Quaternary rift flank uplift of the Peninsular Ranges in Baja and southern California by removal of mantle lithosphere

Karl Mueller,1 Grant Kier,1 Thomas Rockwell,2 and Craig H. Jones1,3

Received 2 November 2007; revised 13 January 2009; accepted 12 May 2009; published 9 September 2009.

[1] Regional uplift in southern California, USA, and northern Baja California, Mexico, is interpreted to result from flexure of the elastic lithosphere driven largely by heating and thinning of the upper mantle beneath the Gulf of California and eastern Peninsular Ranges. The geometry and timing of faulting in the Salton Trough and Gulf of California, the history of recent rock uplift along the Pacific coastline, and geophysical data constrain models of lithospheric heating and thinning based on unloading of a continuous elastic plate. High topography that marks the ~400-km-long rift shoulder in northern Baja California mimics the pattern of uplift observed along the Pacific coastline as defined by marine terraces. We interpret this to indicate that recent rock uplift has occurred across the entire width of northern Baja Peninsula and increases from west to east. Pliocene strata deposited at sea level along the Pacific coastline in southern California have not been uplifted significantly above Quaternary marine terrace deposits. This suggests the onset of rock uplift along the Pacific coast here is post-Pliocene and occurs after Miocene crustal extension in the Salton Trough and Gulf of California. Heating of the mantle lid beneath the Peninsular Ranges in northern Baja California thus coincides with crustal extension limited to localized oceanic spreading in the Gulf of California. Citation: Mueller, K., G. Kier, T. Rockwell, and C. H. Jones (2009), Quaternary rift flank uplift of the Peninsular Ranges in Baja and southern California by removal of mantle lithosphere, Tectonics, 28, TC5003, doi:10.1029/2007TC002227.

1. Introduction

[2] Quaternary uplift of coastal southern California and northern Baja California has long been recognized by the presence of well-preserved flights of marine terraces along the Pacific coast [Arnold, 1903; Kennedy et al., 1982; Kern and Rockwell, 1992; Muhs et al., 2002]. The cause of uplift, however, has not been studied in detail, nor has it appeared particularly significant until the recognition of active blind thrust faults in offshore regions of the southern California borderland by Rivero et al. [2000]. They attribute the observed coastal uplift in southern California to slip on a blind thrust system that includes one segment (the Ocean-side detachment) extending down-dip beneath the coastline, implying significant seismic hazard for this region. In contrast, Johnson et al. [1976], Muhs et al. [1992] and Orme [1998] have argued that regional uplift in coastal southern California and northern Baja California is due to aseismic tectonic or epirogenic processes Table 1.

[3] In this paper, we explore whether, and to what extent, Miocene to Recent thinning and/or heating of the lithosphere beneath the Gulf of California and Salton Trough is likely to have produced rock uplift similar to that observed in the adjacent Peninsular Ranges. We first test whether rifting of a thin elastic plate as constrained by present geological and geophysical conditions in the Gulf of California can accurately predict observed topography. Second, we determine whether modeled rift shoulder uplift produces a signal of total rock uplift along the Pacific coastline that is consistent with observed uplift of Quaternary marine terraces. We then compare the timing of coastal uplift with the history and style of extension in the Salton Trough and Gulf and California and its implications for processes that drive Quaternary rock uplift in the Peninsular Ranges.

2. Geologic Setting and the Gulf of California

[4] The Peninsular Ranges of southern California and northern Baja California form the high-relief, western boundary escarpment of the northern Gulf of California and Salton Trough. This region, termed the Gulf Extensional Province by Gastil et al. [1975] is characterized by low topography [Larsen and Reilinger, 1991] (Figure 1a) that can be related to a history of extensional and transform faulting that began in the Miocene [Stock and Lee, 1994; Axen and Fletcher, 1998; Axen et al., 2000; Oskin et al., 2001; Oskin and Stock, 2003].

[5] The Gulf of California rift system first developed as the Pacific-Farallon spreading ridge neared North America at the end of the Middle Miocene [Atwater, 1970; Lonsdale, 1989; Stock and Hodges, 1989]. Following the cessation of spreading between ca. 12.5 and 14 Ma along the Pacific coast of Baja California, subsequent divergence of the Pacific Plate from North America was accommodated by extension of continental crust farther east occurring as slip
Table 1. Locations and Elevations of Marine Terraces Measured Along the Pacific Coastline\(^a\)

