In situ U-Pb ages of zircons from the Bishop Tuff: No evidence for long crystal residence times

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ABSTRACT

Ion microprobe U-Pb isotope analyses of zircons from fallout and intercalated ash-flow deposits of the early, most chemically evolved Bishop Tuff (east-central California) yield a weighted mean age of 823 ± 11 ka (mean square of weighted deviates, MSWD = 0.8; n = 22). A similar mean age of 839 ± 36 ka (MSWD = 0.6, n = 15) obtained on unpolished zircon rims suggests that the difference between these ages and that of 760 ka for the Bishop Tuff obtained by recent Ar isotope determinations may be largely analytical in origin. Considerations based only on the effect of ^{230}Th-^{234}U disequilibria on the U-Pb ages yield an apparent maximum mean crystal residence time for zircon of ~100 k.y. The zircon residence times are significantly shorter than those inferred from previous estimates for the timing of phenocryst growth and suggest that, once evolved, the early Bishop Tuff rhyolite was not stored for long before erupting.

Keywords: Long Valley caldera, residence time, zircon, ^{238}U-^{206}Pb, ion probe data, rhyolites, magma chamber.

INTRODUCTION

Cataclysmic eruption of >750 km^3 (Hildreth and Mahood, 1986) of high-silica rhyolite gave rise to the Bishop Tuff and formation of the Long Valley caldera in east-central California (Fig. 1) ca. 760 ka (van den Bogaard and Schirnick, 1995, and references therein). Melt inclusion-bearing quartz crystals in some of the first-erupted portions of the Bishop Tuff yield radiogenic Ar ages (van den Bogaard and Schirnick, 1995) and Sr model ages (Christensen and Halliday, 1996) of 2.0–2.3 Ma and 1.3–2.5 Ma, respectively, suggesting that differentiation, accumulation, and storage of an appreciable portion of the magmatic system occurred long before eruption. The feldspars may be comparably old, which could indicate that the magma was suspended in a fairly narrow portion of the temperature interval between liquidus and solidus for a protracted period of time (Davies and Halliday, 1998, and references therein). Alternatively, the phenocrysts could be the more refractory residue of juvenile intrusions that remelted shortly before eruption (Sparks et al., 1990; Mahood, 1990). The thermal and mass requirements of these scenarios for accumulation of voluminous rhyolites are quite different and, for the Bishop Tuff, are critically dependent on the validity of the largely model phenocryst ages.

We have shown elsewhere that zircons in rhyolitic magmas may crystallize well before eruption (Reid et al., 1997). Given the foregoing results for the Bishop Tuff, a similar scenario would be anticipated for crystallization of its zircons. Differences between the trace element characteristics of various stages of the eruption (Michael, 1983), of melt inclusions in quartz and the whole rocks containing them (Dunbar and Hervig, 1992), and of various glass inclusions in quartz (Lu et al., 1992) suggest that zircon was a liquidus phase throughout the crystallization sequence. In support of this, zircons are present within all phenocryst phases (Hildreth, 1979) and within melt inclusions in quartz (Lu et al., 1992). High U concentrations (≥2500 ppm; Izett and Naeser, 1976) in zircons from the early, more evolved Bishop Tuff makes ^{238}U-^{206}Pb dating a promising means for confirming the age of phenocryst growth. Moreover, assuming that the differentiation process by which crystals and liquids separate is a stochastic one, the grain-by-grain distribution of ages obtained using the ion microprobe could more rigorously quantify the thermochemical evolution of the evolving magma system.

RESULTS

Spot analyses were performed on 80–200 mesh euhedral zircon grains from localities where melt inclusion-bearing quartz phenocrysts have been studied (e.g., van den Bogaard and Schirnick, 1995; Christensen and Halliday, 1996; Lu et al., 1992). One sample, 061498-24-6, consists of pumice lapilli from the Plinian fallout unit F6 (Wilson and Hildreth, 1998; Fig. 1). The other sample, 061498-16/17-p1, is a 15 cm pumice block from the distal portion of the Chidago lobe of the Gorges cooling unit (Ig1Eb; Wilson and Hildreth, 1998), where it is interstratified with, and therefore erupted contemporaneously with, the fallout. These portions of the deposit represent some of the most evolved rhyolite associated with the Bishop Tuff (Gardner et al.,

