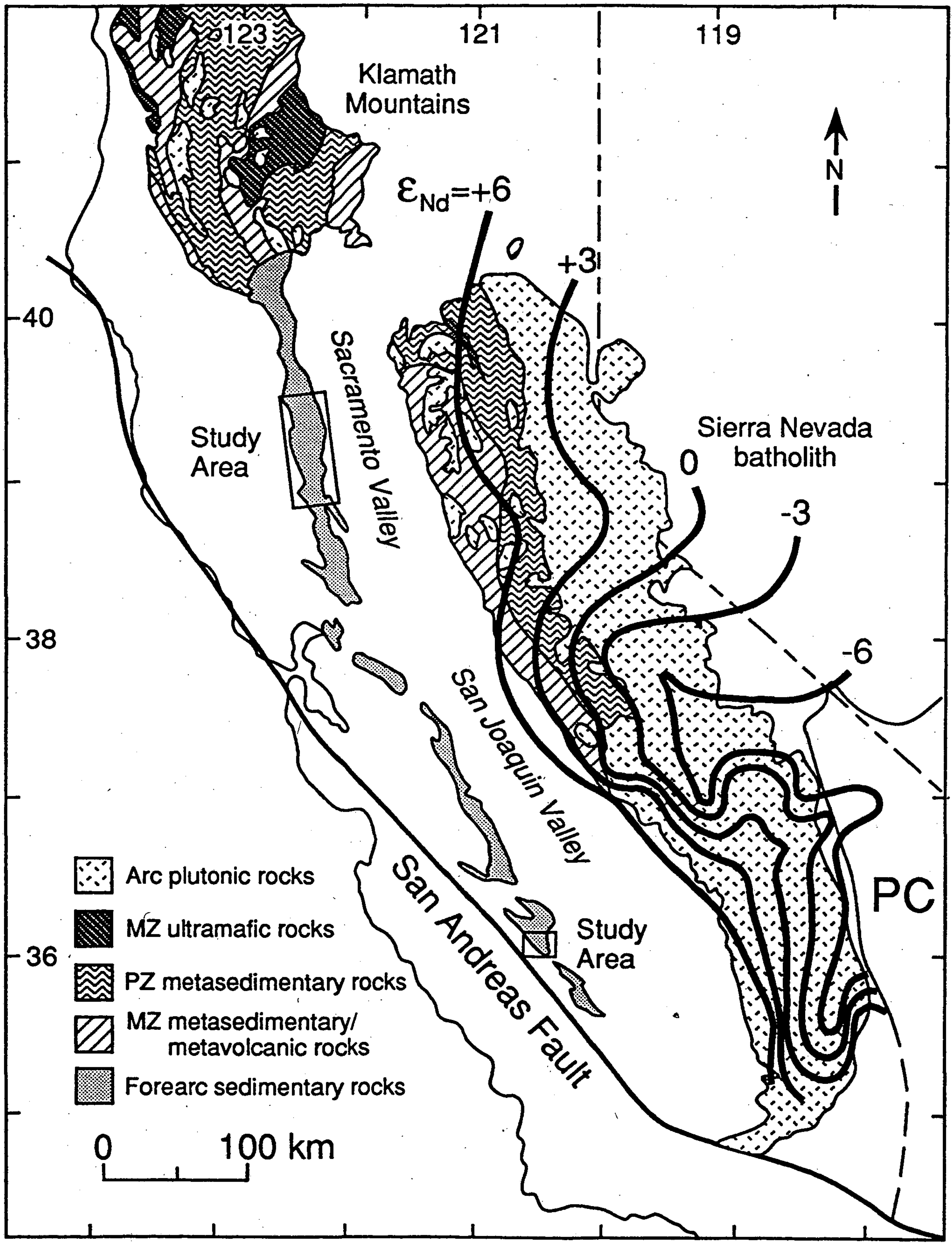

123
121
119
Klamath Mountains
N
εNd=+6
+3
0
-3
-6
40
38
36
Study Area
Sacramento Valley
Sierra Nevada batholith
San Joaquin Valley
San Andreas Fault
Study Area
PC
Arc plutonic rocks
MZ ultramafic rocks
PZ metasedimentary rocks
MZ metasedimentary/ metavolcanic rocks
Forearc sedimentary rocks
0
100 km