<table>
<thead>
<tr>
<th>Terrace Locality</th>
<th>Location Number</th>
<th>Latitude (deg)</th>
<th>Longitude (deg)</th>
<th>(5a) Elevation (m)</th>
<th>(5e) Elevation (m)</th>
<th>Uplift Rate (mm/a)</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>San Joaquin Hills</td>
<td>1</td>
<td>33.61</td>
<td>117.92</td>
<td>19</td>
<td>32</td>
<td>0.21–0.24</td>
<td>Grant et al. [1999]</td>
</tr>
<tr>
<td>Oceanside</td>
<td>2</td>
<td>33.35</td>
<td>117.52</td>
<td>9</td>
<td>22</td>
<td>0.13</td>
<td>Kern and Rockwell [1992]</td>
</tr>
<tr>
<td>North San Diego County</td>
<td>3</td>
<td>33.17</td>
<td>117.35</td>
<td>9</td>
<td>22</td>
<td>0.13</td>
<td>Kern and Rockwell [1992]</td>
</tr>
<tr>
<td>San Diego (outside fault zone)</td>
<td>4</td>
<td>32.91</td>
<td>117.24</td>
<td>9</td>
<td>22</td>
<td>0.13</td>
<td>Kern and Rockwell [1992]</td>
</tr>
<tr>
<td>Mt. Soledad</td>
<td>5</td>
<td>32.85</td>
<td>117.27</td>
<td>14</td>
<td>22</td>
<td>0.13</td>
<td>Kern and Rockwell [1992]</td>
</tr>
<tr>
<td>Point Loma</td>
<td>6</td>
<td>32.69</td>
<td>117.25</td>
<td>9–10</td>
<td>22</td>
<td>0.13</td>
<td>Kern and Rockwell [1992], Ku and Kern [1974], and Muhs et al. [1992]</td>
</tr>
<tr>
<td></td>
<td>7</td>
<td>32.5</td>
<td>117.15</td>
<td>20–23</td>
<td>~0.13</td>
<td></td>
<td>Valentine and Rowland [1969]</td>
</tr>
<tr>
<td>Tijuana Playa</td>
<td>8</td>
<td>32.3</td>
<td>117.08</td>
<td>23</td>
<td>0.14</td>
<td></td>
<td>Valentine [1957] and Valentine and Rowland [1969]</td>
</tr>
<tr>
<td>Punta Descanso</td>
<td>9</td>
<td>32.24</td>
<td>117.03</td>
<td>23</td>
<td>0.14</td>
<td></td>
<td>Kennedy et al. [1986]</td>
</tr>
<tr>
<td>Alisitos/La Fonda</td>
<td>10</td>
<td>32.1</td>
<td>116.91</td>
<td>27</td>
<td>0.17</td>
<td></td>
<td>(SLA in fault zone)</td>
</tr>
<tr>
<td>Ensenada</td>
<td>11</td>
<td>31.85</td>
<td>116.65</td>
<td>~20</td>
<td>~0.12</td>
<td></td>
<td>Kennedy and Rockwell (unpublished data, 1995)</td>
</tr>
<tr>
<td>Punta Banda</td>
<td>12</td>
<td>31.73</td>
<td>116.77</td>
<td>27–43</td>
<td>0.16–0.29</td>
<td></td>
<td>Rockwell et al. [1989]</td>
</tr>
<tr>
<td>South Maximinos</td>
<td>13</td>
<td>31.65</td>
<td>116.7</td>
<td>29–30</td>
<td>0.18</td>
<td></td>
<td>Rockwell et al. [1989] and Kennedy et al. [1986]</td>
</tr>
<tr>
<td>Punta Santo Tomas</td>
<td>14</td>
<td>31.54</td>
<td>116.72</td>
<td>8–9</td>
<td>0.13</td>
<td></td>
<td>Emerson [1956]</td>
</tr>
<tr>
<td>Punta Cabras</td>
<td>15</td>
<td>31.3</td>
<td>116.48</td>
<td>6</td>
<td>0.08–0.10</td>
<td></td>
<td>Addicott and Emerson [1959] and G. L. Kennedy (personal communication, 2004)</td>
</tr>
<tr>
<td>Punta Baja</td>
<td>16</td>
<td>29.94</td>
<td>115.81</td>
<td>8–10</td>
<td>&gt;10</td>
<td>0.13</td>
<td>Emerson and Addicott [1958], Otltie et al. [1984], and Kennedy (unpublished data, 1995)</td>
</tr>
<tr>
<td>Isla de Guadalupe</td>
<td>17</td>
<td>28.95</td>
<td>118.14</td>
<td>6</td>
<td>0</td>
<td></td>
<td>Lindberg et al. [1980]</td>
</tr>
<tr>
<td>Punta Rosallita</td>
<td>18</td>
<td>28.66</td>
<td>114.27</td>
<td>n.o.</td>
<td>6</td>
<td>0</td>
<td>Emerson and Hertlein [1960] and Kennedy and Rockwell (unpublished data, 1995)</td>
</tr>
<tr>
<td>Turtle Bay, Viscaino Peninsula</td>
<td>19</td>
<td>27.67</td>
<td>114.88</td>
<td>12</td>
<td>24–27</td>
<td>0.15–0.16</td>
<td>Emerson [1980], Emerson et al. [1981], and Otltieb et al. [1984]</td>
</tr>
<tr>
<td>Mulege (On Gulf of California)</td>
<td>20</td>
<td>26.9</td>
<td>112</td>
<td>n.o.</td>
<td>12</td>
<td>0.04–0.05</td>
<td>Ashby et al. [1987]</td>
</tr>
<tr>
<td>Cabo San Lucas</td>
<td>21</td>
<td>22.85</td>
<td>109.9</td>
<td>n.o.</td>
<td>6</td>
<td>0</td>
<td>Muhs et al. [1992]</td>
</tr>
</tbody>
</table>

\(\text{\textsuperscript{a}}\)Locations of measurements along coastline shown in Figure 3 by location number. Stage 5a is 83 ka, and stage 5e is 122 ka. Here n.o., not observed.

\(\text{\textsuperscript{b}}\)Early extensional strain in the Salton Trough at the northern head of the Gulf of California was accommodated by an east rooted system of detachment faults that have been subsequently dismembered by NW trending dextral faults along the western margin of the Salton Trough (Figure 1b [Axen, 1995; Axen and Fletcher, 1998]). The age of syntectonic alluvial strata preserved in the hanging walls of these low-angle normal faults suggest they were active from late Miocene through Pliocene time (beginning ca. 5–8.3 Ma [Axen and Fletcher, 1998; Axen et al., 2000; Dorsey et al., 2007]). Early extension in the Salton Trough is also defined in northernmost Baja California by a separate array of low-angle normal faults rooted to the west beneath the Sierra Juarez [Axen, 1995; Axen et al., 2000]. The cooling history of rocks exhumed in the footwalls of these detachments suggest they were active between ~4–15 Ma during early rift-related extension [Axen et al., 2000]. Evidence for modern extension along this fault system has also been suggested on the basis of offset Quaternary sediments deposited along the western margin of the Sierra Mayor [Axen et al., 1999].

\(\text{\textsuperscript{c}}\)Quaternary strain in eastern Baja California, the Salton Trough and Gulf of California is dominated by dextral shear across the Pacific–North America plate boundary. This is manifest as dextral transform faults in the Gulf of California, and a broad zone of right lateral faulting that extends across the Peninsular Ranges and Salton Trough in southern California and northern Baja California. Faults that accommodate dextral Quaternary plate motions in the Salton Trough include the active NW trending San Jacinto, Elsinore–Laguna Salada and San Andreas/Imperial/Cerro Prieto fault systems; the latter are interpreted to be linked across the Brawley and Mexicali/Cerro Prieto seismic zones [Fuis et al., 1984; Lonsdale, 1989] (Figure 1b). These dextral fault systems also bound actively extending basins such as the Imperial and Mexicali Valleys that are filled with thick successions of Pleistocene and younger strata and marked by earthquake seismicity, Quaternary volcanism and locally elevated heat flow [Larsen and Reilinger, 1991; Mueller and Rockwell, 2000a].
Dextral shear has been accommodated south of the Salton Trough in the northern Gulf of California by closely spaced transform faults that are linked and separated by highly extended sedimentary basins such as the Guaymas, Tiburon, Upper and Lower Delfín and other basins (Figure 1b [Lonsdale, 1989; Stock, 2000; Aragon-Arreola and Martín-Barajas, 2007]). Rapid extension in this part of the Gulf began after ca 6.2 Ma [Oskin et al., 2001]. Transform faults near the mouth of the Gulf of California are linked by short, actively spreading ridges composed...
mostly of igneous rocks of mafic composition \cite{Lonsdale, 1989}. These spreading centers are marked by relatively little sedimentary infill and deep bathymetry, in contrast to similar regions of highly extended crust formed further north in the Gulf of California and Salton Trough that are covered with thick sequences of Pliocene (?)-Quaternary sediment. Quaternary strata in the Salton Trough are metamorphosed at greenschist facies conditions at shallow, upper crustal levels \cite{Fuis et al., 1984; Muffler and White, 1969}, as a result of the high heat flow in this region \cite{Lachenbruch et al., 1985}.

