The record of Middle Jurassic volcanism in the Carmel and Temple Cap Formations of southwestern Utah

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ABSTRACT

Altered volcanic ash beds in the Middle Jurassic Temple Cap and Carmel Formations in southwestern Utah record a pulse of active arc-related volcanism between 166 and 171 Ma. A second pulse between 148 and 155 Ma has previously been documented in the Upper Jurassic Morrison Formation. Volcanic and volcaniclastic rocks of these same ages have also been identified closer to or within the arc in California in the Inyo Mountains, the Cowhole Mountains, the Palen Mountains, and the central Mojave Desert. The upper part of the volcaniclastic Mount Wrightson Formation and the strata of Cobre Ridge in southern Arizona are ca. 170 Ma in age and appear to be time correlative with the Middle Jurassic formations in southwestern Utah.

The altered ash beds found in the Temple Cap and Carmel Formations typically contain phenocrysts of sanidine, quartz, biotite, apatite, zircon, and titanite. Plagioclase was likely present originally in all of the ashes, but was removed by alteration and is now found only in the Temple Cap red beds. Quartz and sanidine are absent in two crystal-poor ash beds that contain two pyroxenes, hornblende, and biotite. Although major and trace element concentrations in the ash beds have been substantially modified, compositions of relict phenocrysts reveal that the magmas were calc-alkaline rhyolites to andesites. Two-pyroxene, two-feldspar, biotite, and biotite-apatite thermometers suggest that crystallization occurred at temperatures ranging from 740 to 910 °C. Hornblende geobarometry yields pressures of 1–2 kilobars for the two ash beds that contain the appropriate buffer assemblage. The mafic silicates all have moderately high Mg/Fe ratios. This fact, combined with the presence of hornblende, biotite, and titanite, suggests that the phenocrysts crystallized at high oxygen fugacities similar to those of the granites of the batholiths of California. The ash probably erupted from a low-lying arc cut by strike-slip faults in what is now southern California and western Nevada.

Major Jurassic unconformities occur near or within the ash-bearing formations in southwestern Utah. Laser-fusion single-crystal $^{40}Ar/^{39}Ar$ measurements have defined the ages of the unconformities and the associated volcanism. The age of the J-1 unconformity, found at the base of the Temple Cap Formation in southwestern Utah, is older than ca. 170.5 Ma. The J-2 unconformity, which lies between the Temple Cap and Carmel Formations, formed between ca. 169 and 168 Ma. The origin of these unconformities is still unclear, but may be related to the Middle Jurassic pulse of magmatism and the oblique plate convergence along the western margin of North America. The age range of ash beds in the Carmel Formation between 166.3 and 168.0 ± ~0.5 Ma is consistent with a Bajocian-Bathonian boundary of ca. 166 Ma.

Keywords: Carmel Formation, geochronology, Jurassic, Temple Cap Formation, volcanism.

INTRODUCTION

The Middle Jurassic Temple Cap and Carmel Formations of southwestern Utah contain numerous altered volcanic ash beds (Wright and Dickey, 1963; Marvin et al., 1965; Nielson, 1988; Everett, 1989; Everett et al., 1989; Christiansen et al., 1994; Blakey and Parnell, 1995; Zhang, 1996) erupted from a low-standing volcanic arc located to the west and southwest (Fig. 1) in present-day California and Arizona (Busby-Spera, 1988). These more distal air-fall ash beds preserve a record of volcanic activity, particularly of large-volume ash eruptions, that may not be as readily available or identifiable along the remnants of the preserved arc. The preservation of datable mineral phases in most of these bentonitic ash beds is excellent and allows precise determination of the timing of the volcanic eruptions.

Controversy over the correlation of Middle Jurassic sedimentary rocks in the Colorado Plateau to their equivalents in the volcanic arc has been difficult to resolve owing to a lack of a precise chronology for the sedimentary rocks. Some authors (Busby-Spera, 1988, 1990; Adams et al., 1991; Riggs et al., 1993; Schermer and Busby, 1994; Fackler-Adams et al., 1997) have correlated the eolian sandstones found in and near the arc to the Carmel and Temple Cap Formations in the Colorado Plateau. Others (Bilodeau and Keith, 1986; Marzolf, 1990, 1991, 1994; Riggs and Busby-Spera, 1990; Marzolf and Lucas, 1996) have

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correlated arc-related eolian sandstones, in some places called the Aztec Sandstone, with the Navajo-Aztec Sandstone to the east. The difference in age between (1) the Carmel and Temple Cap Formations and (2) the Navajo Sandstone is unknown, however. Recent additions to the age database in the arc region (Graubard et al., 1988; Busby-Spera et al., 1989; Dunne and Walker, 1993; Riggs et al., 1993; Schermer and Busby, 1994; Fackler-Adams et al., 1997), along with our new dates for ash beds in the Carmel and Temple Cap Formations, have allowed us to make much more precise correlations between the two areas, as we describe.

Stratigraphy

In southwesternmost Utah, the Lower Jurassic to lower Middle Jurassic rocks consist of, from oldest to youngest, the Navajo Sandstone, the Temple Cap Formation, and the Carmel Formation (Hintze, 1988; Luttrell, 1993; Marzolf and Lucas, 1996). Two major unconformities, the J-1 and J-2, separate the Navajo from the Temple Cap and the Temple Cap from the Carmel, respectively (Pipiringos and O'Sullivan, 1978; Peterson and Pipiringos, 1979). Other Middle Jurassic and Upper Jurassic formations found elsewhere in Utah are missing in southwestern Utah.

The Temple Cap Formation is limited to southwestern Utah and extends slightly into northern Arizona. It is composed in its eastern exposures of two members, a lower reddish silty sandstone to mudstone member, the Sinawava Member, and a light-gray, cross-bedded eolian sandstone member, the White Throne Member (Peterson, 1994). However, in southwesternmost Utah in our study area, the White Throne Member is replaced by red beds similar to the Sinawava Member, and the two members are lumped together because they cannot be distinguished. Bentonitic ash beds occur throughout the entire red-bed sequence in the study area; as many as eight different ash beds have been identified in measured sections (Peterson and Pipiringos, 1979).

The Carmel Formation in southwestern Utah is divided into four members. From oldest to youngest, they are the Co-op Creek (or Judd Hollow) Member, the Crystal Creek Member, the Paria River Member, and the Winsor Member (Blakey et al., 1983; Blakey, 1994b; Peterson, 1994). The altered volcanic ash beds in southwestern Utah are found in the two lower members (Nielson, 1988). Seven different ash beds have been identified in some measured sections (Nielson, 1988; Everett, 1989).

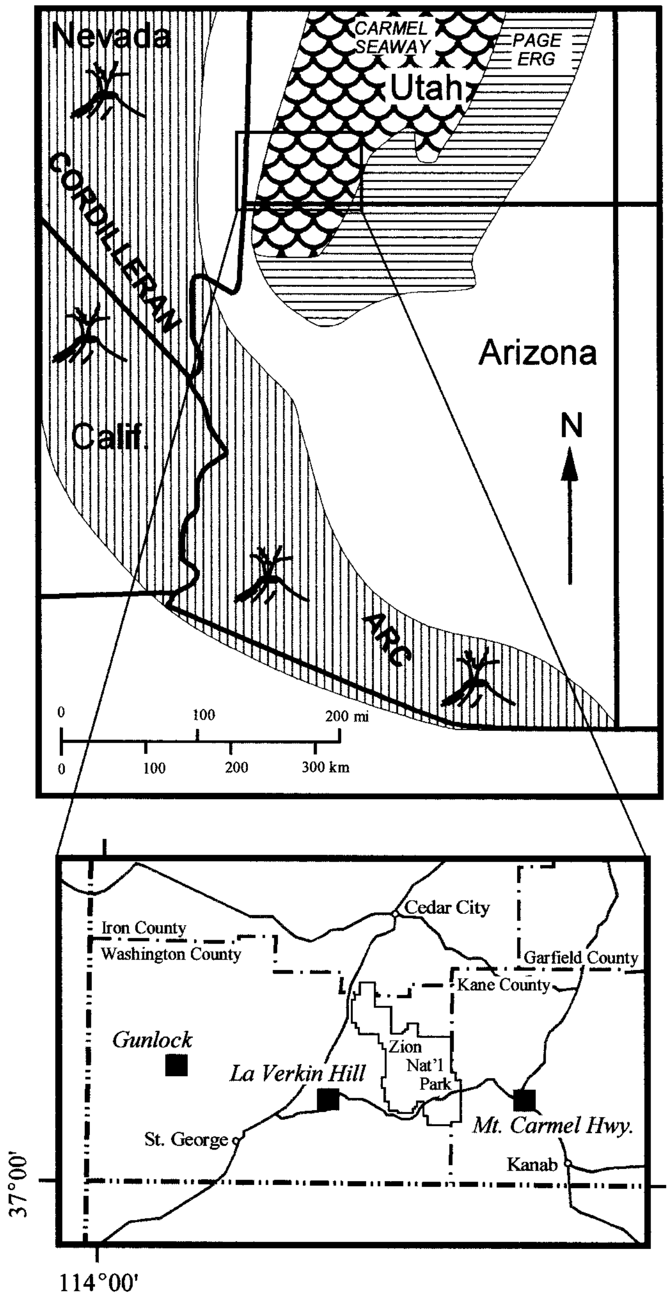

Figure 1. Index map showing paleogeography and the locations of study sections. The Gunlock locality includes, in close proximity, the Gunlock section, Hill 3815 section, and the Manganese Wash cutbank section.