Figure 1a

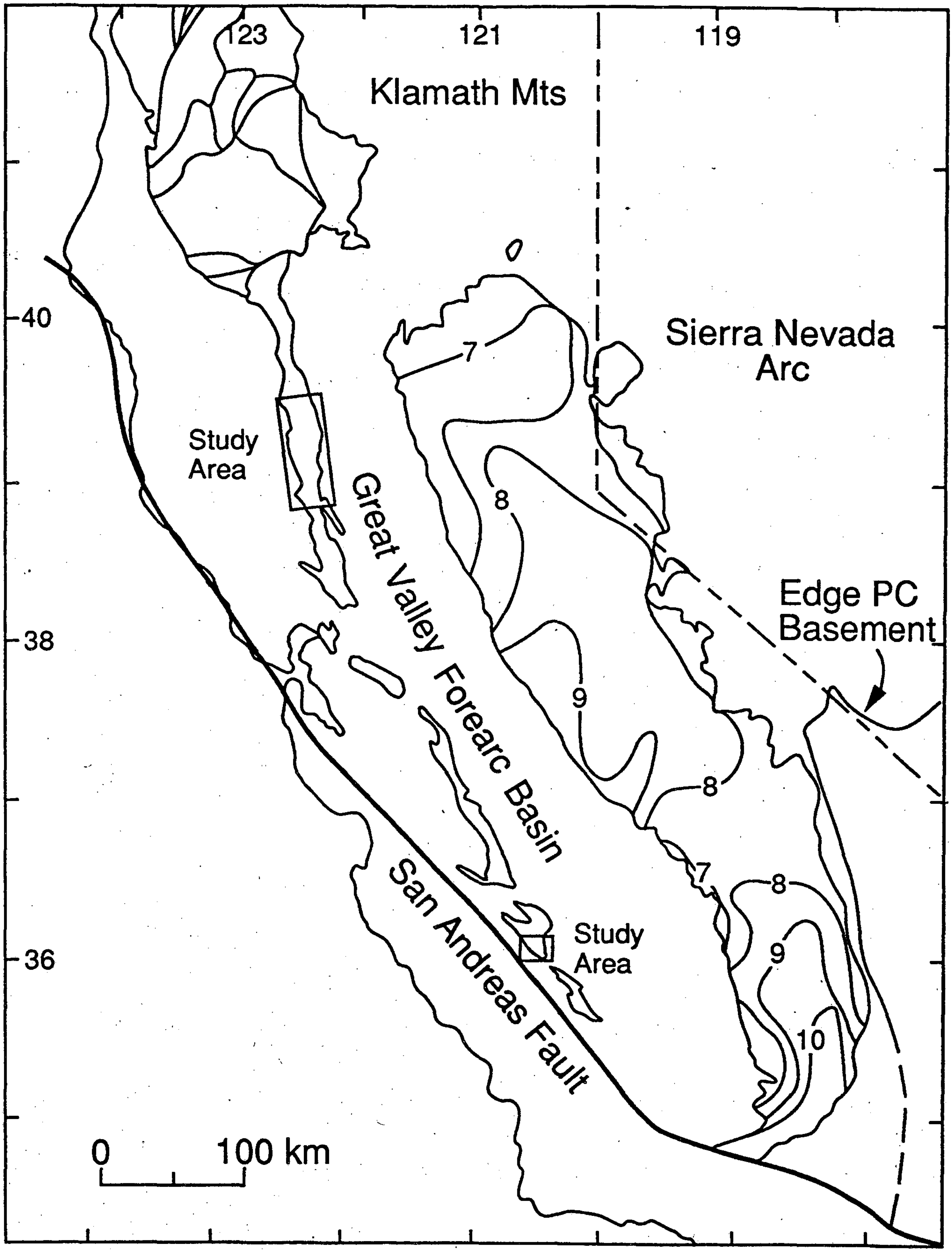

123
121
119
Klamath Mts
Sierra Nevada Arc
40
38
36
Study Area
Great Valley Forearc Basin
7
8
9
8
7
8
9
10
Edge PC Basement
San Andreas Fault
Study Area
0
100 km

Figure 1b

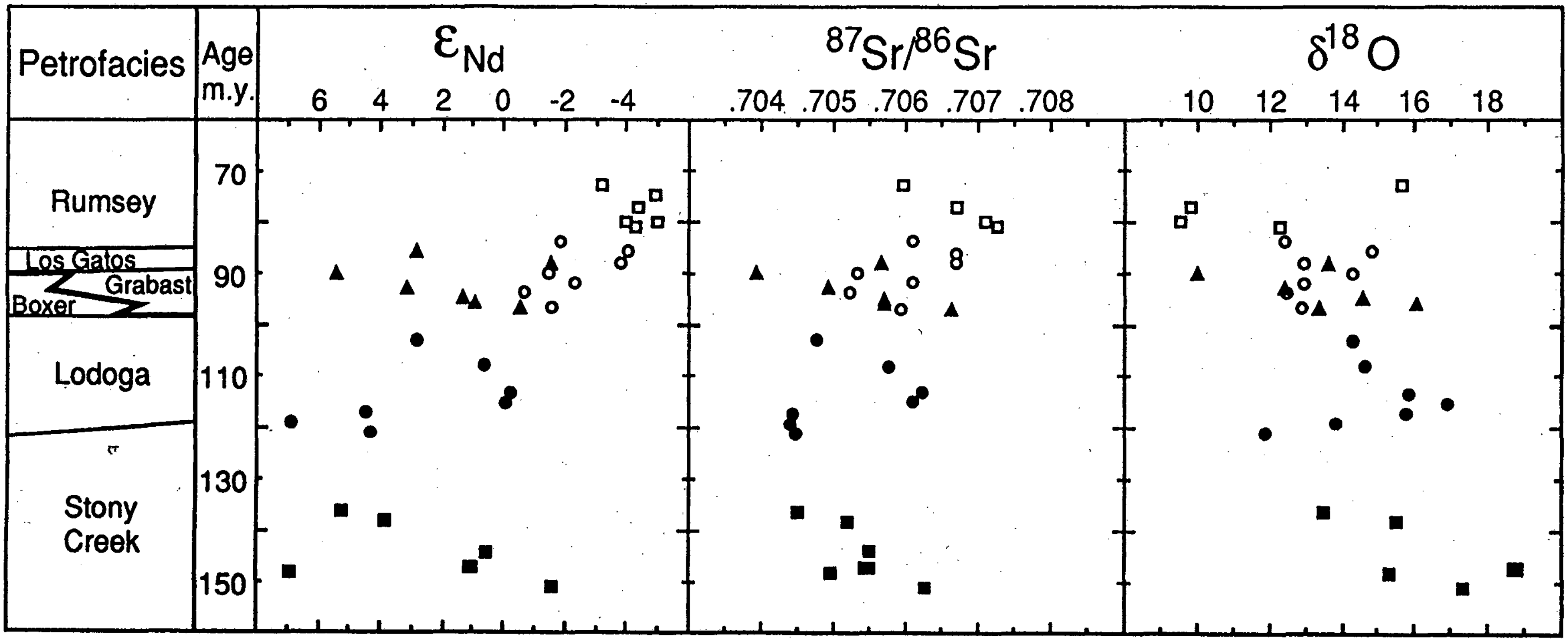

Petrofacies
Age m.y.
ε_{Nd}
6 4 2 0 -2 -4
$^{87}Sr/^{86}Sr$
.704 .705 .706 .707 .708
$\delta^{18}O$
10 12 14 16 18
Rumsey
Los Gatos
Grabast
Boxer
Lodoga
Stony Creek
70
90
110
130
150