Quaternary extension in the Salton Trough is likely associated with mafic magmatism at deeper crustal levels \cite{Fuis et al., 1984}, which can be argued as beginning at ca. 4 Ma \cite{Axen et al., 2006}. Thin crust (21–22 km) in the Salton Trough \cite{Parsons and McCarthy, 1996; Zhu and Kanamori, 2000}, and a temperature gradient of \text{33.3°C/km} \cite{Lachenbruch et al., 1985} all suggest that crustal extension, magmatism and heating of the upper mantle are active processes beneath the region \cite{Schubert and Garfunkel, 1984}.

Spreading in the northern Gulf of California apparently underwent a westward shift along much of its length in Pliocene time, as first identified by Nagy and Stock \cite{2000} in the Tiburon-Delphin and Guaymas Basins at ca. 3 Ma, and subsequently by Aragon-Arreola and Martín-Barajas \cite{2007} in this same region. A similar westward shift in active spreading has also been recognized near the mouth of the Gulf near the Alarcon Rise \cite{Lizarralde et al., 2007}.

Quaternary extension in the Gulf of California is also accommodated by normal faults that bound the eastern edge of the Peninsular Ranges and dip east, forming an escarpment that extends the length of the Baja Peninsula. These structures include the east rooted Sierra San Pedro Martir fault system \cite{Gastil et al., 1975; Dokka and Merriam, 1982; Stock, 2000}, which bounds the highest rift shoulder segment along the Baja California Peninsula (Figures 1a, 1b, and 2). The San Pedro Martir fault has a steep, continuous footwall scarp with large (~5 km) displacement and scarps that offset late Quaternary alluvium (Figure 2 \cite{Dokka and Merriam, 1982; Brown, 1978}). The San Pedro Martir fault segment of the Main Gulf Escarpment terminates abruptly to the south at the Puertecitos Volcanic Province that is associated with the Motamí accommodation zone (Figures 1b and 2 \cite{Stock, 2000}).

The Peninsular Ranges rise steeply west of the Gulf Escarpment between 30°–34°N latitude (Figures 1a and 2) to a maximum height of 3095 m with elevations gradually decreasing westward along a concave upward slope \cite{Lee et al., 1996}. Maximum elevations along the crest of the range vary between 1500 and 3000 m with the highest topography located in the Sierra San Pedro Martir (Figure 3a). The Peninsular Ranges are composed primarily of Mesozoic igneous rocks that were exhumed and cooled through zircon fission track closure temperatures during subduction in late Cretaceous to Eocene time \cite{Cerveny et al., 1991; Ortega-Rivera, 2003; Lovera et al., 2006}.

Igneous rocks of the Peninsular Ranges were subsequently covered by Eocene conglomerate derived from a source in northern Sonora and deposited in a channel system that extended west across southern California to the California borderlands \cite{Minch et al., 1976; Kies and Abbott, 1983; Abbott and Smith, 1989}. Now largely stripped of the overlying conglomerate across the top of

![Figure 2](image-url). Oblique view of Baja Peninsula from 90 m SRTM data. View to south. Note Gulf Escarpment is defined by 1–3 km high fault scarps shown in shadow. Illumination is from southwest.
The Peninsular Ranges, the nonconformity, or basement cover contact can be identified as a low-relief, eroded surface that dips gently toward the Pacific coast and is mapped as high as 1830 m, approximately 60 km east of the Pacific coast at N32°50' [Gastil, 1961; Lovera et al., 2006].

The Peninsular Ranges must have been relatively low in elevation prior to extension and rift flank uplift, on the basis of the distribution of Eocene conglomerate transported from Sonora westward across southern California [Kies and Abbott, 1983] and northern Baja California [O'Connor and Chase, 1989]. The provenance of clasts in Eocene conglomerates also suggests little input from local granitic sources in the Peninsular Ranges when they were deposited [Abbott and Smith, 1989], consistent with a lack of significant relief. In addition, the modern river channel network developed on the Laguna Mountain, Sierra Juarez and Sierra San Pedro Martir rift shoulder segments is not deeply incised except near the Pacific coastline. We interpret this as additional evidence for low relief prior to recent rock uplift otherwise paleochannels cut into batholithic rocks would be more deeply incised at higher elevations in rift shoulder segments. On the basis of this evidence, we assume that prior to uplift, the Peninsular Ranges lay at a relatively low elevation (sloping very gently toward the Pacific coast), and therefore initial topography in our models is approximated as sea level. The onset of surface uplift in the eastern Peninsular Ranges in northern Baja California is not well constrained, but desert soils, or aridisols of mid Pliocene age exposed in the Salton Trough is interpreted to form in a rain shadow consistent with an existing rift flank uplift at this time [Peryam et al., 2008].

Modeling presented later in this paper attempts to fit the total uplift of the initially gently west dipping Eocene surface, thus we are only concerned with the net change in buoyancy since the Eocene. This uplift is presumed to be a...
product of Miocene and younger normal faulting, erosion, and lithospheric thinning or heating.

3. Marine Terrace Uplift Along the Pacific Coast

[16] The Pacific coast west of the Peninsular Ranges, from 30° to 33°N, is characterized by flights of marine terraces that imply continuous uplift during the late Quaternary. Many of the most prominent terrace platforms are present in areas where faults locally influence and control coastal structure and geomorphology, such as in the Palos Verdes Peninsula [Woodring et al., 1946; Woodring, 1948; Muhs et al., 1992], San Joaquin Hills [Grant et al., 1999], the Mt Soledad region of San Diego [Kern and Rockwell, 1992], and the Punta Banda region of Baja California [Rockwell et al., 1989] (Figures 3a and 3b). Shortening above blind thrusts, or in restraining bends of strike-slip faults, produces local uplift in these areas that is superposed on regional uplift that occurs at a lower rate.

[17] It is the regional and remarkably uniform uplift signal along the Pacific coast, however, that is of primary interest in this study. Local structures could only produce such a signal if a fault paralleled the coast and had nearconstant displacements. A west facing escarpment mapped offshore of the Pacific coast in northern Baja California could reflect vertical motion just offshore, but existing earthquake focal mechanisms indicate strike-slip motion [Rebollar et al., 1982]. Although we cannot completely reject the possibility that this fault contributes in some way to coastal uplift, the great length and uniformity of the regional uplift of terraces seems incompatible with local fault control. While rock and surface uplift in Baja California is certainly modified by local faulting on the scale of tens of kilometers or more, we seek to characterize and constrain the larger-scale pattern of elevation change that extends across and along the northern peninsula.