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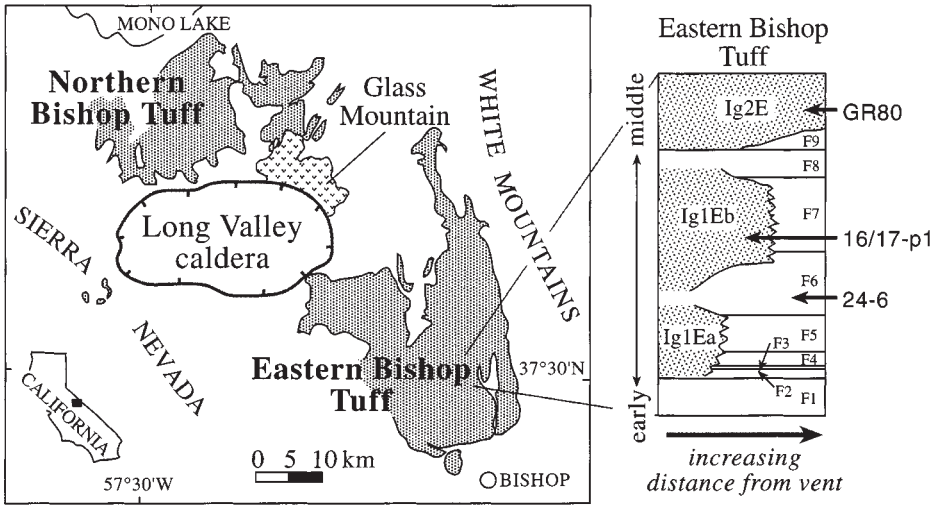

Figure 1. Map of Long Valley area and stratigraphic column for southeastern portion of Bishop Tuff (BT), showing sample locations and relative positions of samples with respect to eruption sequence. Modified after Wilson and Hildreth (1998); units in stratigraphic column with labels beginning in Ig are ash flows; units with labels beginning in f are fallout.

GSA Data Repository item 200048 contains additional material related to this article.

1991; Knesel and Davidson, 1997). No xenoliths were noted during crushing. Most zircons are 80–100 μm across and 150–200 μm in length. Many had glass adhering to them, suggesting that they were isolated grains rather than inclusions in other minerals. The majority of the analyses (24 spots on 21 grains) were performed on epoxy-mounted grains which were extensively polished in order to expose the interiors of the grains. We also analyzed the unpolished rims of several zircons (n = 15) from a welded portion (sample GR80) of the Tableland cooling unit (Ig2Eb; Wilson and Hildreth, 1998) of the middle Bishop Tuff. Analytical details and results for zircon interiors are presented in Table 1; results for zircon rims are presented in Table 2.[1]

One zircon from the fallout deposit gives discordant $^{206}Pb/^{238}U$ ages (n = 2) of comparable magnitude (ca. 210 Ma) to those obtained on zircons from Sierra Nevadan granitoids in the immediate vicinity of Long Valley caldera (Chen, 1982; Stern et al., 1981). The zircon xenocryst is likely derived from lithic fragments of the Triassic Wheeler Crest Quartz Monzonite (cf. Hildreth and Mahood, 1986), but may be the refractory residue of crustal anatexis.

[1]GSA Data Repository item 200048, Table 2, Results of U/Th/Pb ion microprobe analyses of Bishop Tuff zircon rims, is available on request from Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301-9140, editing@geosociety.org, or at www.geosociety.org/pubs/drpint.htm.