Figure 2

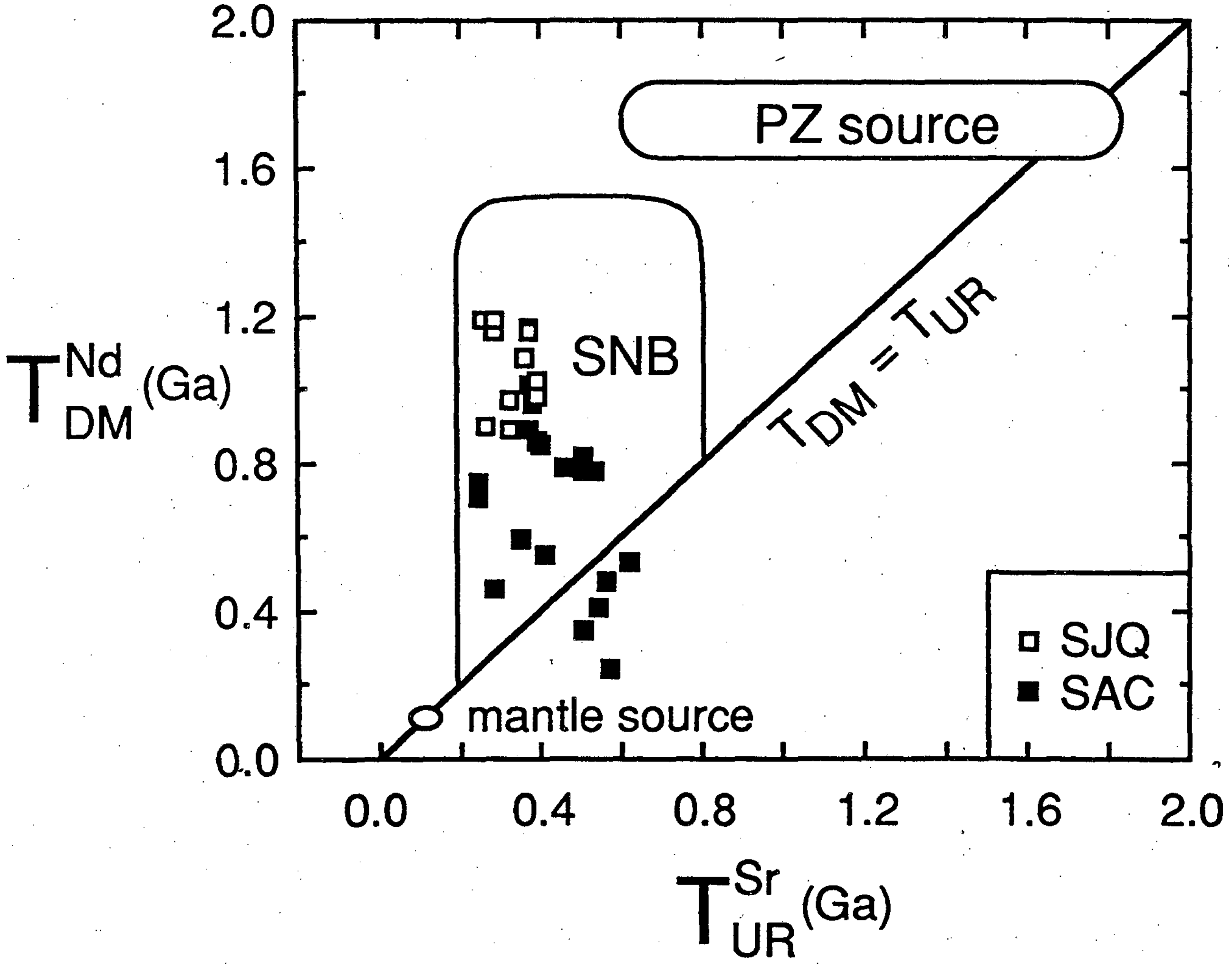

2.0
1.6
1.2
0.8
0.4
0.0
T^{Nd}_{DM} (Ga)
0.0
0.4
0.8
1.2
1.6
2.0
T^{Sr}_{UR} (Ga)
PZ source
SNB
$T_{DM} = T_{UR}$
mantle source
SJQ
SAC