[18] The late Quaternary rate of regional, or background uplift is well known for much of the Pacific coast in southern California and northern Baja California and averages from 0.13 to 0.14 mm/a from southern California southward to the Ensenada region (Figures 3a and 3b). South of the Agua Blanca fault, the uplift rate is similarly low at about 0.13 mm/a to 30° north latitude (interpreted from Ellis and Lee [1919], Emerson [1956], Valentine [1957], Emerson and Addicott [1958], Addicott and Emerson [1959], Emerson and Hertlein [1960], Valentine and Rowland [1969], Orme [1974], Emerson [1980], Lindberg et al. [1980], Emerson et al. [1981], Ortlied et al. [1984], Kennedy et al. [1986], Ashby et al. [1987], Rockwell et al. [1989], Kern and Rockwell [1992], Muhs et al. [1992, 2002], and G. L. Kennedy and T. K. Rockwell, unpublished survey data, 1995). Terraces along the Pacific coastline south of 28.6° north latitude to the southern tip of the Baja California Peninsula have not been uplifted except where they are locally deformed along strike slip faults (Figures 3a and 3b [Muhs et al., 1992; Kennedy and Rockwell, unpublished data, 1995]).

[19] Although the absence of terrace uplift along the Pacific coast is largely the result of a narrower rift flank uplift as defined by regional topography in the southern peninsula, the east rooted Santa Margarita and San Lazaro faults normal faults defined just offshore south of the Vizcaino Peninsula [Fletcher and Eakins, 2001; Fletcher et al., 2000b, 2000c, 2007] may also affect the record of coastal uplift here. We therefore focus our attention on the uplift signal north of about 28.6°N.

[20] The highest observed marine terrace in southern California located away from active structures is the Rifle Range terrace in San Diego County that now stands at 155 m above sea level [Kern and Rockwell, 1992]. Higher terraces only exist within the Mount Soledad, Palos Verdes and San Joaquin uplifts in southern California where active faulting produces locally higher topography. Pliocene to early Pleistocene marine sediment of the San Diego Formation deposited at sea level are preserved in the San Diego region at nearly the same elevation of the highest Quaternary terraces. This can be interpreted to suggest that total Quaternary uplift does not exceed about 155 m, on the basis of the eustatic record for Pliocene time [e.g., Dowsett and Cronin, 1990]. Given the present elevation of the San Diego Formation and the Quaternary marine terraces in the San Diego region, we thus argue that coastal southern California has undergone about 155 m of total surface uplift since middle to late Pliocene time.

[21] Southward into Baja California, the highest marine terraces along the Pacific coast have been described as Pliocene in age on the basis of the presence of extinct fauna [Emerson and Hertlein, 1960; Hertlein and Allison, 1959]; the maximum elevation of Pliocene marine fauna occur at elevations greater than 100 m in northern Baja California, and decrease to about 20–30 m elevation at Santa Rosalalita near the Vizcaino Peninsula. Age controls for these deposits are poor, and it is possible that some of the extinct fauna described in these strata are early Quaternary in age. Nevertheless, these observations are consistent with the much better described terraces and their inferred ages described in southern California. We use 155 m as the maximum post-Pliocene surface uplift where terraces are not affected by local structures.

[22] An extensive terrace including the Mesa Los Indios and Mesa El Tigre surfaces located near the international border in northernmost Baja California, is preserved at elevations of 350–380 m. Remnant benches above 400 m may represent older and poorly preserved terrace remnants, but the ~350 m surface northwest and inland from La Mision is the highest, well-formed and well-preserved marine terrace in the vicinity. The surface is capped by a cobbles lag, as observed from an aerial overflight of the area, although no fieldwork has yet been undertaken to locate marine fauna. This terrace may be very old, and possibly Tertiary in age.

[23] Uplifted marine deposits preserved along the Gulf of California also provide constraints on regional strain in the Peninsular Ranges. U-series dating of corals in southern Baja California along the Gulf of California indicate low rates of uplift that range from 0.3 to 0.19 mm/a [DeDiego-
Forbis et al., 2004]. Near Mulage, Ashby et al. [1987] document a rate of about 0.05 mm/a for the past ~120 ka. These values are more than an order of magnitude lower than geodetically determined rates derived from the crest of the eastern Peninsular Ranges, in northern Baja California (e.g., 4.8 ± 1.5 mm/a [Outerbridge et al., 2005]). We do not consider published uranium series age determinations using molluscs in Baja California [e.g., Gastil, 1983] owing to problems associated with the open system exchange of isotopes known to occur in samples of this type.

4. Rift Shoulders

[24] Rift shoulders are common topographic features along extensional basins and continental rift margins throughout the world [Vening Meinesz, 1950]. Examples include the Transantarctic Mountains in Antarctica [Bott and Stern, 1992; Stern and Ten Brink, 1989], and the flanks of the Rhine Graben [Weissel and Karner, 1989], the Rio Grabin in Greece [Poulimenos and Doutsos, 1997], and the East African Rift [Vening Meinesz, 1950; Bott, 1992; Zeyen et al., 1997]. As the hanging wall subsides to form a rift basin, the flanking footwall block is uplifted. Maximum subsidence occurs along the rift axis, while uplift of rift shoulders occurs beyond the area offset by normal faults [Chéry et al., 1992]. Rift shoulders range in height from 1000 to 5000 m along continental rift systems and vary in shape and amplitude as a function of both the uplift force and the rigidity of the lithosphere [Chéry et al., 1992]. Rift shoulders are asymmetric, with a steep escarpment facing the rift and gradually decreasing elevation along a concave-up slope away from the rift [Stern and Ten Brink, 1989].

[25] As early as 1976, topography observed across the Peninsular Ranges and Salton Trough was qualitatively compared to other rift shoulders around the world (Figure 3a [Johnson et al., 1976]) or related to a suppressed root [O’Connor and Chase, 1989]. The maximum elevations along the range crest adjacent to the western edge of the Salton Trough vary from ~3000 m in the south to ~1500 m in the north. These regions of high-relief form distinct rift segments that are coincident with both east and west rooted extensional fault systems of Neogene age [Axen, 1995] and younger Quaternary pull-apart basins and spreading centers (Figures 1b and 3a; see separate rift shoulder segments labeled Laguna, Juarez and Martir).

5. Origin of Buoyancy Forces

[26] To explore the relationship between extension-driven uplift and the uplift of the terraces along the California and Baja California coasts, we first consider simple flexural models with single line loads and then develop solutions that incorporate 2-D loads in cross section. The simple line load models help reveal the tradeoffs in determining the contributions of loads produced by different sources needed to produce the observed flexure; they are a good initial approximation to reality because the topography we seek to match is outside much of the area where the load is created. We presume that any loads created by variations in the composition of the batholith [e.g., Gastil, 1983] were isostatically compensated before the initiation of rifting and before the development of the broad gentle surface we use as a reference, and therefore we are only concerned with the change in loads during the Neogene.