Ages of Formations

The age of many of the Jurassic formations here and throughout the Colorado Plateau region is generally poorly defined, although significant advances have been made in understanding the age of the Upper Jurassic Morrison Formation (Kowallis et al., 1991, 1998) and the Carmel Formation (Everett et al., 1989; Kowallis et al., 1994). Marvin et al. (1965) were the first to attempt to obtain accurate radiometric ages for the altered ash beds in the Carmel Formation in southwestern Utah. They used K-Ar and Rb-Sr methods to date biotite from ash beds at several localities. The low K content (<5%) of their biotites indicates that the biotites were altered, and several of their ages were considerably younger than expected for the Middle Jurassic Carmel Formation. The few K-Ar ages that seemed to

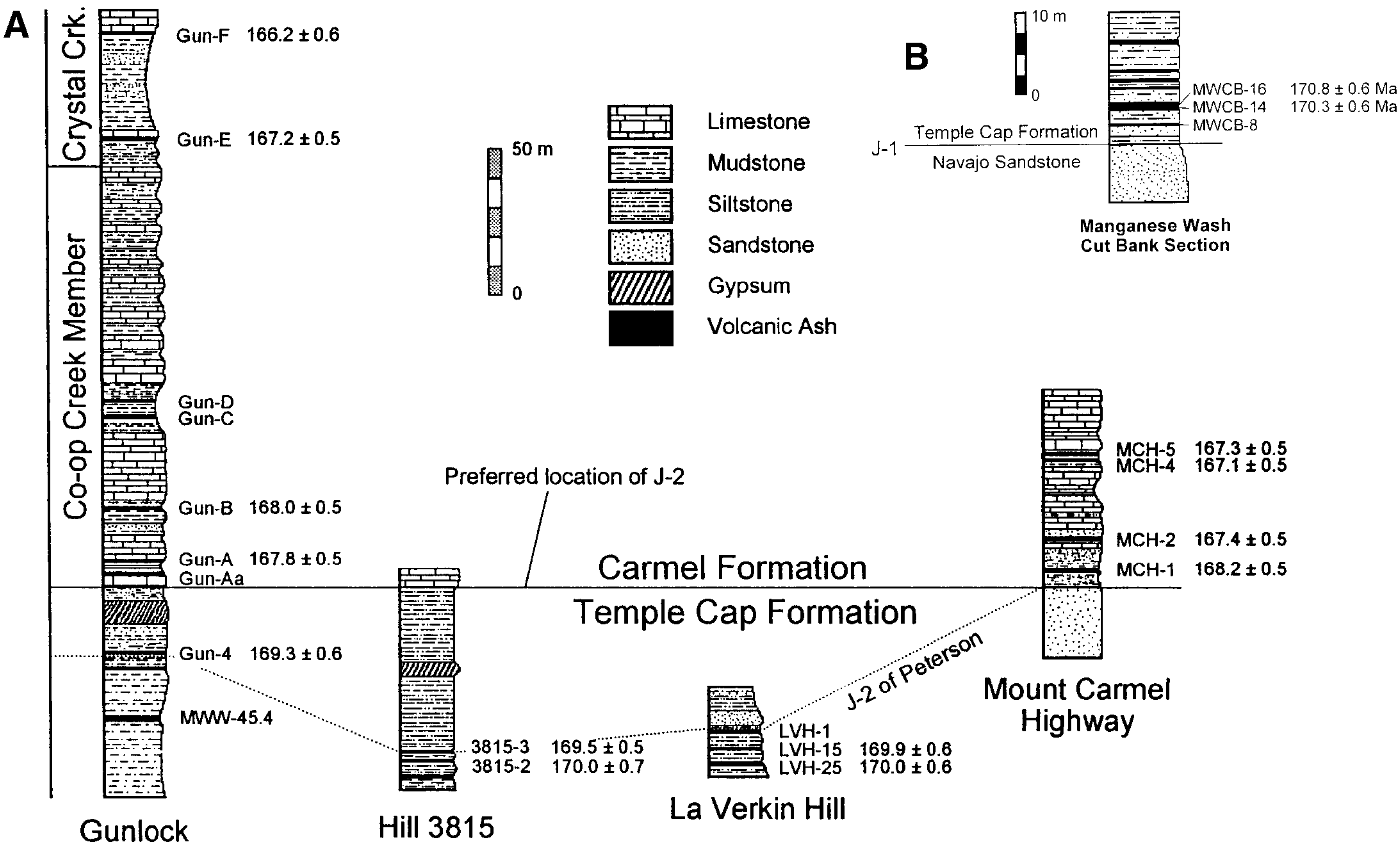

Figure 2. (A) Stratigraphic sections of Carmel and Temple Cap Formations showing the locations of dated ash beds and their relationship to the J-2 unconformity. (B) Short section of Temple Cap Formation and Navajo Sandstone showing ash bed dated near J-1 unconformity.

be most reliable (samples with the highest K content) gave ages of ca. 168 Ma (corrected here by using decay constants of Dalrymple, 1979). We have resampled the localities of Marvin et al. (1965) and added many more precise ages and localities to the database.

Although considerable effort has been expended on analyzing and characterizing the volcanic rocks within the Middle Jurassic arc (Dunne, 1986; Karish et al., 1987; Busby-Spera, 1988; Busby-Spera et al., 1990; Adams et al., 1991; Dunne and Walker, 1993; Riggs et al., 1993; Schermer and Busby, 1994; Sorensen et al., 1998), very little work has been done to characterize the volcanic deposits farther from the arc (Chapman, 1989; Everett, 1989; Blakey and Parnell, 1995; Zhang, 1996). Just as in the Cretaceous Western Interior Seaway, where an excellent record of volcanism is well preserved (Hattin, 1971; Elder, 1988; Kirkland, 1991; Kauffman and Caldwell, 1993; Obradovich, 1993) without the complicated structure, sedimentation, and deformation characteristic of the Cretaceous arc region, so also in the Middle Jurassic rocks of southwestern Utah we find an excellent record of large, explosive volcanic eruptions, particularly during deposition of the Page Sandstone (Blakey and Parnell, 1995), the Temple Cap Formation, and the lower Carmel Formation.

RESEARCH METHODS

Sample Localities and Collection

Altered ash beds were collected from measured sections of the Temple Cap and Carmel Formations near Gunlock, La Verkin, Mount Carmel Junction, and other localities in southwestern Utah (Figs. 1 and 2). The sample localities are described in detail by Nielson (1988, 1990, Gunlock section), Everett (1989, Mount Carmel Highway section), and Zhang (1996, Manganese Wash cutbank section). Short, partial sections from which samples of ash were collected are the Hill 3815 section (NE¼NE¼ sec. 31, T. 40 S., R. 17 W.) and the La Verkin Hill section (SW¼NE¼ sec. 22, T. 41 S., R. 13 W.).

When an ash bed was located within a section, the area was first cleared of overburden with a shovel or hoe and then ~10 kg of material was carefully collected to prevent contamination of the altered ash from rocks above and below. Most of the ash beds contained visible biotite flakes, had a waxy feel because of swelling clays, and would be called bentonites. Those without obvious biotite flakes were still bentonitic, but are sometimes more difficult to recognize in the field.

The stratigraphically youngest ash bed is Gun-F, which is located near the top of the Crystal Creek Member of the Carmel Formation ~4 m below the first limestones of the overlying Paria River Member (Fig. 2A), whereas the oldest ash bed stratigraphically is MWCB-8 collected ~4.5 m above the base of the Temple Cap Formation (Fig. 2B).

Mineral Separation and Analysis

A split from each sample was crushed and pulverized for chemical analysis, and the rest was utilized in extraction of mineral phases. The samples in which the clays were easily disaggregated were washed in water, and the phenocrysts were concentrated. Samples that did not disaggregate in water were crushed and sieved, and then the 60–200-mesh size fraction was washed over a Wilfley table to concentrate and separate the phenocrysts from the remaining matrix. Phenocryst residues were then processed through standard magnetic and heavy-liquid procedures to produce concentrates of sanidine, quartz, plagioclase, biotite, hornblende, pyroxene, zircon, titanite, and apatite. Some handpicking of minerals was required to obtain relatively pure fractions for electron-microprobe analysis and single-crystal, laser-probe $^{40}Ar/^{39}Ar$ age determinations. Mineral analyses were carried out by using a CAMECA SX-50 electron microprobe at the University of Utah. The acceleration voltage was 15 kV; the beam size was 30 μm, and both natural and synthetic standards were used.

Sanidine (60–100-mesh size) age determinations were performed at the Berkeley Geochronology Center by using the techniques described in detail in Deino and Potts (1990), Deino et al. (1990), Chesner et al. (1991), and Kowallis et al. (1995). Sanidine from the Fish Canyon Tuff of Colorado, with a reference age of 27.84 Ma (Samson and Alexander, 1987; Cebula et al., 1986), was used as the age standard for sample grains 6XXX and 20XXX. Sanidine from a Jurassic Morrison Formation ash with a reference age of 148.97 Ma (calibrated against Fish Canyon Tuff at 27.84 Ma) was used as the standard for samples 506X (Kowallis et al., 1998). Table 1 gives weighted mean averages for laser-fusion single-crystal $^{40}Ar/^{39}Ar$ ages for each of the sanidine separates analyzed. Complete laser-fusion single-crystal $^{40}Ar/^{39}Ar$ age data are available in Data Repository Table DR1[1].

Detrital microcline and sanidine grains are common in the Gun-F ash, but are rare or not present in any of the other samples. Detrital grains in Gun-F range in age from 209 Ma (Late Triassic) to 884 Ma (late Precambrian), with a mode in the Late Triassic at 224 Ma and perhaps a second broad mode at ca. 300–350 Ma in the Mississippian–Pennsylvanian. The Late Triassic mode is the same age as the ash-rich beds of the Petrified Forest Member of the Chinle Formation (Christiansen et al., 1994) from which they were possibly eroded. The provenance of grains older than Triassic is more problematic, although possible bentonitic beds occur in the Mississippian in parts of Utah.