Figure 3

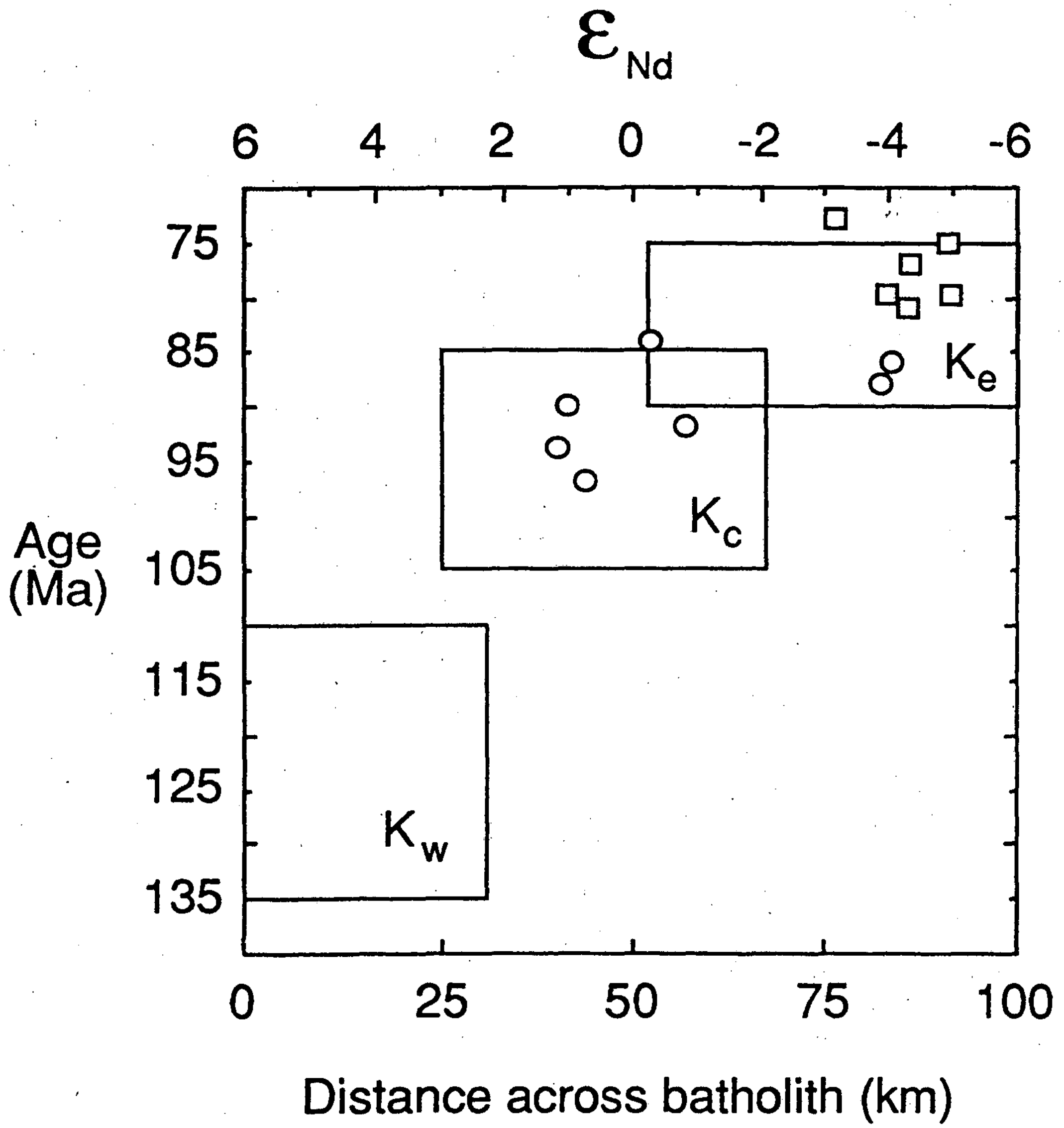
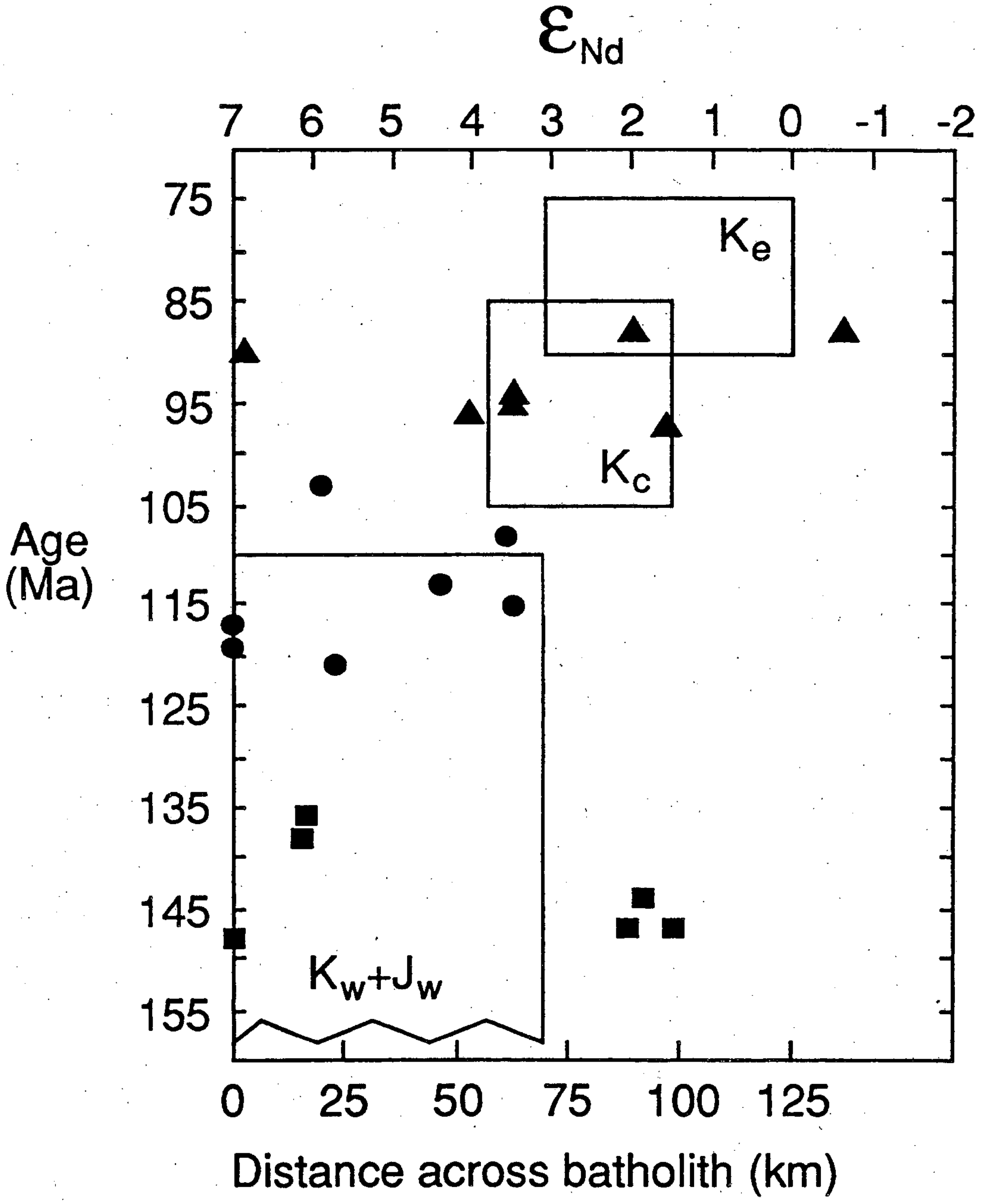