[27] For an elastic plate subjected to an upward force in the vicinity of the rift, the upward force represents removal of rock by erosion and normal faulting at the top of the crust, thermal expansion of the crust and thinning of the mantle lithosphere at depth. This upward force is reduced by the deposition of sediments, the shallower depth of the Moho along the rift axis and any metamorphism of, or intrusion of mafic material into the crust. Contributions to the total load from different sources are shown in Figure 4. The resulting deformation is determined by the magnitude of the upward force, \( V \), which controls the magnitude of uplift, the flexural parameter \( \alpha \), which determines the width of uplift or flexural wavelength, and the distance of the force from the coast \( x_0 \), which simply translates the deformation in the direction perpendicular to the rift. The upward deflection, \( w(x) \), is then defined

\[
\begin{align*}
  w(x) &= w_0 e^{-\frac{x-x_0}{D}} \left( \cos \left( \frac{x-x_0}{\alpha} \right) + \sin \left( \frac{x-x_0}{\alpha} \right) \right)
\end{align*}
\]

where the maximum deflection \( w_0 \) is related to the force through

\[
  w_0 = \frac{V \alpha^3}{8D}
\]
where $D$, the flexural rigidity of the plate, and $\alpha$, the flexural wavelength, are

\[ \alpha = \sqrt{\frac{4D}{\rho_m - \rho_f}} \frac{1}{g} \]

\[ D = \frac{ET^3}{12(1 - \nu^2)} \]

and $T_e$ is the elastic plate thickness [e.g., Turcotte and Schubert, 2002]. We assume a Young’s modulus $E$ of $10^{11}$ N m$^{-2}$ and Poisson’s ratio $\nu$ of 0.25. For this calculation we use mantle density $\rho_m$ of 3250 kg m$^{-3}$ and a fill density of 0.

[26] If the aim of the models is only to fit the western flank of the rift (i.e., the topography of the Peninsular...
Ranges), a tradeoff exists between these three parameters: \( x_f, V \), and \( \alpha \) can be varied to generally fit the overall slope of the profile. The main difference between these parameters within the Peninsular Ranges is a slight change in the curvature of predicted rift flank topography (Figure 5a). In all cases the point where uplift begins to occur (moving from west to east) near the Pacific coastline is located offshore by some 40 km; moving it farther east misfits the topography near the coastline. In general, a similar shape across the Sierra Juarez is obtained by shifting the load farther east while increasing both the flexural rigidity and the load. Because we do not know the preflexure topography in detail, we cannot fully resolve the tradeoff between force, position, and flexural parameter without additional information. We also do not consider the effect of loading and flexure on the Sonoran side of the Gulf of California, owing to the more complex extensional history this region has undergone and uncertainty regarding the prerift topography that may have been present there.

As an alternative end-member, a broken plate model approximates the case where strength in the lithosphere becomes very low within the rift. For a given load position, the elastic plate thickness must be \(~80\%\) greater and the load slightly smaller to produce a deflection comparable to that obtained from a continuous plate. Unless the plate edge is located well to the west of the main basin edge, the effective elastic plate thickness \( T_e \geq 40 \text{ km} \) required to fit the topography on the northern profile seems too high for lithosphere in southern California [e.g., Lowry et al., 2000]. The combination of the higher \( T_e \) and steeper topography associated with the broken plate model make it a poorer choice for the northern profile than the continuous model of Figure 5a.

The origin of the buoyancy driving flexure can be determined by considering the forces in these simple mod-
els. For the profile across the Salton Trough, a line force of about $10^{13}$ N/m applied to a continuous plate within the rift acceptably reproduces the topography (Figures 4 and 5a). The upward buoyancy force from replacing granitic crust with the sedimentary fill in the Salton Trough only amounts to between about $2.5 \times 10^{11}$ N/m and $10 \times 10^{11}$ N/m, roughly an order of magnitude too small to produce observed rift flank uplift. The surface load caused by removing rock in the rift through either erosion or extension is proportional to the area between the flexural curve and the present topography; this varies with the flexural rigidity assumed. In all cases, the surface load is about 40–80% of the necessary load, with the greater deficiencies for loads placed farther west (Figure 5a). In addition, the crust thins and crustal density increases from west to east across the western edge of the Salton Trough [Fuis et al., 1984], producing a significant downward force in the crust along the rift perhaps twice the magnitude of the upward force from the sedimentary basin. Thus much of the buoyancy force needed to drive flexural uplift (i.e., that matches observed rift flank topography) must be derived from the mantle under the rift (Figure 4). This is also true of the broken plate models because the load required from extension is (mis)located 65 km west of active normal faults in the Salton Trough.

[31] A similar analysis for the southern profile across the Sierra San Pedro Martir yields comparable results though the required load is roughly twice that of the northern profile for any particular load position. With a continuous plate, loads between 1.2 and $3.5 \times 10^{13}$ N/m applied to a plate with an elastic plate thickness from 17 to 30 km will match the topography of the Sierra San Pedro Martir, with the thicker plate and larger load corresponding to a load near the center of the Gulf of California. A slightly better fit to the topography is obtained by using a broken plate model with somewhat smaller loads and ~80% greater elastic plate thicknesses than the continuous plate model (Figure 5b). We lack observations here comparable to those along the northern profile constraining the sedimentary thickness, but it seems unlikely that load generated by replacing crystalline rock with sediments will be more than double that in the Salton Trough. The main surface load is that produced by removing crust through faulting and erosion; the load could be somewhat higher than to the north, perhaps as high as $2 \times 10^{13}$ N/m for a load centered in the Gulf of California. Although this is almost 80% of the load needed, the downward load from a shallower Moho and probable crustal density increases into the Gulf of California again indicate an important, if not dominant, contribution from buoyancy in the mantle. Furthermore, the comparison of the two profiles suggests a somewhat weaker plate and a noticeably greater load in the south than the north.

6. Flexural Modeling of Sierra Juarez and Salton Trough

[32] Because the actual loads are not line loads but in fact distributed along the profile, we investigate the loads producing rift flank uplift with more realistic 2-D models. To better determine the source, location, and magnitude of the load uplifting the Sierra Juarez, we modeled the uplift as the result of extensional thinning of the crust from normal faulting and variable thinning of the mantle lithosphere (Figures 6, 7, and 8). The basin, Moho and lithosphere loads are specified for each trial, and the deflection of the plate is calculated. The material removed by erosion or crustal tectonism is then estimated to be equal to the mass between the deflected position of the plate top and the modern topography. The load from this estimate is then added to the original loads and a new deflection is calculated, which in turn leads to a new estimate of the load from erosion or crustal tectonism. We continue to iterate until no new forces are generated. We force our flexural uplift near the coastline to match the 155 m maximum elevation of shoreline terraces described above and seek to match deflections of an early Tertiary, low-relief surface.