TABLE 1. MEAN SINGLE-CRYSTAL $^{40}Ar/^{39}Ar$ LASER PROBE AGES

Sample numbers	Description	Grains dated (number)	Average $^{40}Ar^*$ (%)	Mean age ± 1σ (Ma)
Gun-A	15.1 m above Carmel-Temple Cap contact	10	99.4	167.84 ± 0.49
Gun-B	37.7 m above Carmel-Temple Cap contact	7	99.6	167.95 ± 0.51
Gun-E	154 m above Carmel-Temple Cap contact	8	99.4	167.22 ± 0.49
Gun-F	190 m above Carmel-Temple Cap contact	6	99.6	166.18 ± 0.64
Gun-4	24.0 m below Carmel-Temple Cap contact	8	99.8	169.27 ± 0.59
Hill 3815–2	49.1 m below Carmel-Temple Cap contact	7	99.0	170.04 ± 0.65
Hill 3815–3	47.9 m below Carmel-Temple Cap contact	12	99.8	169.54 ± 0.49
LVH-15	about 50 m below Carmel-Temple Cap contact	8	99.5	169.92 ± 0.59
LVH-25	about 53 m below Carmel-Temple Cap contact	9	99.4	169.95 ± 0.57
MCH-1	4.9 m above Carmel-Temple Cap contact	3	99.5	168.15 ± 0.54
MCH-2	15.1 m above Carmel-Temple Cap contact	5	99.5	167.44 ± 0.51
MCH-4	42.0 m above Carmel-Temple Cap contact	5	99.5	167.09 ± 0.51
MCH-5	43.4 m above Carmel-Temple Cap contact	8	99.5	167.30 ± 0.49
MWCB-14	4.3 m above Temple Cap-Navajo Sandstone contact	8	99.4	170.25 ± 0.55
MWCB-16	4.9 m above Temple Cap-Navajo Sandstone contact	8	99.4	170.81 ± 0.56

Notes: Weighted mean ages are calculated using the inverse variance as the weighting factor (Taylor, 1982), while errors in the weighted mean ages are 1σ standard error of the mean (Samson and Alexander, 1987).
*Radiogenic.

MINERALS IN THE ASH BEDS

Phenocryst Abundance

The relative abundances of phenocryst types were determined for each ash layer. This information provides insights into the original composition of the volcanic ash and helps to determine how many different ash layers were present in the Temple Cap and Carmel Formations. After obtaining a mineral residue from each altered ash bed and separating this residue into light and heavy fractions by heavy-liquid separation, a qualitative estimate was made of the relative abundance of each type of phenocryst in both the light and heavy fractions. These data are shown in Tables 2 and 3 with each mineral recorded as abundant, common, rare, or not present. Biotite, zircon, and apatite were present in all of the samples processed; other minerals, such as quartz, sanidine, and titanite were present in most of the altered ashes; and hornblende, pyroxene, and plagioclase were only found in a few samples. Data tables containing electron-microprobe analyses of most of the mineral phases discussed are available from the GSA Data Repository (see footnote 1).

Feldspar

Feldspar from the samples includes sanidine, plagioclase, and rare xenocrystic or detrital microcline. Given the ubiquity of plagioclase in calc-alkaline volcanic rocks, it is probable that it was present in all of the ashes when they were deposited, but now is only found in two of the ashes deposited in the Temple Cap Formation red beds. The Carmel Formation contains abundant shallow-marine carbonates. In shallow marine environments, the activity of Ca^{2+} in aqueous solutions is limited by the solubility of calcite (Faure, 1998). In calcite-saturated solutions, the $[Ca^{2+}]/[H^+]$ ratio is restricted to low values, and anorthite is unstable. The stability of sanidine is unaffected by these diagenetic conditions, and it appears completely unaltered when examined with a petrographic microscope.

All of the ash beds, except Gun-C and Gun-D, have sanidine. The compositions of sanidines are generally similar and range from Or_{65} to Or_{75} with little or no zoning. The content of Ba in sanidine from MCH-1, Gun-B, and Gun-4 is higher than that in other samples. There are a few grains with Ba contents above 0.05 atoms per formula unit (apfu). In general, the compositions of sanidine are similar to those in other rhyolitic and dacitic ignimbrites (Table DR2; see footnote 1).

Microprobe analyses of plagioclase were obtained from two samples (Gun-4 and MWW-45.4) in the Temple Cap Formation (Table DR3; see footnote 1). Although a small range of plagioclase compositions exists, no zoning of individual plagioclase grains was detected during analysis. The anorthite content of plagioclase ranges from 30% to 40% (Fig. 3). The homogeneity of feldspar compositions suggests homogeneity of the parent magmas. The compositions of the plagioclases from the two ashes in the Temple Cap Formation are similar to those of plagioclase from the dacitic Fish Canyon Tuff (Whitney and Stormer, 1985) and plagioclase from dacitic and rhyolitic rocks in general (Ewart, 1979). However, there are some differences compared to pla-

[1]GSA Data Repository item 2001030, electron microprobe and isotopic data, is available on the Web at http://www.geosociety.org/pubs/ft2001.htm. Requests may also be sent to Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301; e-mail: editing@geosociety.org.

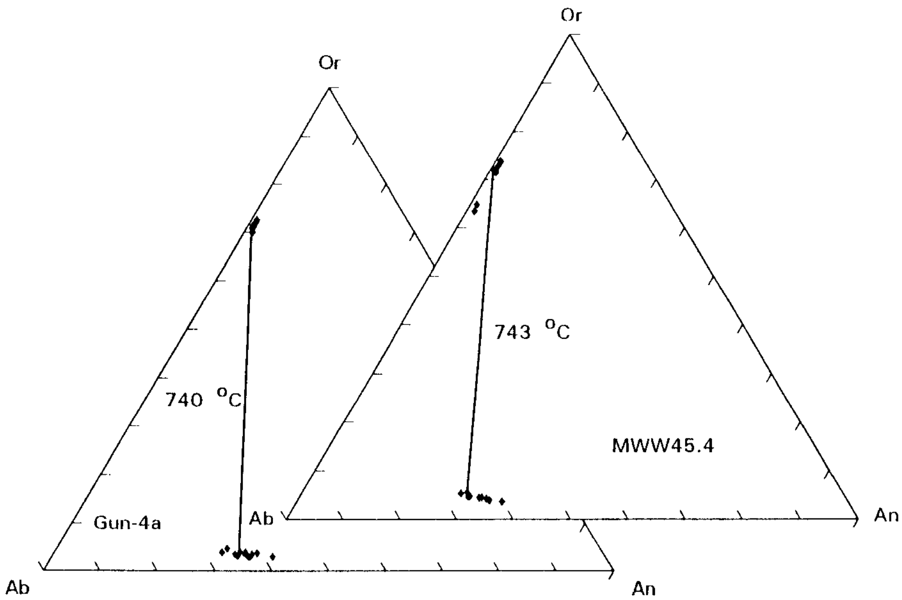

Figure 3. Ternary diagrams for feldspars from samples Gun-4a and MWW-45.4, the only two samples with both plagioclase and sanidine. Temperatures were estimated following the method of Fuhrman and Lindsley (1988).

gioclases in high-silica rhyolites like the Bishop Tuff (Hildreth, 1977) or the strongly zoned Pahranagut Tuff (Best et al., 1995) where Na_2O is higher and CaO and Al_2O_3 are lower.

The two-feldspar geothermometer of Fuhrman and Lindsley (1988) was used to estimate magmatic temperatures for two ash beds that contained both sanidine and plagioclase. The compositions used in the calculations are shown in Figure 3. Average sanidine compositions were matched with an average of the most sodic of the plagioclase grains. The temperature for MWW-45.4 was 743 °C and for Gun-4 was 740 °C at an assumed pressure of 2 kilobars. These eruption temperatures are reasonable for rhyolite magmas with two feldspars, quartz, and biotite. The similarity of the other sanidine compositions in other ash beds suggests that all of the sanidine-bearing rocks were probably erupted at about the same temperature. Only Gun-F and MCH-2 have sanidines with distinctively lower Or contents (about 0.65–0.70 vs. 0.70–0.75). Those ash beds that lack sanidine (Gun-C and Gun-D) probably erupted at temperatures that exceeded 750 °C.

Biotite

Biotite occurs in all of the ash samples and is the most abundant mafic phase in all but ash beds Gun-C and Gun-D. In most respects, the biotite from almost all of the ash beds is similar (Table DR4; see footnote 1). Only MCH-5 is distinct, with a higher Mn content than the others. Compositions of biotites in terms of molar Fe/(Fe + Mg) and total Al relative to ideal end members fall in the area of calc-alkaline igneous rocks from the western United States (Christiansen et al., 1986; Fig. 4). They are more magnesian than the biotite of the Bishop Tuff and other highly evolved rhyolites (Hildreth, 1977; Christiansen et al., 1986), but similar to the composition of biotite in the dacitic Fish Canyon Tuff (Whitney and Stormer, 1985). However, caution must be exercised in interpreting these results since K and Fe in biotite may be mobilized during alteration of volcanic ash beds (Blaylock, 1998), and all of the biotites have low totals, even with calculated water added in, probably as a result of alteration.