ε_{Nd}
6
4
2
0
-2
-4
-6
75
85
95
105
115
125
135
Age (Ma)
K_e
K_c
K_w
0
25
50
75
100
Distance across batholith (km)

Figure 4a



ε_{Nd}
7
6
5
4
3
2
1
0
-1
-2
K_e
K_c
K_w+J_w
Age (Ma)
75
85
95
105
115
125
135
145
155
0
25
50
75
100
125
Distance across batholith (km)

Figure 4b

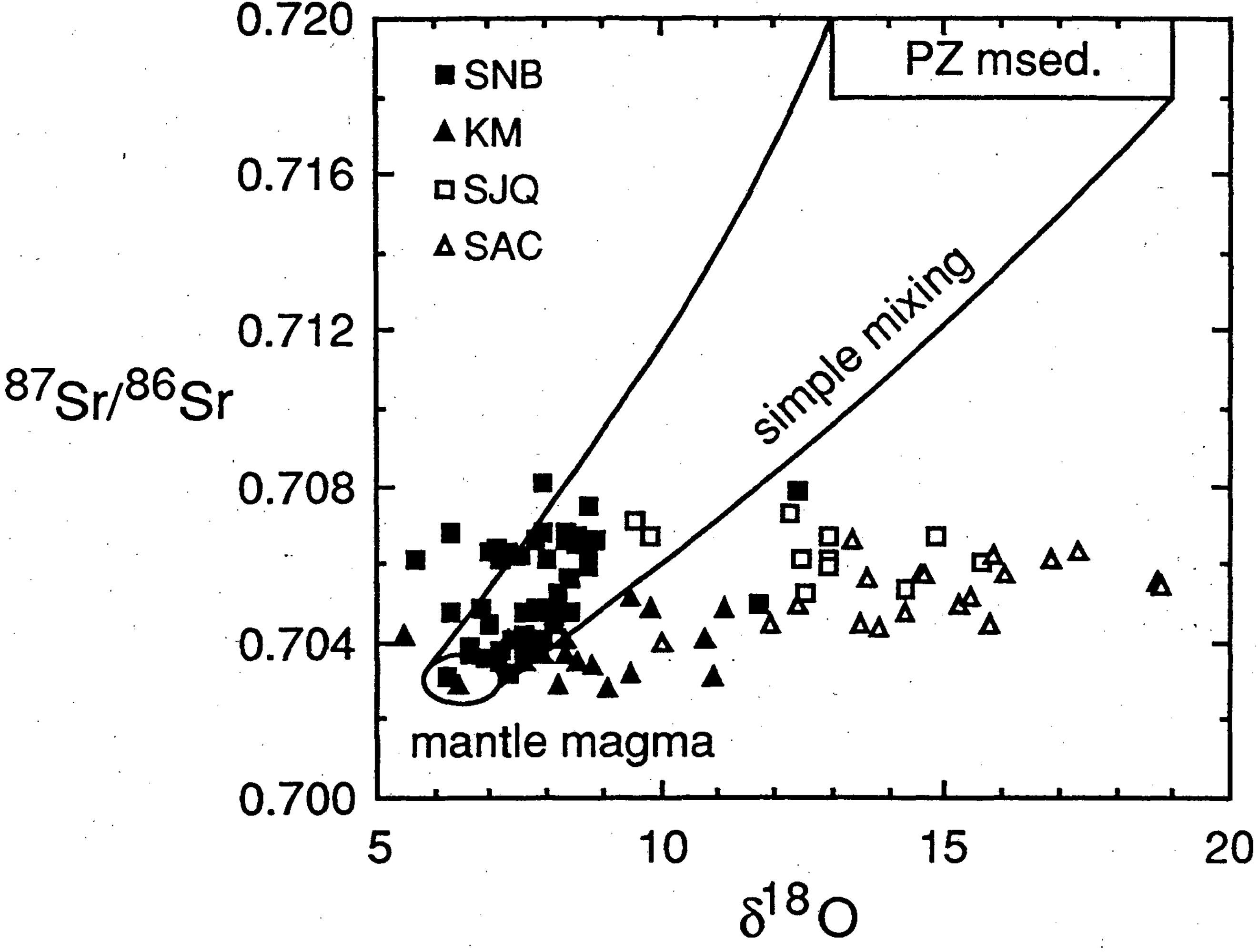

0.720
0.716
0.712
0.708
0.704
0.700
$^{87}Sr/^{86}Sr$
■ SNB
▲ KM
□ SJQ
△ SAC
PZ msed.
simple mixing
mantle magma
5
10
15
20
$\delta^{18}O$