[33] The load produced from variation in the thickness of the crust are specified from seismological inferences of crustal thickness [Ichinose et al., 1996]. We use a 15 km thinner crust under the sediment-filled Salton Trough with thinning tapering to zero 50 km from the edges of the basin. Our model of crustal thickness mimics that of Ichinose et al. [1996] who defined crustal thickness across the Peninsular Ranges and Salton Trough of southern California near our northern transect (see crustal depth points defined in Figure 8). Their work was based on using P-to-S converted phases of teleseismic body waves on a broadband array and has been found to be representative of the Peninsular Ranges to the north and south [Lewis et al., 2000, 2001; Yan and Clayton, 2007]. We have assigned a density contrast of 400 kg m$^{-3}$ to the Moho for our flexure models. This presumes a relatively flat Moho prior to the Neogene. If more crustal thinning occurred, then a more buoyant mantle is required, and if less thinning, then a less buoyant mantle.

[34] Thinning of the upper mantle is considered by replacing lithosphere with less dense material similar to the asthenosphere beneath the Salton Trough and eastern Peninsular Ranges. The replaced region is modeled by a 80- to 100-km-thick, rectangular or trapezoidal body 100 kg/m$^2$ less dense than the mantle lithosphere (Figures 6, 7, and 8). Note that as the load is in proportion to the density contrast and thickness, a thicker (or thinner) variation will produce the same load if the density contrast varies proportionally. Both for simplicity and because there is an insignificant difference between a rectangular body and a trapezoid with an edge less than about 60 km wide, most models assume a rectangular body.

[35] The fit of models to observed topography is judged principally by the overall fit to the topography of the west slope of the Sierra Juarez. Because we force each model to pass through the 155 m elevation of the swath topography, much of the difference between models is near the crest. Ideally, acceptable models would pass somewhat below modern topography, making the Eocene surface dip gently toward the ocean before flexure had occurred. We use the west edge of the deepest basin in the Salton Trough (140 km east of the coast) as our zero point ($x = 0$) in this modeling.
6.1. Elastic Plate Thickness

[36] We first consider plausible plate thicknesses \( T_e \) in southern California and northern Baja California. There is a tradeoff between the loads imposed and \( T_e \): a thickness of 0 requires a load exactly reflecting topography, while a stiffer plate permits the load to be located farther east. An upper bound for \( T_e \) in this region is 30 km [e.g., Lowry et al., 2000]. Using a constant \( T_e \) of 30 km and holding the eastern edge of the mantle load at \( x = 100 \) km, we find that the mantle load should not extend west of the west edge of the deep basin (Figure 6a). These solutions require large mantle loads equivalent to the removal of 170 km or more of mantle lithosphere 100 kg m\(^{-3}\) denser than asthenosphere (equivalent to line loads of about \( 2 \times 10^{13} \) N/m), and these require the tectonic removal of 3 or more km of crust in the eastern Chocolate Mountains. We find the magnitude of the load too large to be plausible given lithospheric thicknesses in California [e.g., Yang and Forsyth, 2006].

[37] Smaller values of \( T_e \) require loads more consistent with lithospheric variations expected in the region. A \( T_e \) of 20 km requires a load extending to the west of the western edge of thinned crust and into the eastern Sierra Juarez, with about 95 km of lithosphere removed (Figure 6b). Unlike the \( T_e = 30 \) km case, the eastern edge of the load can be located far enough to the west to allow for minimal removal of crust in the Chocolate Mountains. A \( T_e \) of 15 km represents the lowest possible flexural rigidity capable of reproducing the topography above a sharp-edged mantle load (Figure 6c). In this case, the load magnitude trades off dramatically with the amount of material that must be removed within the rift to match the topography, thus greatly limiting the range of acceptable solutions. These results indicate that for any \( T_e \) less than or equal to about 20 km, mantle loading must extend west of the major basin bounding fault by at least 50 km and west of the westernmost range-bounding faults by more than 15 km.

Figure 6. (a–c) Series of flexural models across northern transect assuming varied amounts of mantle lithosphere replaced by asthenosphere and different values for \( T_e \). Location of transect and all subsequent models shown in Figure 1a. Basin fill geometry in Salton Trough is derived from Fuis et al. [1984]. Uppermost 6 km of crust is highly (>30 times) vertically exaggerated. Solutions are based on a constant \( T_e \) for a given model, with increasing \( T_e \) from Figures 6a to 6c. All models maintain an elevation of 155 m at the Pacific coastline, as constrained by uplift of marine terraces. Figure 6a illustrates solutions for a relatively narrower and deeper region of mantle lid replaced by asthenosphere beneath the eastern Salton Trough for a relatively thick \( T_e \) of 30 km. Figure 6b shows intermediate solutions for replaced mantle lid centralized beneath the Salton Trough and a \( T_e \) of 20 km. Figure 6c defines region of replaced mantle lid that is thinner and wider but with a relatively thinner \( T_e \). Parameters in models illustrate the tradeoff between the location and magnitude of various loads and resulting rock uplift as illustrated by rift flank topography in the Sierra Juarez.
A more plausible scenario is that the effective elastic plate thickness thins into the rift. In general, elastic plate thickness thins with increased heat flow. We focus on a model with $T_e = 24$ km away from the rift and $T_e = 10$ km in the rift (from $x = 0$ to $x = 33$ km) with a gradient in elastic plate thickness about 30 km wide on each side of the rift. In this case (Figure 7), we find that topography is quite sensitive to the position of the left edge of the mantle load. This is in part because of the feedback of forcing material to be removed in the rift: as the mantle load grows, the erosional load grows as well. The right edge of the load has little effect on the topography in the Sierra Juarez, but this does change the amount of material that must be removed from the Chocolate Mountains. If the Chocolate Mountains are an equivalent flexural shoulder to the rift, then the mantle load ends somewhere between 65 and 80 km east of the axis of the rift basin.

Just as with solutions with a constant $T_e$, below about 20 km, the variable $T_e$ case requires the mantle load to extend about 50 km to the west of the main basin bounding fault and about 20 km west of the eastern flank of the Sierra Juarez. Significantly, this is also 20–25 km west of the decrease in elastic plate thickness. These results indicate that the mantle load must extend well to the west of a thinner elastic plate, which is equivalent to the part of the crust that has been heated and damaged by rifting. The magnitude of the load is equivalent to removal of about 80 km of mantle lithosphere. Thus we find that the buoyancy in the mantle extends a minimum of 20 km (and probably closer to 50 km) west compared with the western limit of substantial crustal deformation.