The F contents of the biotites is lower than those of biotites in the Fish Canyon Tuff, whereas Cl contents are generally similar to those in the Fish Canyon Tuff, except in ashes Gun-A, Gun-F, Gun-4, and MCH-2. Biotite in Gun-F has the highest Cl content compared to other samples. Halogen intercept values for biotites are similar to biotites from subduction-related intrusions associated with porphyry copper deposits (Munoz, 1984). Plotted on a $\log(X_{Mg}/X_{Fe})$ coordinate framework (where X is mole fraction; Ague and Brimhall, 1988), most biotites fall into the highly oxidized I-type fields, but scatter across the I-MC (moderately contaminated I-type granite) and I-SC (strongly contaminated I-type granite) fields (Fig. 5). Most biotite-bearing rocks in the Sierra Nevada batholith, including those of Jurassic age, are I-MC and I-SC types (Ague and Brimhall, 1988). None of our volcanic rocks had biotites like the mostly Cretaceous I-SCR (I-type, strongly contaminated and reduced) granites of the Sierra Nevada.

Among our samples, only Gun-B has both amphibole and biotite analytical data. In this ash, the partitioning of Mn between amphibole and biotite is consistent with an equilibrium distribution of Mn between the two minerals as estimated from compositions of various granitoid and volcanic rocks in the United States (Speer, 1984). Temperatures calculated from biotite compositions, by using an equation in Luhr et al. (1984, p. 93), ranged from 747 °C to 858 °C.

Pyroxene

Pyroxene occurs as a phenocryst in only a few ash beds. Orthopyroxene was found in four samples (Gun-B, Gun-C, Gun-D, and Gun-E) and clinopyroxene in the same four plus two others (Gun-AA and MWCB-8). Orthopyroxene compositions range from En_{63} to En_{70}, and all clinopyroxene is augite (Fig. 6). In most respects, the pyroxenes from these ash beds are like those in high-K to medium-K

TABLE 2. RELATIVE ABUNDANCES OF MAJOR PHENOCRYST PHASES IN ASH RESIDUES

Phenocryst mineral	MWW 45.4	Gun 4	Gun Aa	Gun A	Gun B	Gun C	Gun D	Gun E	Gun F	MCH 1	MCH 2	MCH 4	MCH 5	Hill 3815–2	Hill 3815–3	LVH 1	LVH 15	LVH 25	MWCB 8	MWCB 14	MWCB 16
Sanidine	c	*	c	A	c			A	c	A	A	A	A	c	A	c	A	A		A	c
Plagioclase	*	c																			
Biotite	A	A	A	A	A	*	*	A	c	A	A	A	c	c	c	c	A	A	*	A	c
Quartz	c	c	A	A	*			A	A	A	c	A	A	c	A	c	A	A		A	c

Note: A—abundant; c—common; *—rare.

TABLE 3. RELATIVE ABUNDANCES OF MINOR PHENOCRYST PHASES IN ASH RESIDUES

Phenocryst mineral	MWW 45.4	Gun 4	Gun Aa	Gun A	Gun B	Gun C	Gun D	Gun E	Gun F	MCH 1	MCH 2	MCH 4	MCH 5	Hill 3815–2	Hill 3815–3	LVH 1	LVH 15	LVH 25	MWCB 8	MWCB 14	MWCB 16
Apatite	*	c	*	c	c	*	*	c	A	c	c	c	c	*	c	*	c	c		*	*
Hornblende						*	*												*		
Pyroxene			*		*	c	*	*											*		
Titanite	c	*	*		c	*	*				*	*	c		c		*				
Zircon	*	c	c	A	c	*	*	c	c	c	c	c	c	*	*	*	c	c	*	*	*

Note: A—abundant; c—common; *—rare.

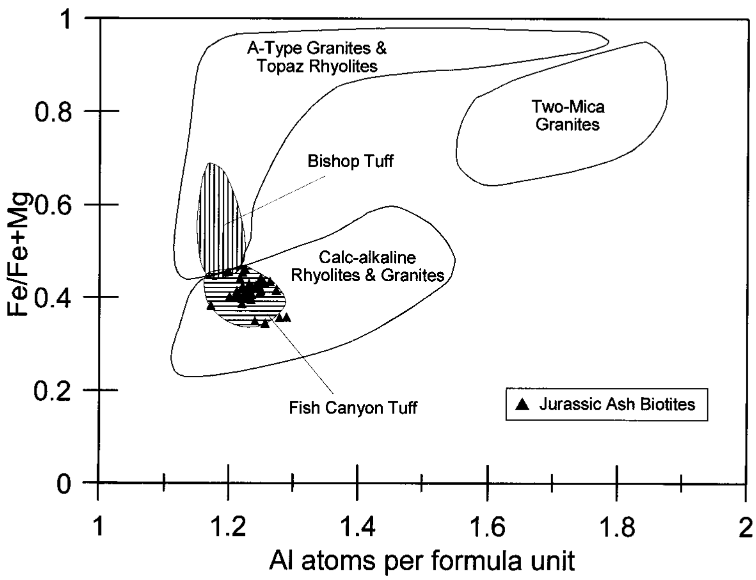

Figure 4. Compositions of biotite from Temple Cap and Carmel ash beds compared to Fish Canyon and Bishop Tuffs in terms of molar Fe/(Fe + Mg) and total Al relative to ideal end member. Fields for different types of granite are from Christiansen et al. (1986).

calc-alkaline andesites (Gill, 1981), dacites, and rhyolites (Ewart, 1979), but are different from those in rift-related, high-silica rhyolites like the Bishop Tuff (Hildreth, 1977), which are much richer in Fe (Table DR5; see footnote 1).

Augite and orthopyroxene appear to have been in equilibrium in Gun-D as judged by reasonable magmatic temperatures (~910 °C, calculated from average compositions using Lindsley and Andersen, 1983). In contrast, Gun-C has two populations of augite, $Wo_{45}En_{35}Fs_{21}$ and $Wo_{42}En_{45}Fs_{13}$. Augite in the second group appears to be in equilibrium with coexisting orthopyroxene and yielded a temperature of ~885 °C. Al, Na, Fe, Mn, and Ti contents are slightly higher, and Si content is slightly lower in the first group of augites from the Gun-C ash than in any other sample; perhaps these are xenocrysts. Likewise, orthopyroxene and clinopyroxene were apparently not in equilibrium in two sanidine-bearing ashes (see Fig. 6, Gun-B and Gun-E)—they yielded unreasonably high (1117 °C in Gun-B) and low (654 °C in Gun-E) temperatures. These pyroxenes may be xenocrystic as well. The other two pyroxene-bearing ashes with sanidine only have augite.

Amphibole

Amphibole is a trace constituent in a few of the altered ash beds. The compositions of the amphiboles in Gun-C and Gun-D, which lack sanidine and quartz, are distinctly different from those in the other two samples, which have sanidine, quartz, and biotite (Fig. 7). Core and rim analyses on amphibole grains do not show any significant zoning. Nonetheless, MWCB-8 has two amphibole populations. The first group is similar to those in Gun-B, and the second group is similar to those in Gun-D. Fe, Al, F, and Mn contents of the hastingsite in Gun-C are higher than those in the magnesiohornblendes in Gun-B and MWCB-8. In addition, Si and Mg contents are lower in amphiboles in Gun-C (Table DR6; see footnote 1).

The compositions of magnesiohornblendes in Gun-B and MWCB-8 are similar to those from the Fish Canyon Tuff (Whitney and Stormer, 1985) and to granites of the Sierra Nevada batholith, California (Dodge et al., 1969), although they are slightly more magnesian (Fig. 7). Gun-C, Gun-D, and some amphiboles from MWCB-8 are lower in Mg and Si content than those in the Fish Canyon and the Sierra Nevada batholith. All of the amphiboles probably indicate an association with subduction-related, calc-alkaline rocks similar to those found in other studies from the western United States (Dodge et al., 1969; Wender and Nash, 1979; Hausel and Nash, 1977; Keith, 1982; Christiansen et al., 1986) (Fig. 7).

Pressures were calculated for sample Gun-B, the only sample with the proper mineral buffer assemblage, by using the method of Johnson and Rutherford (1989). This method is based upon the partitioning of Al among hornblende, liquid, quartz, sanidine, plagioclase, and biotite. Calculated pressures range between 1.7 and 2.9 kilobars.

Apatite

Apatite occurs as euhedral prismatic crystals in all samples. Apatite compositions (Table DR7; see footnote 1) are similar to those from dacitic to rhyolitic volcanic rocks like the Fish Canyon Tuff (Whitney and Stormer, 1985). However, they are quite different from those of the highly evolved Bishop Tuff (Hildreth, 1977) with less Si and total REEs (rare earth elements) and more Ca and P (Fig. 8). Subduction-related dacitic and rhyolitic rocks often have high-Cl apatite (Stormer, 1972). The Cl concentrations in apatite (Fig. 9) from most samples in the Temple Cap, Carmel, and Page Sandstone Formations are relatively high and similar to apatite in the Fish Canyon Tuff (Whitney and Stormer, 1985) and other dacitic to rhyolitic volcanic rocks (Stormer and Carmichael, 1971; Stormer, 1972). A few samples (MCH-5 and Gun-Aa) have F contents similar to those in the highly evolved Bishop Tuff (Fig. 9) even though they differ from Bishop Tuff apatite in other elements (see Fig. 8). The

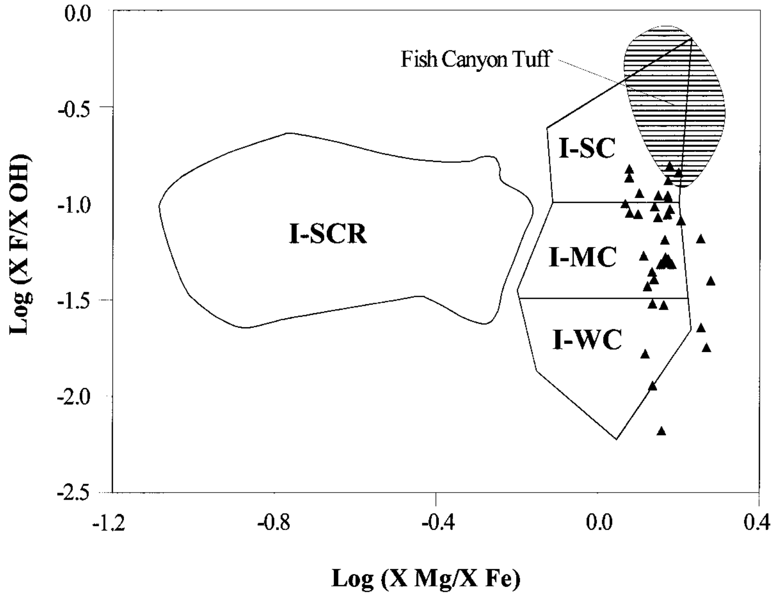

Figure 5. Compositions of biotites from Temple Cap and Carmel Formation ash beds compared to Fish Canyon Tuff in terms of log (X_{Mg}/X_{Fe}) vs. log (X_F/X_{OH}), where X is mole fraction. Granite fields from Ague and Brimhall (1988): I-SCR—I-type, strongly contaminated (by graphitic pelites) and reduced; I-SC—I-type, strongly contaminated; I-MC—I-type, moderately contaminated; I-WC—I-type, weakly contaminated.