Figure 5

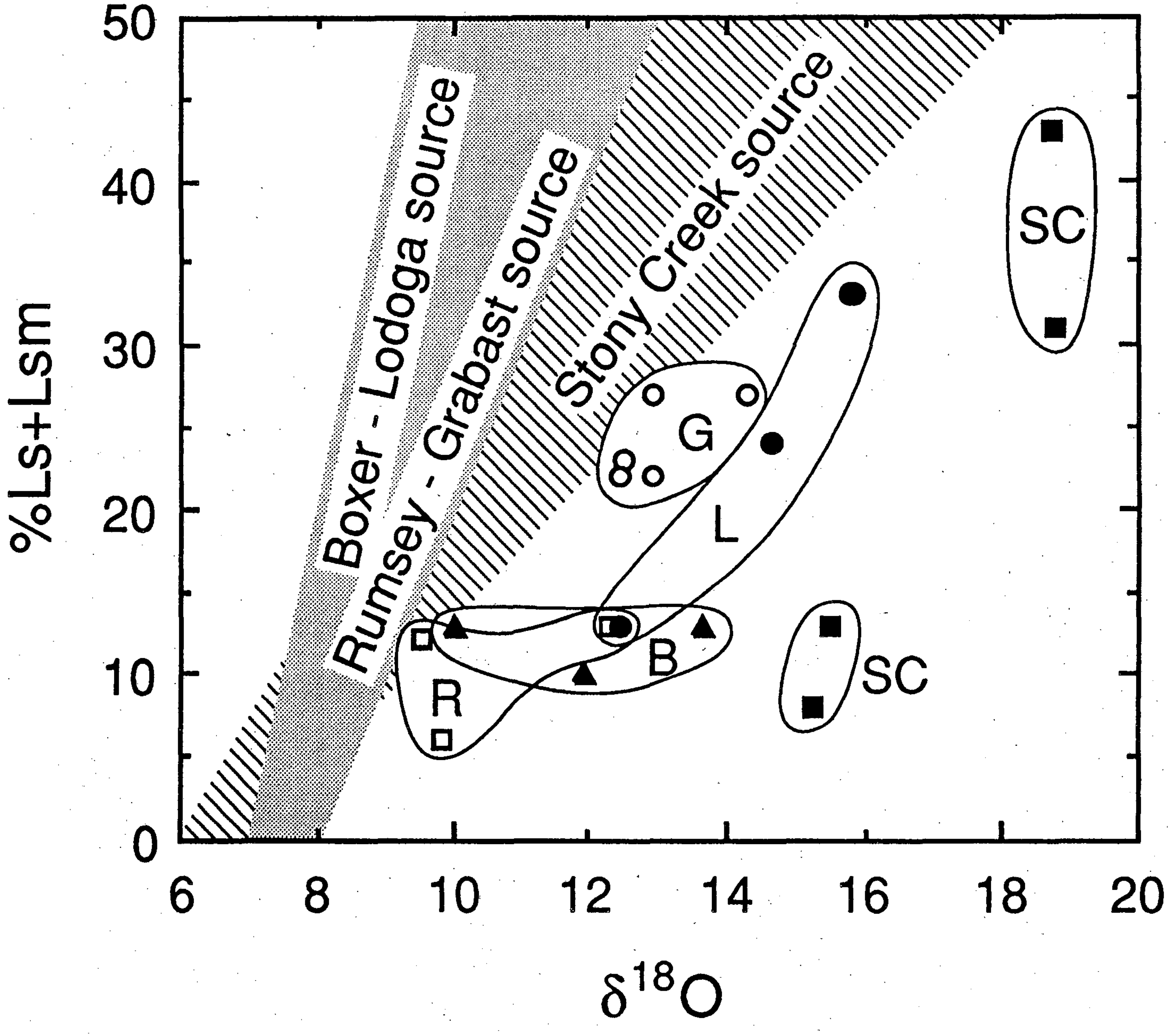