### 6.2. Timing of Uplift and Loading

We noted above that the Salton Trough has experienced multiple episodes of extensional deformation since the Miocene. However, the coastal uplift is post-Pliocene. One possibility is that the entire mantle load causing uplift of the rift shoulders postdates the Pliocene. This would require that earlier crustal thinning be rooted into the mantle west of the Sierra Juarez and the Salton Trough. The least amount of mantle loading can be found by noting that the terraces are near the nodal point for flexural uplift from a load in the interior. In Figure 8, we show that a 20–30 km westward shift in the west edge of the mantle load is sufficient to generate the 155 m uplift at the coast. This lateral growth of the load also produces about half of the topography of the Sierra Juarez, suggesting that the elevation of this range could also be mostly Pliocene and younger. However, the magnitude of uplift of the crest of the Sierra Juarez depends on the dip of the west edge of buoyant mantle; if the dip allows buoyant material farther west than shown in Figure 8, then it is possible to raise the coastal areas while only raising the crest a few hundred meters, or, in the most extreme case, only 30 m.

Oxygen isotope analysis of paleosols and magnetostratigraphy in the Fish Creek–Vallecito basin in the western Salton Trough suggest that a rain shadow existed there prior to ca. 3.7 Ma [Peryam et al., 2008]. Aridisol soil chemistry also suggests general aridity in this region between ca. 1.0 and 3.7 Ma [Peryam et al., 2008]. While this does not tightly constrain the elevation of a supposed flank uplift, it is generally consistent with our arguments for
Figure 8. Flexural model across Sierra Juarez that shows an east to west progression of rift flank uplift in the Peninsular Ranges. Model is intended to illustrate an evolving rift that migrates toward the Pacific from the Salton Trough with 155 m of coastal uplift. Basin geometry similar to previous models, with a variable $T_e$ across the Salton Trough. White circles denote base of crust as defined by receiver functions from Ichinose et al. [1996].
the timing of rift flank uplift driven by increased mantle buoyancy.

7. Discussion

[42] Thinning, or strong heating of the upper mantle under the eastern Peninsular Ranges is consistent with the previous results of Lewis et al. [2000] who require a significant mantle component to support topography west of the Main Gulf Escarpment. Ichinose et al. [1996] also observe the lack of correlation between topography and crustal thickness (e.g., depths to the Moho) and suggest compensation from lateral density variations in the lower crust and upper mantle.

[43] All models include removal of upper crustal material by either erosion or normal faulting so that the modern topography within the rift is honored. We have not attempted to explicitly include erosion to the west of the rift, nor have we tried to estimate a priori the total amount of material removed from within the rift. Erosion west of the rift is volumetrically minor and does not significantly affect our results. Estimating the total volume of material removed within the rift is problematic, largely because the original Moho geometry is unknown and any surficial geologic constraints are complicated by poor control on fault geometry at depth and possible lateral flow of lower crust. Despite these difficulties, the magnitude of buoyancy in the mantle is only overestimated if the crust was originally thinner at the site of the Salton Trough prior to extension (relative to the region to the west).

[44] Simple models of crustal extension coupled with near complete replacement of mantle lithosphere predict relief that broadly matches observed basin geometry and rift flank swath topography across the Peninsular Ranges and Salton Trough in southern California and Baja California (Figures 5, 6, 7, and 8). Thinning of the lithosphere in this region has produced a western rift flank uplift composed of the Laguna Mountains, Sierra Juarez and Sierra San Pedro Martir that rise to an average elevation of ~2 km above sea level. Uplift of the western rift flank apparently extends westward to the Pacific coastline, where late Quaternary marine terraces are raised as much as 155 m above sea level. A narrow and lower rift flank uplift bounds the eastern side of the Salton Trough that is also predicted by our models and marked by the Chocolate Mountains. While little evidence exists to exactly constrain the timing of uplift of the higher parts of the rift flanks, the age of uplifted marine terraces and shallow marine deposits along the Pacific coast suggest regional uplift may be very young and largely Pleistocene in age. This is supported by late Pliocene to Pleistocene soils developed in the western Salton Trough that imply the presence of a rain shadow and hence significant relief in the eastern Peninsular Ranges at this time. It is important to note that rates defined by the marine stage 5e terrace only provide an average uplift rate for the last ~120 Ka, and that total uplift defined in southern California of ~155 m must have been initiated well before this time (but after shallow marine Pliocene strata were deposited, and at about 1.3 Ma on the basis of dating of older terraces with amino acid racemization results and uplift rates).

[45] Unpublished geodetic data [Outerbridge et al., 2005] support our work by suggesting that rapid rates (4.8 ± 1.5 mm/a) of modern uplift may occur in the higher parts of the Peninsular Ranges province. The remarkably consistent rate of slow background uplift of marine terraces along the Pacific coastline and negative gravity along the range crest is interpreted as evidence for large-scale flexure of the lithosphere driven largely by a broad mantle upwelling beneath eastern Peninsular Ranges, Salton Trough and northern Gulf of California. At the scale of the width of the Baja Peninsula, this suggests that coastal uplift is relatively unaffected by the segmentation of late Quaternary extensional faults in the Salton Trough and Gulf of California [Axen, 1995]. A lower rift flank uplift preserved south of the Sierra San Pedro Martir in southern Baja also suggests buoyant mantle adjacent to the Gulf of California, although to a lesser extent than farther north.

[46] Our work thus suggests a fundamental increase in buoyancy in the upper mantle in the northern Gulf of California that is presumably related to the thermal conditions that have affected this region. Rift flank uplift of the Peninsular Ranges is thus partly coeval with crustal extension driven by transform tectonics, as the timing of coastal uplift does not match well-documented evidence for prior Miocene extension along low-angle detachment faults in this region. We envision the mantle load that drives rift flank uplift as migrating westward from the Salton Trough and Gulf of California by some combination of lithospheric thinning and thermal erosion (Figure 8). The onset of uplift at the Pacific coastline must, however, be Quaternary in age. Our models also hint that the current topography of the Chocolate Mountains may be related to the relatively recent development of the Gulf of California, although there are essentially no constraints on the true timing of uplift in this region. We also have not considered the effects of substantial Miocene thinning along the east rooted Colorado River extensional corridor farther to the east and its effect on the current topography of the Chocolate Mountains. An interesting aspect of the models presented in this paper is the enormous magnitude of mantle lid thinning required to reproduce rift flank topography. For example, the best fit solutions suggest 80 km of lid loss over a 120–150 km wide region, perpendicular to the Gulf of California. While a number of previous workers have argued for significant heating of the mantle, or loss of the mantle lid [Fuís et al., 1984; Lewis et al., 2001] beneath parts of the rift, none have suggested that crust in the northern Gulf of California and Salton Trough overlies asthenosphere over the width of the rift floor and adjacent eastern Peninsular Ranges as indicated by our models. However, Park et al. [1992] suggested the imposition of a major heat source in the past 5 My under the Peninsular Ranges and Lewis et al. [2001] argued for a significant mantle component to support topography east of the Main Gulf Escarpment (located about 20 km east of the range crest).