Cl content of Gun-F is particularly high (about double that of our other samples), which suggests that apatite in Gun-F was formed in a very Cl rich environment. Biotite from Gun-F is also high in Cl, but apatite in Gun-F has a higher Cl content than the coexisting biotite.

Titanite

Titanite phenocrysts occur in most samples. No significant zoning of titanite was found (Table DR8; see footnote 1). Titanites in Gun-C and Gun-4 are different from those in other samples and those from the Fish Canyon Tuff (Whitney and Stormer, 1985). Ca content is higher, whereas Fe and REE contents are lower, in titanite in Gun-C and Gun-4 than in other samples (Fig. 10). Gun-4 also has a higher Ti content and lower Al content than other samples have.

Fe content generally increases with increasing Al content and with decreasing Ti content. Total REE content increases with decrease in Ca. These trends have also been observed in titanites from granodiorites (Sawka and Chappell, 1988; Paterson et al., 1989) and dacites (Nakada, 1991). Nakada (1991) concluded that Al (not Fe) substitutes mainly for Ti, whereas the REEs substitute for Ca. Many of the Jurassic ash beds have the mineral assem-

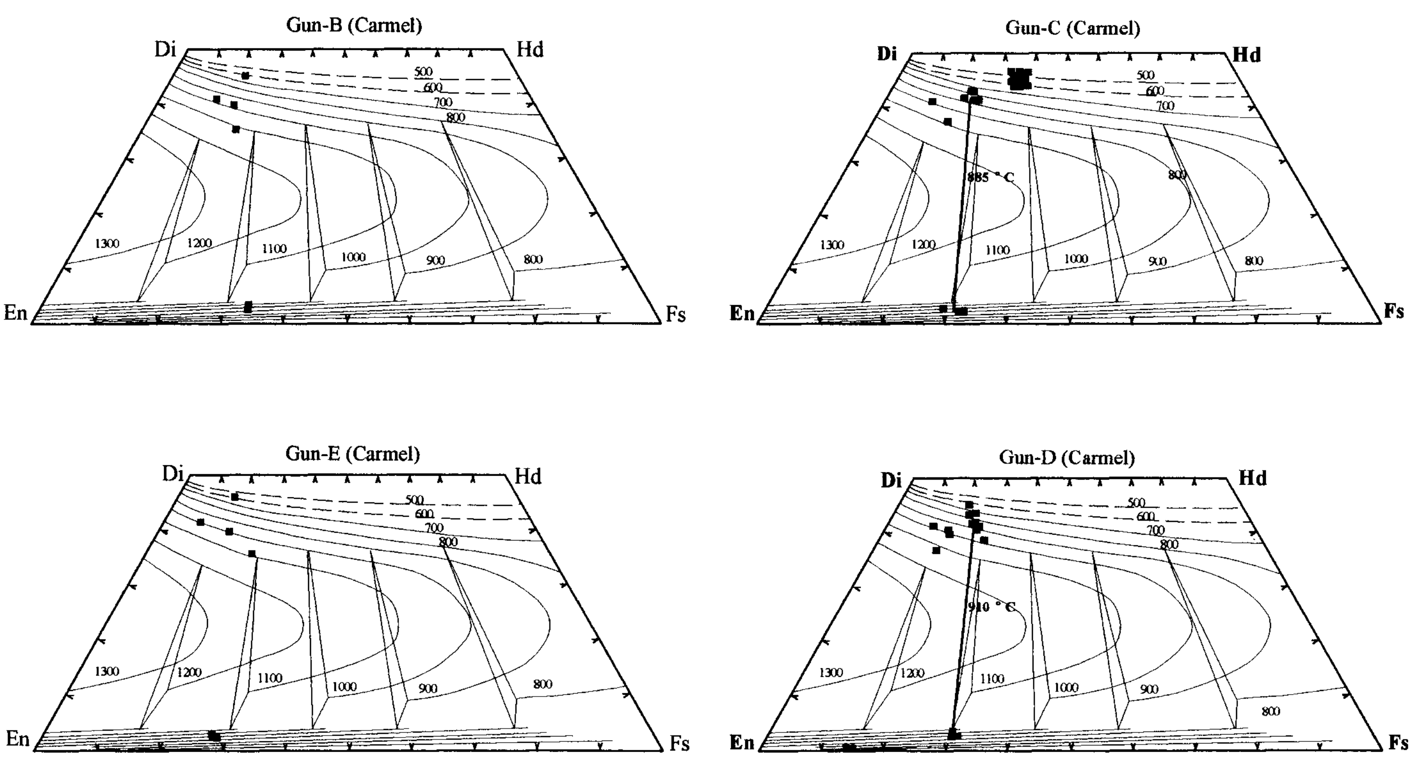

Figure 6. Pyroxene compositions from four Carmel Formation ash beds. Tie lines in Gun-C and Gun-D diagrams are proposed equilibrium orthopyroxene-clinopyroxene grains with appropriate temperatures. Isotherms in °C are from Lindsley and Andersen (1983).

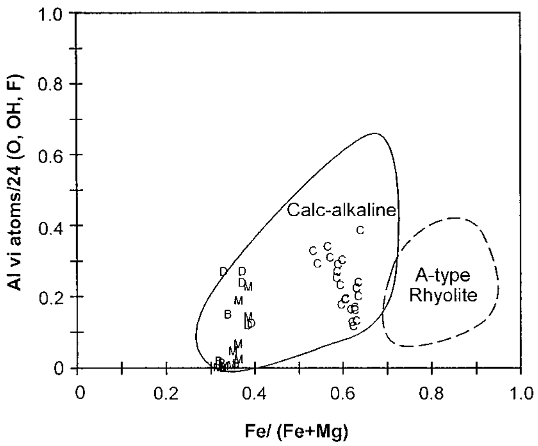

Figure 7. Compositions of amphibole in terms of molar Fe/(Fe + Mg) and octahedral Al relative to ideal end members. Fields from Christiansen et al. (1986). Sample symbols: B—Gun-B, C—Gun-C, D—Gun-D, M—MWCB-8.

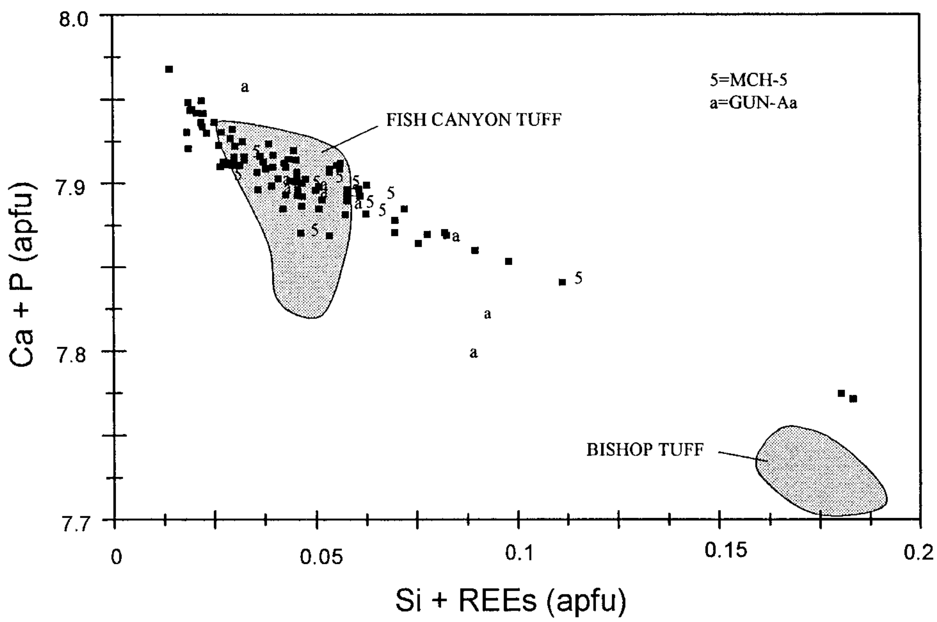

Figure 8. Diagram showing variation of (P + Ca) vs. (Si + REEs) in apatite from ash beds in the Temple Cap and Carmel Formations compared with apatite from the Fish Canyon Tuff (Whitney and Stormer, 1985) and the Bishop Tuff (Hildreth, 1977). REE analyses are not complete, but include La + Ce + Nd + Sm. Scatter of data from samples MCH-5 and Gun-Aa are shown; other samples have similar scatter.

blage titanite + magnetite + quartz, again similar to the Mesozoic batholiths of the Sierra Nevada batholith. Wones (1989) reported that this assemblage is indicative of relatively high oxygen fugacities. Although the presence of titanite and clinopyroxene alone may not indicate high oxygen fugacities (Xirouchakis and Lindsley, 1998), the combined assemblage of titanite with quartz, biotite, and hornblende has been shown by Carmichael (1991) to be a clear indicator of high oxygen fugacity.