TABLE 1. RESULTS OF U/Th/Pb ION MICROPROBE ANALYSES OF BISHOP TUFF ZIRCON INTERIORS

Spot number	$f^{206}Pb^{*\dagger}$	7/6 corrected $^{206}Pb^*/^{238}U$ ($\times10^4$)§	7/6 corrected age (ka)#
Fallout (061498-24-6)			
r8g12s1	0.99	3610	229 ± 3 x10³
r8g12s2	0.99	3180	202 ± 2 x10³
r7g8s1	0.96	1.24	889 ± 64
r8g8s1	0.81	1.22	879 ± 52
r8g10s1	0.96	1.05	765 ± 44
r8g10s2	0.94	1.16	834 ± 41
r8g1s1	0.95	1.19	854 ± 46
r8g1s2	0.96	1.13	820 ± 42
r7g2s1	0.96	1.23	882 ± 45
r7g5s1	0.90	1.07	780 ± 56
r8g9s1	0.95	1.18	851 ± 47
r8g7s1	0.93	1.11	806 ± 54
r8g2s1	0.99	1.15	831 ± 45
r7g11s1	0.92	0.96	711 ± 47
r8g13s1	0.50	1.31	938 ± 197
Early ash flow (061498-16/17-p1)			
r1g11s1	0.97	1.16	835 ±54
r1g13s1	0.96	1.08	785 ±43
r1g1s1	0.87	1.09	790 ±59
r2g5s1	0.77	1.20	865 ±51
r1g9s1	0.98	1.20	862 ±36
r5g2s1	0.58	1.10	800 ±63
r1g4s1	0.94	1.09	794 ±50
r2g7s1	0.98	1.14	821 ±53
r1g3s1	0.92	1.08	787 ±57

Note: Analytical techniques as described in Dalrymple et al. (2000): 10-15 nA O^- primary beam; 3×10^{-5} torr O_2 flooding; spots ~35-45 μm in diameter; MRP~5500.

†Fraction of radiogenic ^{206}Pb based on common $^{207}Pb/^{206}Pb$ correction.

§Corrected for common Pb using $^{207}Pb/^{206}Pb = 0.818$ (Halliday et al., 1989).

#Corrected for Th/U fractionation assuming $Th/U_{magma} = 3.15$ (Mahood and Hildreth, 1983) and a weighted mean value of $Th/U_{zircon} = 0.57$ based on range of Th/U = 0.52 to 0.71 obtained from five spot analyses; see text for details.

The $^{207}Pb/^{206}Pb$ values obtained for the rest of the polished zircons generally cluster near the steady-state production value of 0.0462 (Fig. 2), showing that most of the ^{206}Pb in these zircons is radiogenic in origin (cf. Table 1). Paired $^{207}Pb/^{206}Pb$-$^{238}U/^{206}Pb$ characteristics of the early fall and flow zircons are similar, as anticipated from the stratigraphic proximity of the rocks they represent. Corrected for common $^{207}Pb/^{206}Pb$, the data cluster at a $^{206}Pb^*/^{238}U$ value which, at steady state, corresponds to an age of ca. 730 ka, but this apparent age neglects the effect of fractionation between daughter nuclides of the U-series decay chain that accompanied crystallization. The most abundant of the U-series daughter nuclides besides ^{234}U is ^{230}Th, and, relative to U, it is preferentially excluded during zircon crystallization. For the polished zircons, Th/U ratios were determined on a subset of the grains and these exhibit a limited range. Using the observed magnitude of zircon Th/U fractionation ($f = Th/U_{zir}/Th/U_{magma}$) to correct for initial ^{234}U-^{230}Th disequilibria in zircon (cf. Scharer, 1984), the weighted mean ages are 824 ± 14 ka (mean square of weighted deviates, MSWD = 1.1; n = 13) for the fallout zircons and 821 ± 17 ka (MSWD = 0.5; n = 9) for the flow zircons.

To better assess the accuracy of ion microprobe analyses on such young zircons, we carefully oriented unpolished zircon grains of GR80 and a standard (91500) flush to the surface of the ion microprobe mount in order to analyze the rims. These zircons, from a welded portion of the ignimbrite, were chosen for rim analyses because they are larger, have surfaces that appeared free of adhering glass, and have a greater probability of late-stage rim growth. The $^{207}Pb/^{206}Pb$ ratios obtained on the rims scatter more widely than those of the interiors (Fig. 2), reflecting the generally lower U concentrations and, in some cases, higher common Pb concentrations. These differences from the characteristics of the zircon interiors could be related to crystallization from a U-depleted boundary layer created as zircon grew while the melt quenched. For this set of analyses, acquisition of ^{208}Pb abundances permitted an alternate correction for common Pb using $^{208}Pb/^{206}Pb$. The two ages obtained in this fashion are in good agreement, indicating that any unresolved mass interferences at ^{207}Pb and ^{208}Pb, if present, are negligible. Corrected on a spot-by-spot basis for ^{230}Th fractionation, the rims yield a ^{207}Pb-corrected age of 839 ± 36 ka (MSWD = 0.6; n = 15) and ^{208}Pb-corrected age of 845 ± 36 ka (MSWD = 0.1). Both are essentially identical to those of the polished interiors of zircons from the early Bishop Tuff.