[47] Whereas the western rift shoulder is marked by discrete segments along the length of Baja California, it
appears to be uplifting at significantly lower rates south of the Sierra San Pedro Martir, on the basis of the morphology of deeply embayed range fronts, lower overall rift flank relief and rates of uplift defined by Late Pleistocene marine deposits near La Paz and Los Cabos [Muhls et al., 1992; DeDiego-Forbis et al., 2004]. Therefore, while the southern Gulf of California has undergone crustal extension across spreading ridges comparable to similar transform and ridge segments farther to the north, mantle buoyancy changes in the Quaternary appear to be lower to the south, on the basis of rift shoulder relief. This is consistent with recent surface wave studies in the region that show very low mantle wave speeds in the northern Gulf of California but higher wave speeds in the south [Zhang et al., 2007].

[48] Our work suggests that the regional background uplift of the Pacific coastline extends from southern California to nearly the Vizcaino Peninsula and is not related to local shortening above active blind thrusts [Rivero et al., 2000] but is instead related to a combination of crust and mantle thinning centered on the Gulf of California and eastern Peninsular Ranges. This broad uplift appears to best correspond to a phase in development of the new spreading system in the Gulf of California when early regional extension becomes more localized but before the system evolves into a better defined seafloor spreading boundary.

[49] Our hypothesis links uplift of the Pacific coast and Sierra Juarez to either conductive or convective heating in the mantle lithosphere under the Peninsular Ranges. Further tests of this idea will either constrain the thermal evolution of the mantle lithosphere under the Peninsular Ranges or better constrain the temporal evolution of topography toward the crest of the range. Our interpretation requires substantial post-Pliocene removal or modification of mantle lithosphere well beyond the edges of Neogene crustal strain recorded in the western Salton Trough; the mechanism(s) accommodating such deformation remain undetermined but could include low-angle normal faulting, lateral heat conduction, or convective instabilities in the lithosphere.

[50] Acknowledgments. We thank Garry Kerner for his comments on our early modeling work and Tim Melbourne and Mike Oskin for their discussions of mantle structure and the Neogene tectonic history of the inner California borderland and the Gulf of California, respectively. Kelin Whipple, Kip Hodges, Suzanne Jannecke, Rebecca Dorsey, Brian Wernicke, and in particular, John Fletcher and Gary Axen provided many helpful suggestions, resulting in a much improved final version of this paper. This research was supported by the Southern California Earthquake Center (SCEC). SCEC is funded by NSF cooperative agreement EAR-8920136 and USGS cooperative agreements 14-08-0001-A0899 and 1434-HQ-97AG01718. The SCEC contribution number for this paper is 580.

Research was further supported by NEHRP grant 01 HQCR0031 to K. Mueller. The views and conclusions contained in this document are those of the authors and should not be interpreted as necessarily representing official policies of the U.S. government.

References


Cerveny, P. F., R. J. Dorsey, and B. A. Burns (1991), Late-Neogene transpression, northern Gulf of California: palaeoearthquakes and lithospheric response, paper presented at the American Geophysical Union Fall Meeting.

Cerveny, P. F., R. J. Dorsey, and B. A. Burns (1991), Late-Neogene transpression, northern Gulf of California: palaeoearthquakes and lithospheric response, paper presented at the American Geophysical Union Fall Meeting.

Cerveny, P. F., R. J. Dorsey, and B. A. Burns (1991), Late-Neogene transpression, northern Gulf of California: palaeoearthquakes and lithospheric response, paper presented at the American Geophysical Union Fall Meeting.

Cerveny, P. F., R. J. Dorsey, and B. A. Burns (1991), Late-Neogene transpression, northern Gulf of California: palaeoearthquakes and lithospheric response, paper presented at the American Geophysical Union Fall Meeting.

Cerveny, P. F., R. J. Dorsey, and B. A. Burns (1991), Late-Neogene transpression, northern Gulf of California: palaeoearthquakes and lithospheric response, paper presented at the American Geophysical Union Fall Meeting.

Cerveny, P. F., R. J. Dorsey, and B. A. Burns (1991), Late-Neogene transpression, northern Gulf of California: palaeoearthquakes and lithospheric response, paper presented at the American Geophysical Union Fall Meeting.

Cerveny, P. F., R. J. Dorsey, and B. A. Burns (1991), Late-Neogene transpression, northern Gulf of California: palaeoearthquakes and lithospheric response, paper presented at the American Geophysical Union Fall Meeting.

Cerveny, P. F., R. J. Dorsey, and B. A. Burns (1991), Late-Neogene transpression, northern Gulf of California: palaeoearthquakes and lithospheric response, paper presented at the American Geophysical Union Fall Meeting.

Cerveny, P. F., R. J. Dorsey, and B. A. Burns (1991), Late-Neogene transpression, northern Gulf of California: palaeoearthquakes and lithospheric response, paper presented at the American Geophysical Union Fall Meeting.

Cerveny, P. F., R. J. Dorsey, and B. A. Burns (1991), Late-Neogene transpression, northern Gulf of California: palaeoearthquakes and lithospheric response, paper presented at the American Geophysical Union Fall Meeting.

Cerveny, P. F., R. J. Dorsey, and B. A. Burns (1991), Late-Neogene transpression, northern Gulf of California: palaeoearthquakes and lithospheric response, paper presented at the American Geophysical Union Fall Meeting.

Cerveny, P. F., R. J. Dorsey, and B. A. Burns (1991), Late-Neogene transpression, northern Gulf of California: palaeoearthquakes and lithospheric response, paper presented at the American Geophysical Union Fall Meeting.

Cerveny, P. F., R. J. Dorsey, and B. A. Burns (1991), Late-Neogene transpression, northern Gulf of California: palaeoearthquakes and lithospheric response, paper presented at the American Geophysical Union Fall Meeting.


Rivero, C., J. H. Shaw, and K. Mueller (2000), Oceanic Province, Baja California, Mexico, to develop-ment of marine terraces within the Agua Blanca series ages, faunal correlations and tectonic defor-mation in the Dead Sea and Salton Trough-Gulf of Calif.