Zircon Morphology

Studies by Pupin and his colleagues have suggested that the relative sizes of crystal faces in zircons have some relationship to the temperature and composition of the magma from which they crystallize (Pupin and Turco, 1972; Pupin et al., 1978; Pupin, 1980, 1985). More recently, Vavra (1990) and Benisek and Finger (1993) evaluated the internal crystal-growth patterns of zircon crystals and found that the growth rate and relative sizes of crystal faces were not in fact related to the temperature of the magma, but are caused by the amount of $ZrSiO_4$ supersaturation and by certain trace elements found in zircon crystals. Even with this modified understanding of zircon morphology, it is still possible to use the plots of different zircon morphologies as proposed by Pupin to identify likely rock types of the source. Figure 11 is a plot of the A and T Pupin indices (Pupin, 1980) showing the fields for crustal (mostly continental collision) granites, calc-alkaline (mostly subduction-related) granites, and alkalic (anorogenic) granites. Average zircon morphologies from the Middle Jurassic ash beds in this study lie within the calc-alkaline field and support the conclusions made in earlier sections of this paper that these ashes are from arc-derived, calc-alkaline magmas.

MAGMA COMPOSITION

The mineral contents of most of the ash beds show that they were rhyolitic. The presence of sanidine and quartz is diagnostic. A smaller group of ash beds was probably dacitic (e.g., Gun-4), lacking sanidine and having abundant amphibole. Three ash beds, interlayered with the more silicic ashes, have mineral compositions and mineral assemblages typical of andesites. These ash beds lack sanidine and quartz, have predominantly mafic minerals, including high-Al hornblende, biotite, and ortho- and clinopyroxene (Gun-C, Gun-D, and MWCB-8).

Volcanic ash beds, even when altered, have been shown to maintain a chemical signature that is useful in identifying the rock type and in correlating ash beds over wide areas (Huff, 1983; Kolata et al., 1986; Cullen-Lollis and Huff, 1986; Huff and Kolata, 1989; Kowallis et al., 1989). Immobile trace element characteristics for these ash layers in the Temple Cap and Carmel Formations are consistent with derivation from subduction-related (I-type) magmas and are unlike dacitic or rhyolitic magmas formed during continental collision or in ‘‘anorogenic’’ settings above mantle plumes or in continental rifts (Fig. 12). REE patterns (Fig. 13) show that some ash beds have deep Eu anomalies, consistent with derivation from highly fractionated rhyolitic magmas.

AGES, CORRELATIONS, AND TECTONISM

Temple Cap Formation

The Temple Cap Formation is ~120 m thick in southwestern Utah. New radiometric

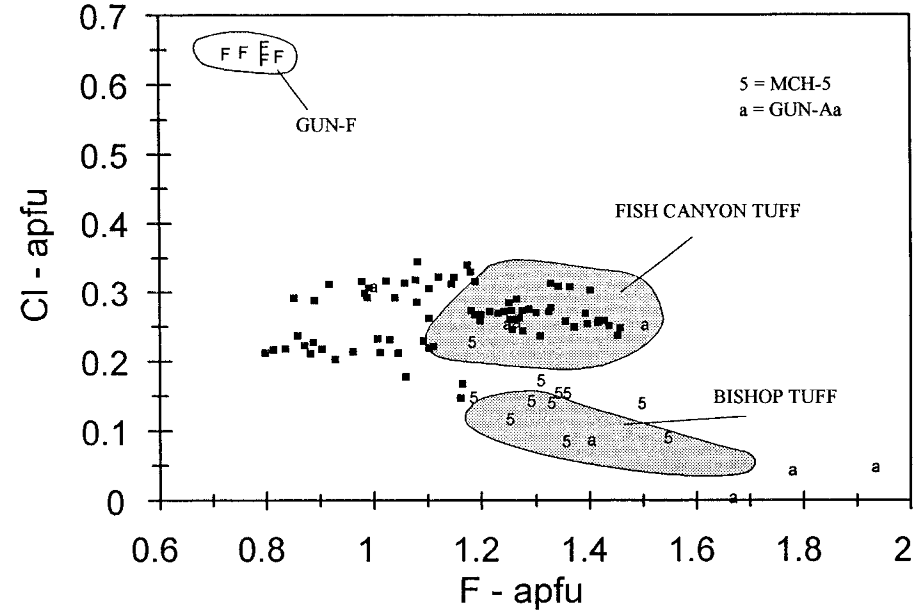

Figure 9. Cl vs. F variation in apatite from ash beds in the Temple Cap and Carmel Formations compared with apatite from the Fish Canyon Tuff (Whitney and Stormer, 1985) and the Bishop Tuff (Hildreth, 1977).

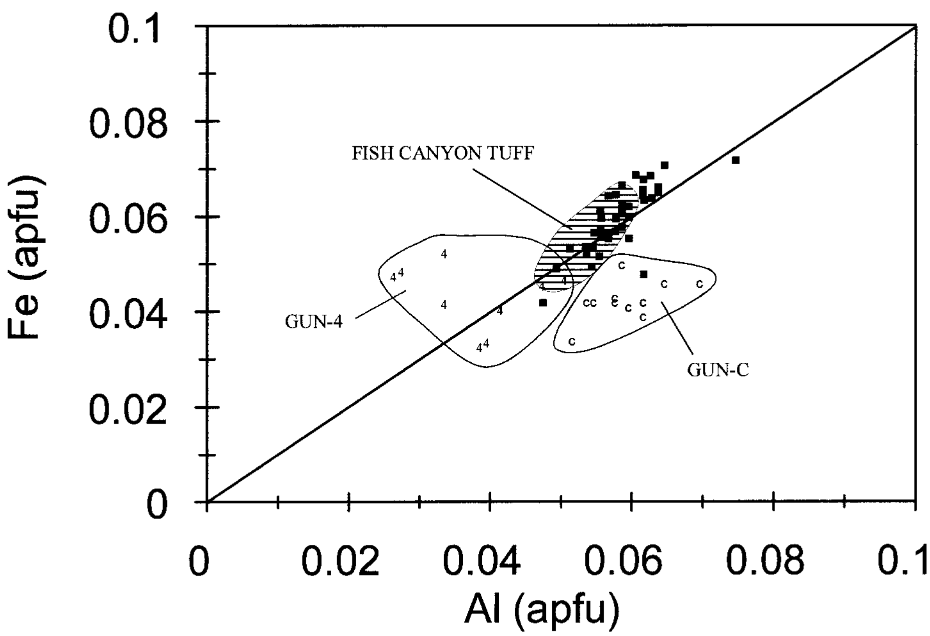

Figure 10. Plot of Fe vs. Al (in atoms per formula unit; apfu) for titanite from the Middle Jurassic ash beds compared to titanite from the Fish Canyon Tuff (Whitney and Stormer, 1985). Gun-C and Gun-4 are shown separately.

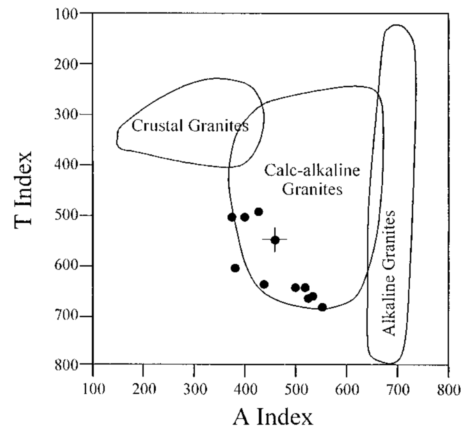

Figure 11. Average zircon morphology of samples from Middle Jurassic ash beds on plot showing source-rock types (developed by Pupin, 1980; see that reference for an explanation of the A and T indices).

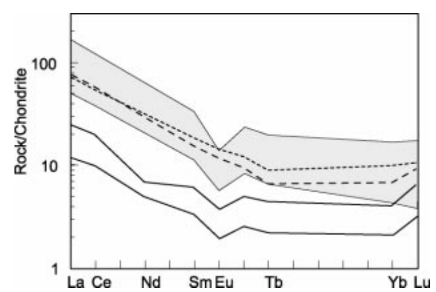

Figure 12. Chondrite-normalized rare earth element patterns for volcanic ash beds in the Carmel and Temple Cap Formations. Of the 13 samples analyzed, 9 have negative Eu anomalies and fall in the gray band. Two samples lack Eu anomalies and appear to be less evolved than the others. Two other samples (from the Mount Carmel Highway section) have low REE concentrations, probably because of dilution with sedimentary carbonate.

ages from the Temple Cap Formation include (see Fig. 2) 170.3 ± 0.6 Ma and 170.8 ± 0.6 Ma for samples MWCB-14 and MWCB-16 from an ash bed near the base of the formation ~4.5 m above the J-1 unconformity and the contact with the Navajo Sandstone (both samples come from the same 0.6-m-thick ash bed; MWCB-14 is from the base where phenocrysts were more concentrated, and MWCB-16 is from the top); 170.0 ± 0.6 Ma (LVH-25), 170.0 ± 0.7 Ma (Hill 3815–2), 169.9 ± 0.6 Ma (LVH-15), and 169.5 ± 0.5 Ma (Hill 3815–3), all from ~50 m below the top of the Temple Cap Formation at La Verkin Hill and Hill 3815 sections; and 169.3 ± 0.6 Ma (GUN-4) from ~24 m below the Temple Cap Formation–Carmel Formation boundary (i.e., the J-2 unconformity). It is probable that samples LVH-25 and Hill 3815-2 are from the same ash layer; samples LVH-15 and Hill 3815-3, which are stratigraphically above LVH-25 and Hill 3815-2, respectively, may also be from a single ash layer. These tentative correlations are based on the fact that the ash beds occupy the same stratigraphic position, have similar mineral types and abundances, and give ages that are statistically the same.