CRYSTAL RESIDENCE TIMES IN THE BISHOP TUFF

Recent Ar isotope estimates of 760 ± 1 ka for the age of the Bishop Tuff, based on analyses of sanidine (van den Bogaard and Schirnick, 1995, and references therein), may be somewhat young because of absolute age uncertainties in neutron-fluence monitors (Renne et al., 1998). Even so, a bias of at least ~1% between $^{40}Ar/^{39}Ar$ and ^{238}U-^{206}Pb ages between 28 and 4500 Ma, apparently due to uncertainties in ^{40}K decay constants and standard data (Renne et al., 1999), suggests that a U-Pb reference eruption age for the Bishop Tuff of at least 768 ka applies. The age of the zircon rims obtained from the ion microprobe analyses is somewhat older than this and may reflect uncertainties in the technique at this level of Pb isotope measurement. A relatively small mass interference at ^{206}Pb would, for example, be unrecognizable—and insignificant—in much older zircons. Preferential incorporation of intermediate U-series daughter nuclides other than ^{230}Th would also mean that the ages are overcorrected somewhat for disequilibria, although this effect is likely to be minor. Relative to most zircon ages, these uncertainties are quite small and further demonstrate the promise of the ion microprobe for dating even very young zircons (cf. Dalrymple et al., 2000).

On the basis of previous results, the mean age of zircon crystallization in the early Bishop Tuff would be expected to be at least as old as that of sanidine, which, based on Sr model ages, is estimated to >1 Ma (Davies and Halliday, 1998; data of Christensen and DePaolo, 1993). Instead, the interiors of early Bishop Tuff zircons yield a collective mean age of 823 ± 11 ka (MSWD = 0.8). The statistical fit of these data to a single age suggests that the distribution of ages exhibited by the zircons does not reflect a protracted episode of mineral growth. Considering also the similarity of this age to that