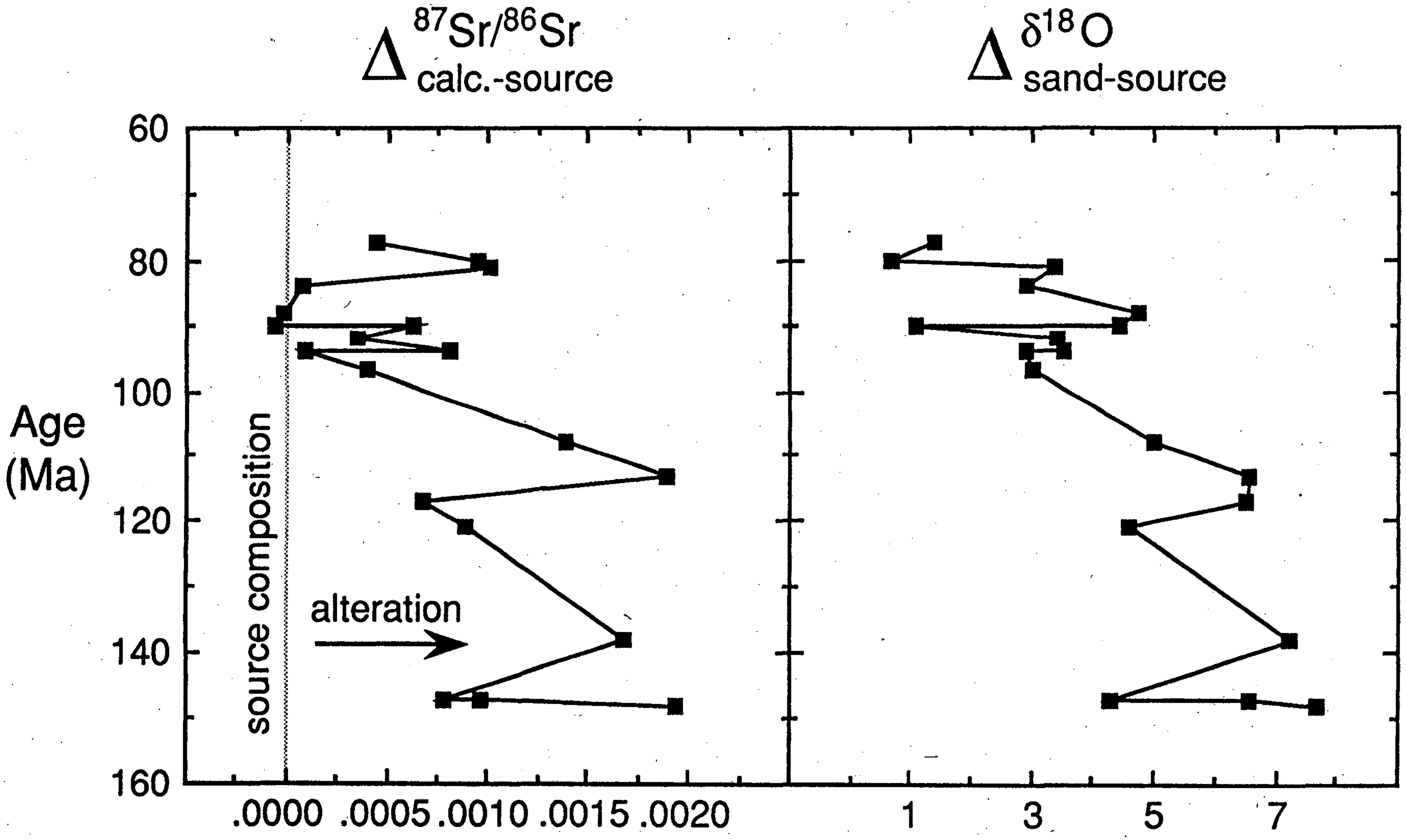
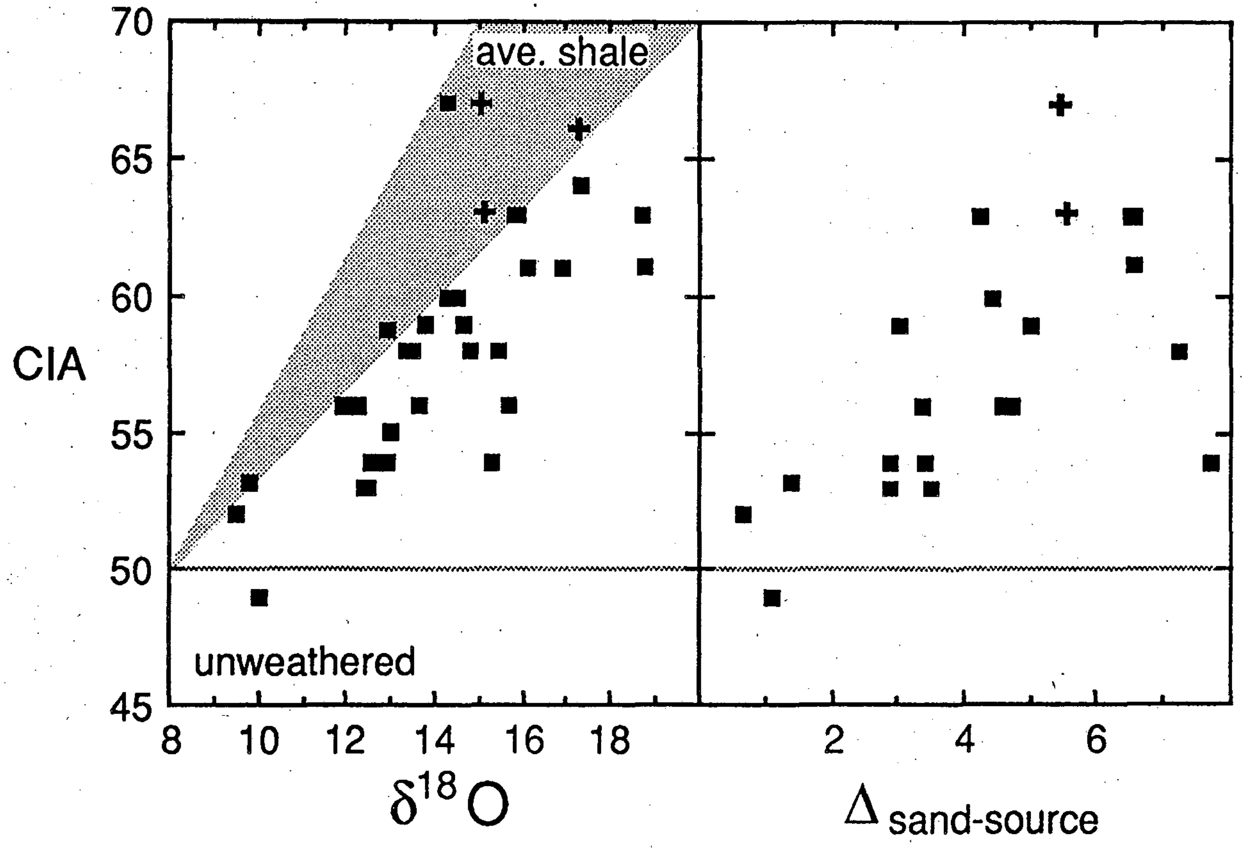