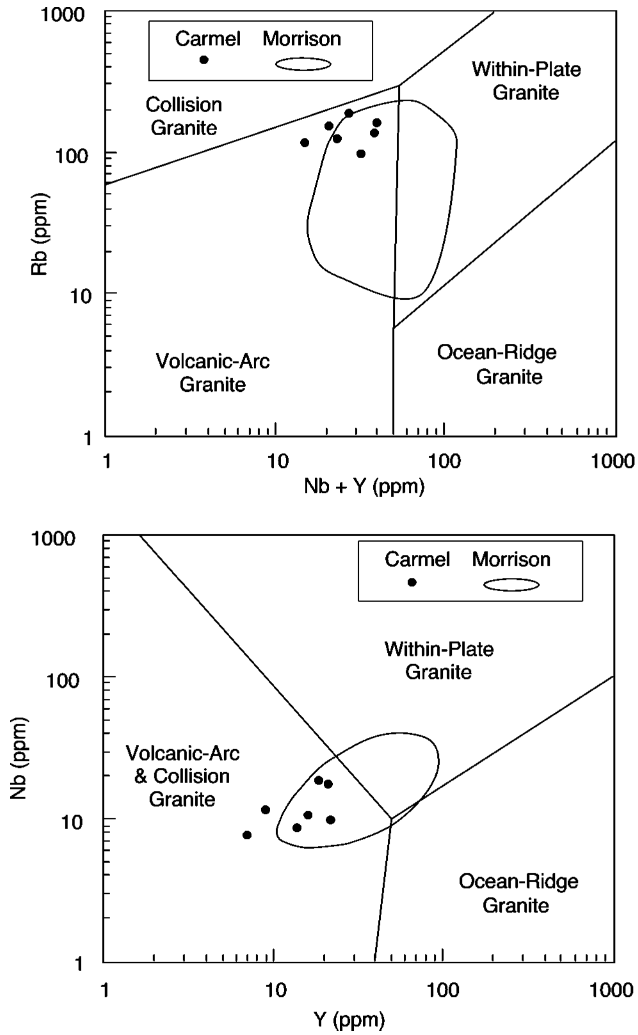

Figure 13. Trace element compositions of Temple Cap and Carmel ash beds fall in the volcanic arc field on the discriminant diagrams of Pearce et al. (1984).

The radiometric ages reported here for the Temple Cap Formation would place it in the middle Bajocian according to either the Harland et al. (1990) or Pálfy et al. (1998) Jurassic time scale. Included fossils are rare, and those found in this formation are not age diagnostic. The Temple Cap ash beds are the same age as ash-flow tuffs and lavas erupted in California in the Inyo and White Mountains (Hanson et al., 1987; Dunne and Walker, 1993; Dunne et al., 1998), the Palen Mountains (Fackler-Adams et al., 1997), the Cowhole Mountains (Wadsworth et al., 1995; C.J. Busby, E.R. Schermer, and J. Mattinson, 2000, written commun.), and the Central Mojave Desert (Graubard et al., 1988; E.R. Schermer and C.J. Busby, 2000, written commun.) (Fig. 14). They also are age equivalent to volcaniclastic rocks in the upper member of the Mount Wrightson Formation in the Santa Rita Mountains of southern Arizona (Riggs et al., 1993).

Carmel Formation

Radiometric ages from the Carmel Formation come from the two lowest members, the Co-op Creek and the Crystal Creek Members, and range from 168.2 $\pm$ 0.5 (MCH-1 collected ~5 m above the Temple Cap–Carmel contact) to 166.2 $\pm$ 0.6 (Gun-F) collected ~4 m below the contact of the Paria River and Crystal Creek Members of the Carmel Formation. The Co-op Creek and Crystal Creek Members are of middle to late Bajocian age on the basis of pelecypods and correlations to the ammonite-bearing Twin Creek Limestone in northern Utah (Imlay, 1964). A preliminary, unpublished (data in possession of Kowallis and Deino) laser-fusion single-crystal $^{40}Ar/^{39}Ar$ age of ca. 168 Ma on sanidine from an altered Twin Creek Limestone ash bed (collected from along the Weber River in northern Utah) solidifies the correlation between these formations. By using either the Harland et al. (1990) or Pálfy et al. (1998) time scale, the ages from the Carmel Formation agree with Imlay's work and place these two members in the upper Bajocian.

The volcanic ashes in the Carmel Formation are age equivalent to volcanic rocks in the central Mojave (E.R. Schermer and C.J. Busby, 2000, written commun.) and with the volcanic rocks of the McCoy Mountains Formation in the Palen Mountains of California (Fackler-Adams et al., 1997). Considering the rather large uncertainties in the ages of the volcanic sequences in California and Arizona (generally $\pm$3 to $\pm$8 m.y.), it is possible that the Carmel could also be time equivalent to the Jurassic volcanic rocks in the Inyo, White, and Cowhole Mountains (Fig. 14).

Age and Significance of the J-1 and J-2 Unconformities

One of our hopes in dating the ash beds of the Carmel and Temple Cap Formations was that we would be better able to determine the timing and duration of the J-1 and J-2 unconformities. To do so, ash samples were collected from just above the J-1 surface and bracketing the J-2 surface.

Samples MWCB-14 and MWCB-16 collected between 4 and 5 m above J-1 give an upper bound on the J-1 unconformity in southwestern Utah of ca. 170.5 Ma. The amount of time represented by the J-1 unconformity cannot be determined from our age data, but it appears that this surface represents a more substantial hiatus than the J-2 surface. Pipiringos and O'Sullivan (1978) proposed that the J-1 unconformity represented 2–3 m.y. If this were the case, then the upper Navajo Sandstone in southwestern Utah could be as young as 173–174 Ma. This age would make it Aalenian according to the Harland et al. (1990) time scale or lowermost Bajocian according to the better-defined Pálfy et al. (1998) Jurassic time scale. However, it is unclear how Pipiringos and O'Sullivan (1978) determined the length of this hiatus; it is possible that it represents a larger or smaller period of time.

Over much of the region where the J-2 is exposed, it can be identified by a lag of chert pebbles (O'Sullivan and Pipiringos, 1997). In southwestern Utah, this chert-pebble lag is not easily located and may be absent in some localities. Therefore, the identification of the J-2

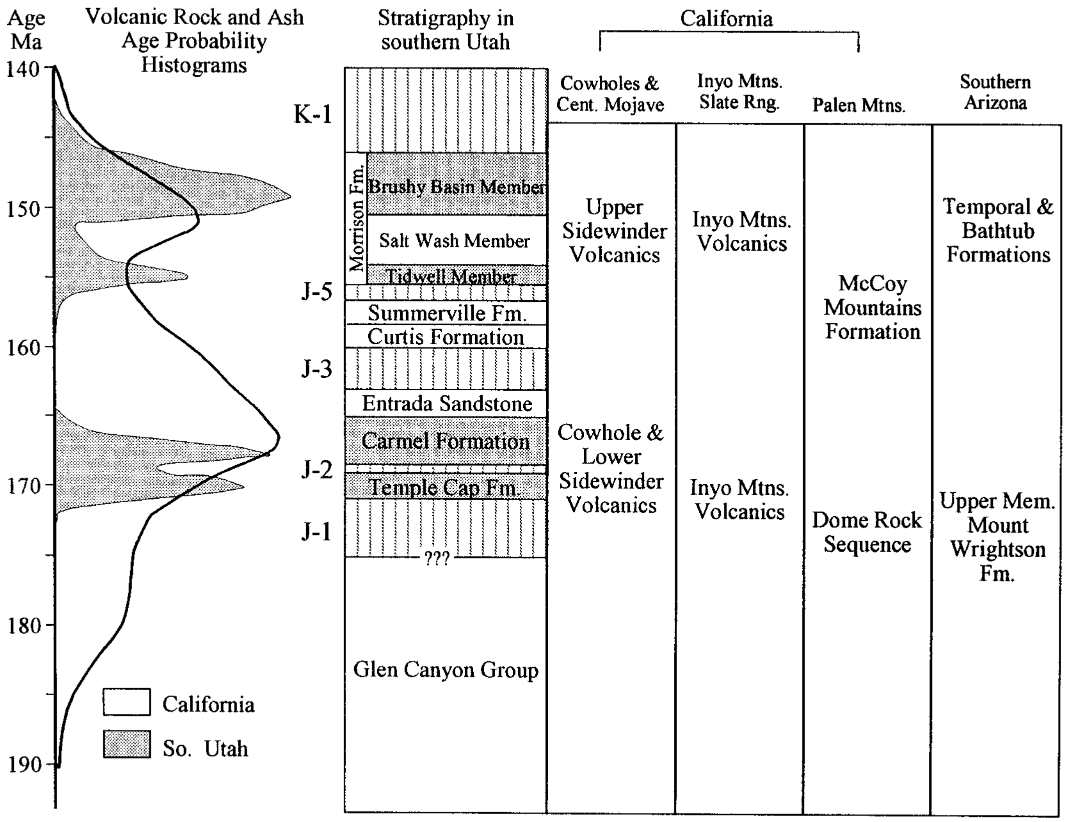

Figure 14. Correlation chart showing ages of formations in southwestern Utah compared with ages of volcanic rocks in California and Arizona. Also shown are regional unconformities and a relative probability histogram (after Hurford et al., 1984; Kowallis et al., 1986) for the ages of volcanic rocks from California and southern Utah.

unconformity can be problematic. F. Peterson (1993, personal commun.) identified the location of the J-2 unconformity in our measured stratigraphic sections on the basis of the occurrence of chert granules. However, the isotopic ages do not show a significant time gap at this level. His pick for the J-2 is shown in Figure 2A along with our preferred location for the J-2, which is based on a short break in the isotopic ages of ~1 m.y. between ca. 168 and 169 Ma.