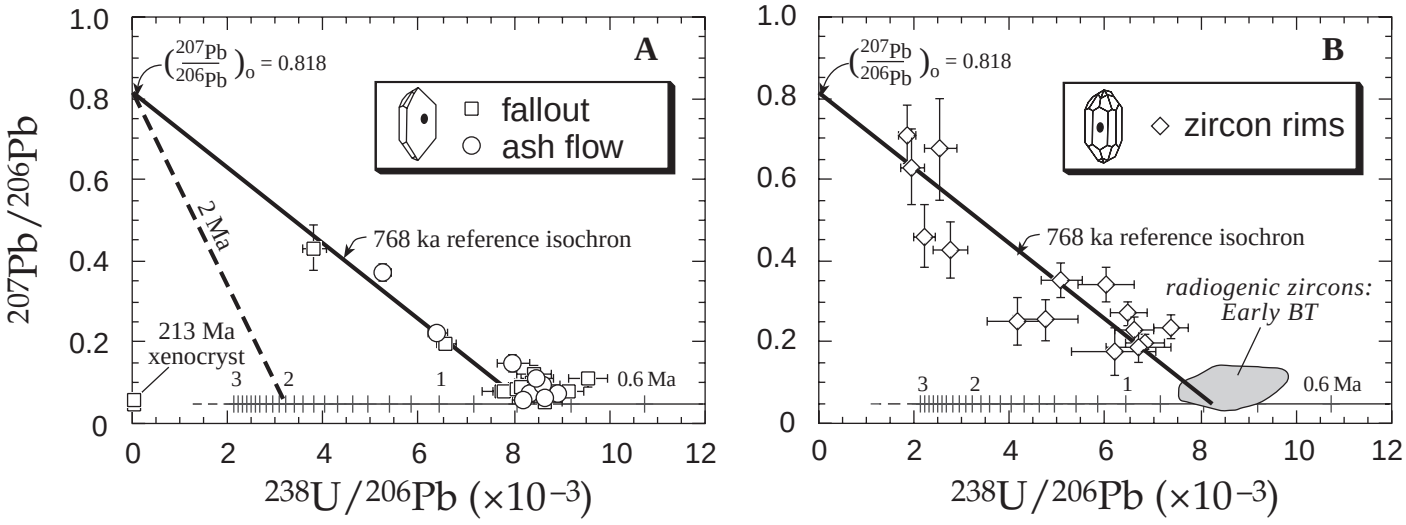

Figure 2. $^{207}Pb/^{206}Pb$ vs. $^{238}U/^{206}Pb$ diagrams illustrating results for (A) zircon interiors, early Bishop Tuff fallout and flow deposits (samples 061498-24-6 and 061498-16/17-p1, respectively) and (B) zircon rims, middle Bishop Tuff (BT) (sample GR80). Values for concordia illustrated for ages <3 Ma, assuming steady-state $^{207}Pb/^{206}Pb$ production.

obtained on the zircon rims, the most straightforward interpretation of our results is that the zircons grew within a few tens of thousands of years of eruption. The ages reported in Table 1 are, however, necessarily minimum ages because they assume that the zircon and magma have not continued to equilibrate with respect to the U-series nuclides in the time interval, if any, between crystal growth and eruption. Of nuclides in the U-decay chain, U, Th, and Pb, and therefore probably also Ra, are unlikely to exchange significantly between zircon and magma, even at relevant magmatic temperatures of 800 °C (Lee et al., 1997; Cherniak et al., 1997; Reid et al., 1997). Loss of ^{222}Rn ($t_{1/2}$ ~4 days), for which diffusive rates in zircons at magma conditions have not been quantified, might be capable of limiting ^{206}Pb accumulation. Fortunately, we can independently constrain maximum magmatic residence times for most of the zircons from considerations based on ^{234}U-^{230}Th disequilibria alone. Uncorrected for U-series disequilibria, the mean $^{206}Pb/^{238}U$ age is younger than the age of eruption. Rather than being an analytical artifact, this difference probably reflects suppressed accumulation of ^{206}Pb while ^{230}Th returned to secular equilibrium with ^{234}U: Had crystallization preceded eruption by a long period of time (more than hundreds of thousands of years), uncorrected zircon ages would have been at least as old as those of eruption. Thus, the difference between the uncorrected zircon ages and that of eruption is a measure of the magnitude of ^{234}U-^{230}Th disequilibria at the time of eruption.

Maximum crystallization ages for the zircons that incorporate the constraint imposed by ^{230}Th-^{234}U disequilibria can be obtained by assuming no accumulation of radiogenic Pb until the time (t) of eruption:

$$\text{maximum } t_{\text{crystallization}} = t_{\text{eruption}} - 1/\lambda_{230} \ln\{\lambda_{230}/\lambda_{238}/[f-1]\cdot[\Delta(^{206}Pb^*/^{238}U)_{\text{meas.-erupt.}}]\}, \quad (1)$$

where $\Delta(^{206}Pb^*/^{238}U)_{\text{meas.-erupt}}$ is the difference between $^{206}Pb^*/^{238}U$ measured and that for steady-state ingrowth of ^{206}Pb since eruption and f is the magnitude of ^{234}U-^{230}Th fractionation at the time of crystallization, as defined here. For the limiting case of an eruption age of 768 ka and an average measured $^{206}Pb^*/^{238}U$ value of 1.14×10^{-4} for the zircon interiors, the calculated maximum mean age of crystallization would be 870 ka and the mean crystal residence time would therefore be ~100 k.y. This interpretation assumes a temporally restricted episode of zircon growth. Because the maximum crystallization age is not linearly dependent on the difference between the eruption age and measured $^{206}Pb^*/^{238}U$, we cannot preclude a priori the possibility that those few zircons that yield ages older than ca. 860 ka have significantly greater residence times than the mean. If, as results for the zircon rims would seem to indicate, the ion microprobe ages are systematically too old, then even these zircons probably crystallized no more than ~200 k.y. before eruption, and this early only in the very unlikely event that preeruption ^{206}Pb accumulation was negligible.