50
40
30
20
10
0
%Ls+Lsm
6
8
10
12
14
16
18
20
δ18O
Boxer - Lodoga source
Rumsey - Grabast source
Stony Creek source
SC
G
8
L
B
R
SC

Figure 6



$\Delta^{87}Sr/^{86}Sr_{calc.-source}$
$\Delta\delta^{18}O_{sand-source}$
Age (Ma)
60
80
100
120
140
160
source composition
alteration
.0000
.0005
.0010
.0015
.0020
1
3
5
7

Figure 7



CIA
70
65
60
55
50
45
ave. shale
unweathered
8
10
12
14
16
18
$\delta^{18}O$
2
4
6
$\Delta_{sand-source}$

Figure 8

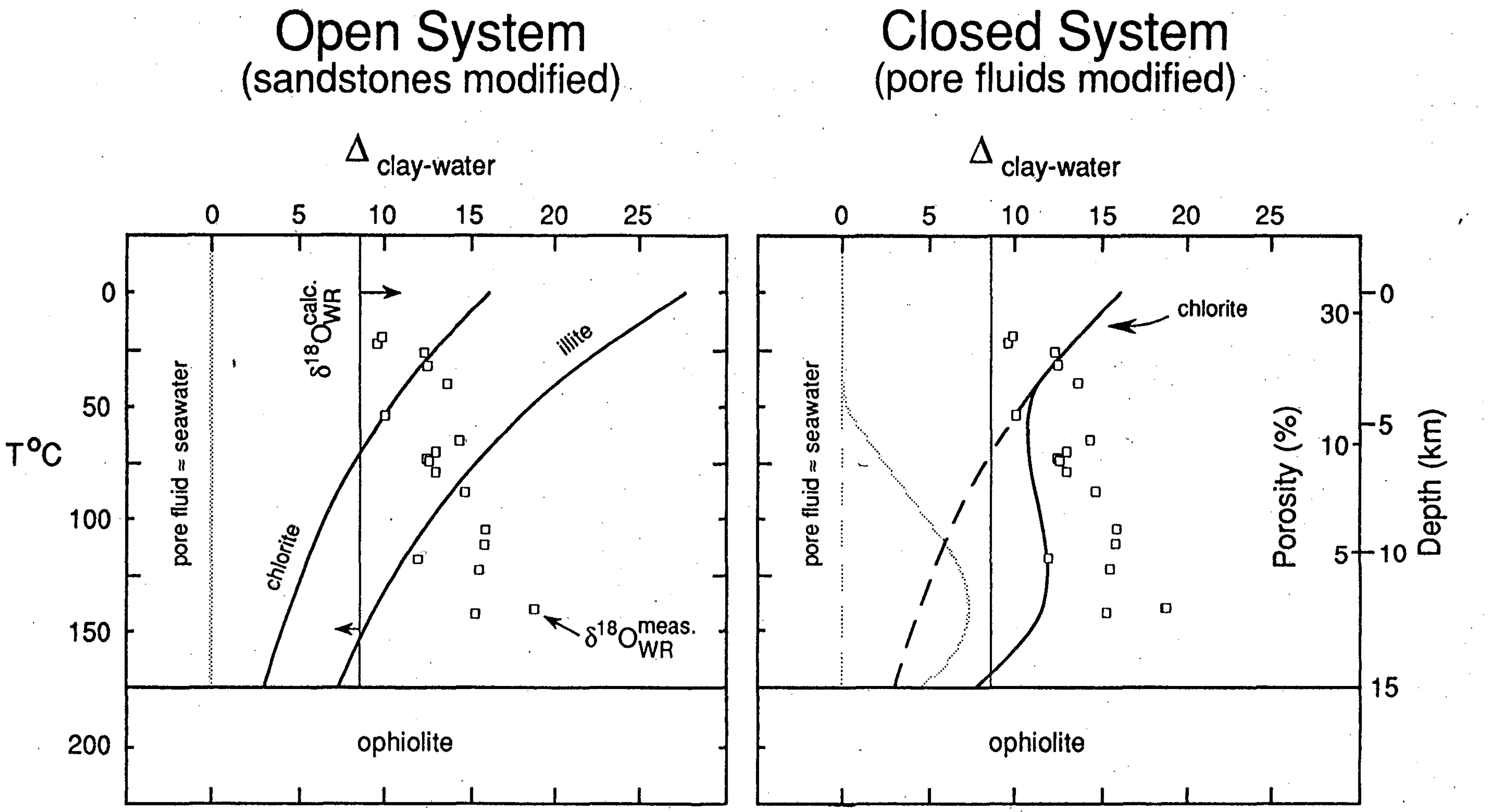

Open System
(sandstones modified)
Closed System
(pore fluids modified)
$\Delta_{\text{clay-water}}$
0
5
10
15
20
25
$\Delta_{\text{clay-water}}$
0
5
10
15
20
25
T°C
0
50
100
150
200
pore fluid = seawater
$\delta^{18}O_{WR}^{calc.}$
chlorite
illite
$\delta^{18}O_{WR}^{meas.}$
ophiolite
pore fluid = seawater
chlorite
ophiolite
Porosity (%)
30
10
5
Depth (km)
0
5
10
15

Figure 9

LAWRENCE BERKELEY LABORATORY
UNIVERSITY OF CALIFORNIA
TECHNICAL INFORMATION DEPARTMENT
BERKELEY, CALIFORNIA 94720