The amount of time represented by the J-2 unconformity in southwestern Utah agrees with earlier observations that the J-2 unconformity, although apparently widespread throughout the western United States, must have formed in a relatively short period of time, perhaps 1 m.y. or less (Pipiringos and O'Sullivan, 1978; Peterson and Pipiringos, 1979). The J-2 hiatus increases eastward as expected for a eustatic change; however, Blakey (1994a) concluded that the J-2 surface is probably of both tectonic and eustatic origin.

Tectonic Synthesis

From the data presented here and our earlier work on similar altered ash beds in the Morrison Formation (Kowallis et al., 1991, 1998; Christiansen et al., 1994), it appears that two pulses of volcanic activity were recorded in the strata of southern Utah during the Middle and Late Jurassic. These same pulses of volcanic activity can also be identified in volcanic rocks found in the Jurassic arc region of California and Arizona (Fig. 14). Volcanic rocks of these same ages were probably also erupted from the arc in Nevada, but only plutonic rocks of this age occur in this region (Fig. 15). Ward (1995) proposed that several tectonic-magmatic cycles, separated by "major chaotic tectonic events," occurred along the western North American arc from Jurassic to the present. Ward's (1995) tectonic and magmatic cycles consist of (1) rapid subduction and associated trench-normal contraction with mostly small-volume, intermediate-composition (andesitic), arc-related volcanic eruptions, (2) slowing of subduction and a change to trench-normal extension with eruption of large-volume silicic tuffs and emplacement of major batholiths, (3) continued extension with associated major mylonitization in metamorphic core complexes and the formation of rifts, and finally (4) widespread strike-slip faulting and the return to rapid subduction. Lawton and McMillan (1999) presented a similar, three-phase tectonic model based on evidence from the Jurassic Mexican Borderland rift in southern Arizona and from the Cenozoic southern Rio Grande rift in New Mexico. If such tectonic cycles are real, it is unlikely that the small-volume, intermediate-composition eruptions of the first stage of these cycles would produce significant layers of air-fall ash in the Jurassic strata to the east. However, during the second stage of a cycle where trench-normal extension occurs, large-volume, caldera-forming eruptions of rhyolitic and dacitic ash could easily produce the ash layers preserved in the sections we studied.

The two periods of Jurassic volcanism, documented in the Middle Jurassic Carmel and Temple Cap Formations and in the apparently more extensive record found in the Upper Jurassic Morrison Formation, occur within one of the major periods of trench-normal extension identified by Ward (1995) along the western margin of North America between ca. 145 and 175 Ma. Moreover, the andesitic ashes that do occur are not found below the dacitic-rhyolitic ashes, but instead are interstratified with them (see Fig. 2). According to Ward's (1995) model, the ashes found in the Temple

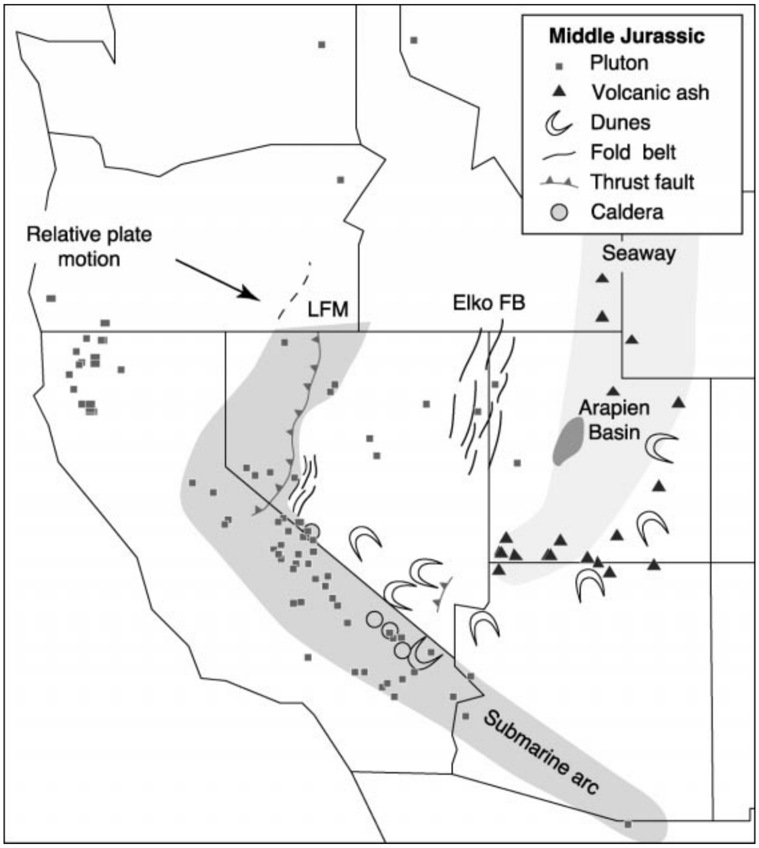

Figure 15. Geologic setting for the eruption and deposition of the Carmel Formation–Temple Cap Formation ash beds and the J-1 and J-2 unconformities. The ash beds accumulated in a shallow-marine to tidal-flat environment flanked on the east and west by dune fields. Middle Jurassic plutons and volcanic rocks outline a long, partly submarine magmatic arc that developed along the west coast of North America, probably as a result of oblique convergence and left-lateral strike-slip faulting in the arc. Plutons in northern California are allochthonous and were probably accreted to North America at a slightly later date. The northern part of the Middle Jurassic arc must have been removed by strike-slip motion during the Mesozoic or Cenozoic. The orientations of thrust faults and fold belts are consistent with southeast-directed contraction and with strike-slip movement in the arc. The J-1 and J-2 unconformities are contemporaneous and probably developed in response to far-field affects of these orogenic movements. LFM—Luning-Fencemaker thrust; FB—fold belt. Information used in compiling this figure comes from Busby-Spera (1988), Thorman et al. (1991), Dilek and Moores (1993), Taylor et al. (1993), Christiansen et al. (1994), Lawton (1994), Peterson (1994), DeCelles and Currie (1996), and Lawton and McMillan (1999).

Cap and Carmel Formations were erupted at the beginning of a period of extension, whereas the ashes from the Morrison Formation were erupted at the end. The cause of the two pulses of Jurassic volcanic activity is not obvious, but it is not simply a lack of preserved strata. The Entrada Sandstone and Curtis and Summerville Formations—which lie in between the ash-rich Middle and Upper Jurassic strata and had appropriate environments for preserving ash—are essentially ash free (see histogram in Fig. 14).

Implications for the Jurassic Time Scale

The radiometric ages reported here also have implications for the poorly defined Middle Jurassic geologic time scale. The few ages on which the time scale is based are mostly older K-Ar dates on minerals with low reliability (biotite, glauconite, and hornblende). Fossils in the Carmel Formation place it in the middle to late Bajocian (Imlay, 1964). The age range of ash beds in the Carmel Formation between 166.3 and 168.0 $\pm \sim$0.5 Ma is consistent with a Bajocian-Bathonian boundary of ca. 166 Ma as proposed by Harland et al. (1990) and Pálfy et al. (1998), but do not preclude the younger age of 164 Ma proposed by Odin (1994). However, these ages seem to eliminate the older boundary of ca. 169 Ma proposed by Gradstein et al. (1994) and the older 176 Ma boundary age proposed in the DNAG time scale (Palmer, 1983). The considerable differences between time scales for these boundaries demand that care be taken in mixing isotopic ages with biostratigraphic ages.

CONCLUSIONS

Altered volcanic ash beds in the Middle Jurassic Temple Cap and Carmel Formations of southwestern Utah are similar in age and composition to near-source, calc-alkaline, arc-related volcanic rocks located west and southwest of the study area in present-day California and Arizona. They are probably remnants of a low-lying, partly submarine arc formed by oblique convergence. Most of the magmas represented in the Temple Cap and Carmel Formations were originally dacite and low-silica rhyolite. The abundances of different types of phenocrysts, composition of phenocrysts, bulk-rock composition, trace element composition, and zircon morphology all support these conclusions. These ash beds are age equivalent with the Sidewinder volcanic series in the Central Mojave Desert, the Cowhole volcanics in the Cowhole Mountains, the McCoy Mountains Formation in the Palen Mountains, unnamed volcanic rocks in the Inyo and White Mountains, and the upper member of the Mount Wrightson Formation in Arizona. The radiometric ages also provide some constraints on the Middle Jurassic time scale. Middle to upper Bajocian ash beds in the Carmel Formation are most consistent with a Bajocian-Bathonian boundary of 166 Ma as proposed by Harland et al. (1990).

This Middle Jurassic pulse of magmatism is separated from a second, Late Jurassic pulse of magmatism by $\sim$10 m.y. During this hiatus in volcanism, sedimentation continued in depositional environments that could easily have accumulated volcanic ash, but none is found. This gap suggests that some fundamental change occurred in the arc region during this time interval—perhaps a reorientation of the direction or change in rate of subduction.

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