Even though the duration of zircon growth may have been somewhat more protracted than indicated by the U/Pb ages, it appears that either most of the zircons in the early Bishop Tuff grew much later than the major phenocryst phases or that the (largely model) ages obtained for those phenocrysts do not accurately reflect crystal residence times. Either way, the proportion of older ages is significantly less than what would be expected for origin of the Bishop Tuff by repeated intrusion, solidification, and remelting of silicic protoliths (Huppert and Sparks, 1988; Sparks et al., 1990) or by remelting of a crystalline mush (Mahood, 1990). For this reason, we interpret the significance of the zircon ages in the context of a system that is molten throughout its evolution. Both zircon and other phenocryst phases in the early Bishop Tuff grew from a very differentiated magma, as evidenced by the significantly higher U contents of zircon and of melt inclusions in quartz, and the higher Rb/Sr of sanidine and melt inclusion-bearing quartz compared to that of the same phases later in the eruption (data of Izett and Naeser, 1976; Lu et al., 1992; Christensen and DePaolo, 1993; Christensen and Halliday, 1996). Of particular note is evidence that Zr concentrations in melt inclusions in quartz are lower than those of the whole rocks in which the quartz is contained (Dunbar and Hervig, 1992). For a closed system, this would require that the zircon phenocrysts grew before or at the same time as quartz. Taken together with the observation that the gravitational buoyancy of the relatively small zircons is similar to or significantly greater than that of the large (>1 mm) quartz crystals that contain radiogenic isotope signatures, this suggests that, in a system that has evolved by internal differentiation, old zircons should be just as common as old quartz, and our data indicate that they are not.

Whereas phenocryst growth occurred after differentiation, it mostly preceded the development of isotopic heterogeneity in the melt. Feldspars from throughout the Bishop Tuff generally exhibit a relatively limited range of initial Sr isotopic characteristics (0.7060–0.7067; Christensen and DePaolo, 1993; Christensen and Halliday, 1996), although individual crystals from the early Bishop Tuff yield initial ratios (0.7078–0.7131; Davies and Halliday, 1998) that are similar to the highly variable initial Sr isotope characteristics of associated melt-bearing quartz (0.7065–0.7115; Christensen and Halliday, 1996). For evolution in a closed system, the

phase-dependent isotopic differences would be the product of in situ decay and the >1 Ma Sr model ages for feldspars and melt inclusion-bearing quartz would accurately reflect the mean ages of crystallization. In evident support of this scenario, Ar apparent and isochron ages for melt inclusion-bearing quartz are broadly similar to the Sr model ages. Experimental determination of diffusion parameters obtained by fractional release of ^{39}Ar during step heating of melt inclusion-bearing quartz suggests, however, that Ar could not have been quantitatively retained by those quartz for periods of >1 k.y. at magmatic conditions (Boyce et al., 1999), and raises doubts about the relationship between the Ar ages and the duration of magmatic differentiation and storage. In addition, glass inclusions within quartz phenocrysts from the Bandelier Tuff are, just as for the Bishop Tuff, more radiogenic with respect to Sr than interstitial glass, and the difference there is probably due to the localized effects of crustal contamination rather than in situ aging (Wolff et al., 1999). Given the hydrous nature of their melt inclusions (Anderson et al., 1989; Dunbar and Hervig, 1992), the radiogenic quartz phenocrysts in the Bishop Tuff may, by analogy, have grown at temperatures close to that of solidification (e.g., Naney, 1983) near the margins of a chemically stratified rhyolite body, where contamination may be more prevalent (cf. Duffield et al., 1995). Feldspars that likely cocrystallized with quartz are relatively Sr poor (Davies and Halliday, 1998) and apparently constitute a minor proportion of the feldspar Sr isotope budget. Most feldspar growth occurred in the main magma body, in the more than 100 °C temperature interval before the appearance of quartz (a smaller temperature interval likely applied to the early Bishop Tuff rhyolite specifically). In that way, the thermal and material balances of crystallization and assimilation were partitioned so that, although coupled thermally, the chemical and therefore isotopic effects of assimilation were imparted locally. Mixing between the marginal and main magma domains yielded a complex mineral assemblage and a matrix that is both heterogeneous and intermediate in isotopic composition and that gives rise to misleading Sr (and Ar?) model ages. Accordingly, rather than a protracted magmatic residence time, the early Bishop Tuff rhyolite may have been stored for only a few tens of thousands of years before eruption.

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