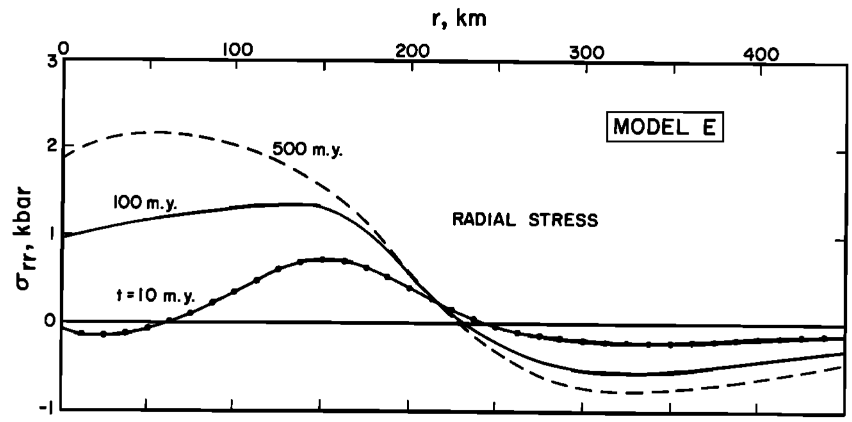

Fig. 18. Surface radial stress σ_{rr} for model E, a combination of uplift heating from model A (Figure 10) and impact heating from model D ($E_B = 7 \times 10^{32}$ erg and $s = 50$ km, Figure 16). An elastic blocking temperature T_e of 800°C has been assumed.

cess. This effect is reflected in the shape of the σ_{rr} distribution at 10 m.y. after basin formation. The zone of contraction contributing most to thermal stress during this time interval occurs not beneath the basin center but near $r \sim 150$ km, where temperatures are below 800°C. The cooled, elastic surface layer is pulled toward this annular region, thus producing zones of mild extension near $r = 0$ and $r = 300$ km. By 100 m.y., most of the crust beneath the basin has cooled below T_e and, as a result, has begun to contribute to and accumulate thermal stress. By this time, the distribution of σ_{rr} begins to resemble models without a blocking temperature. The magnitude of σ_{rr}, however, is everywhere less than in previous models. This is especially evident near the basin center, where σ_{rr} in model D (Figure 17) exceeds 30 kbar at 500 m.y. while σ_{rr} in model E is an order of magnitude less. The position of the surface zone experiencing maximum radial extension is strongly controlled by both the value of T_e and the radial extent of isotherm uplift beneath the basin. The result is a region of extensional stress that is broader, smaller in the magnitude of stress, and located at a greater radial distance from the basin center compared to the same model without a blocking temperature. In thermal stress models with lower adopted values for T_e, these effects are more pronounced. After 100 m.y. in model E, σ_{rr} exceeds the extensional strength of igneous rock only for $r > 250$ km. Even though model D provided a good fit to the topography and the location of fissuring within Orientale among models not including the effects of a blocking temperature, model E demonstrates that with the inclusion of a blocking temperature a value of s less than 50 km is necessary to match the location of fissuring in Orientale. If a blocking temperature of 800°C is appropriate, thermal stress calculations for models otherwise similar to E indicate that s should be about 20 km to predict fissuring at the distance range observed.

DISCUSSION

The models presented above suggest that the emplacement of heat during the formation of an impact basin and the subsequent loss of that heat were important contributors to the topography and tectonics of lunar impact basins. Beyond this qualitative result, we may use the results of these models to place approximate constraints on the quantity and distribution of impact heat implanted during the formation of the Orientale basin. These estimates are based on the assumption that the observed fissuring is a product of thermal stress [*Church et al.*, 1982] and that the distribution of anomalous temperatures resulting from isotherm uplift is relatively well known. Given these assumptions, we note first that isotherm uplift alone predicts poorly the location of fissuring within Orientale. It follows that heat converted from impact kinetic energy must have been at least as important to the early thermal budget of the basin. Expressed differently, E_B is probably comparable to or greater than 10^{32} erg, the total amount of anomalous heat contributed by isotherm uplift.

The distribution of impact heating has been assumed in this paper to follow an exponential decay with distance characterized by a fixed decay constant s (equation (18)). When no blocking temperature is considered, s must be about 50 km (model D, Figures 16–17) to predict correctly the occurrence of fissuring in the distance range 150 to 230 km. With the inclusion of a blocking temperature (model E, Figure 18), an even greater concentration of impact heat near the point of impact is required to match the fissure positions. We therefore suggest that the decay of impact heat density with distance from the point of impact for an Orientale-size event must be rapid and that for the exponential parameterization assumed in this paper s must be less than or equal to 50 km. For comparison, the energy density distribution shown in Figure 5 of *O'Keefe and Ahrens* [1975] for their numerical model of the formation of the Imbrium basin falls off approximately exponentially with distance with a decay constant of about 20 km. As a measure of the parameterization used here, with $s = 50$ km about half the buried impact heat lies inward of $r = 100$ km and 90% of the heat lies inward of $r = 200$ km.

If 5 km can be regarded as an upper bound on the thermal subsidence that has occurred within the central region of the Orientale basin, then the calculations of this paper also permit an estimate of an upper bound on E_B. A superposition of model A (isotherm uplift) and model D ($E_B = 7 \times 10^{32}$ erg and $s = 50$ km) accounts for the entire relief of the central depression. Further, if s is less than 50 km, as suggested by the thermal stress models that incorporate an elastic blocking temperature, subsidence at the center of the basin increases for a given value of E_B (compare models B and C, Figures 14 and 15). On these grounds 7×10^{32} erg is an upper bound on E_B.

It should be recalled that these bounds on E_B have been estimated without regard to sources of stress other than thermal stress. The state of stress immediately following basin formation is unknown. Residual stresses may have remained after shock release and cavity collapse, but such stresses