NPTEL ONLINE CERTIFICATION COURSES

EARTHQUAKE SEISMOLOGY

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Module 05 : Refraction and Reflection seismology
Lecture 02: Refraction seismology in dipping interfaces and crustal structure

CONCEPTS COVERED

- Recap
- Refracted waves for the dipping layers case
- Crustal Structure infer from the refraction seismology
- Summary

Recap

- A major application of seismology is the determination of the distribution of seismic velocities, and hence elastic properties, within the earth.
- Refraction seismology is concerned about the travel time of critically refracted wave at the interface of two medium, to identify the depth of interface, velocity of the layer and underlying half space.
- Travel time as a function of offset $T_D = x/v_0$

$$T_R^2(x) = \frac{x^2}{v_0^2} + 4\frac{h_0^2}{v_0^2}$$

$$T_H(x) = \frac{x}{v_1} + 2h_o\left(\frac{1}{v_o^2} - \frac{1}{v_1^2}\right)^{1/2} = \frac{x}{v_1} + \tau_1$$

- The thickness of successive layers can be found by starting with the top layer:

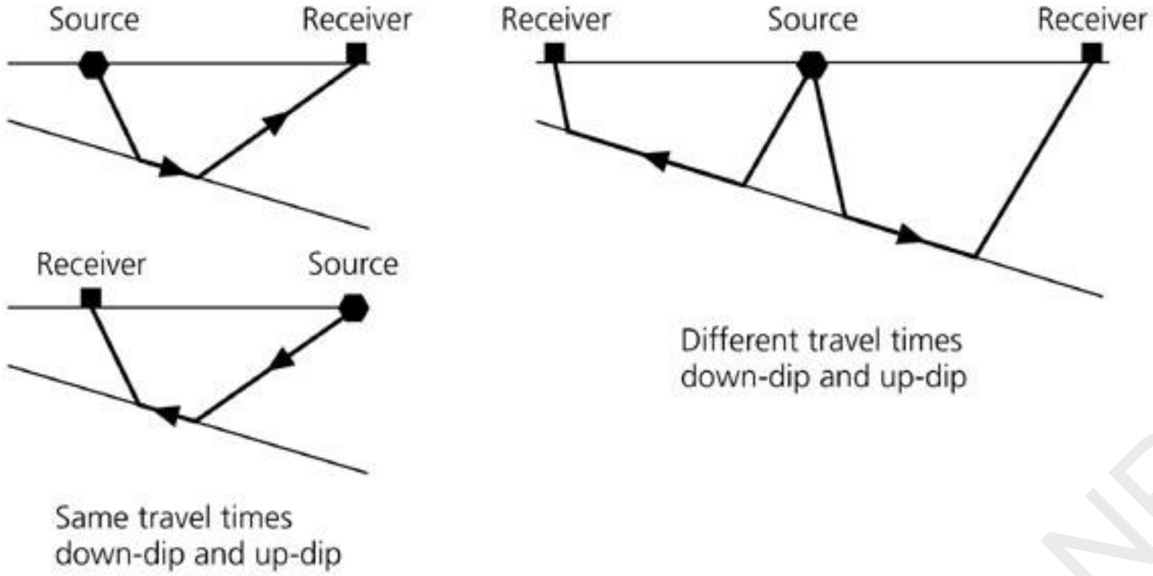
$$h_{n-1} = \frac{\tau_n - 2\sum_{j=0}^{n-2} h_i\left(\frac{1}{v_j^2} - \frac{1}{v_n^2}\right)^{1/2}}{2\left(\frac{1}{v_{n-1}^2} - \frac{1}{v_n^2}\right)^{1/2}}$$

Refraction Seismology

For dipping layers

- **The refraction method can also be applied if the interfaces between layers are not horizontal.**
- **Conducting a reversed profile yields the travel times for ray paths in both the down-dip and the up-dip directions.**
- **Since we do not know the orientation of the interface, i.e. updip and downdip directions, so we arbitrarily consider one as forward and other as reverse direction.**
- **When a refractor dips, the slope of travel-time curve does not represent the "true" layer velocity.**
- **Shooting updip, i.e., geophones are on updip side of shot, apparent refractor velocity is higher.**

Figure 3.2-12: Difference between up-dip and down-dip paths.



Left: If the source and the receiver are interchanged on a reversed refraction profile, the travel time is unchanged. (Principle of reciprocity)

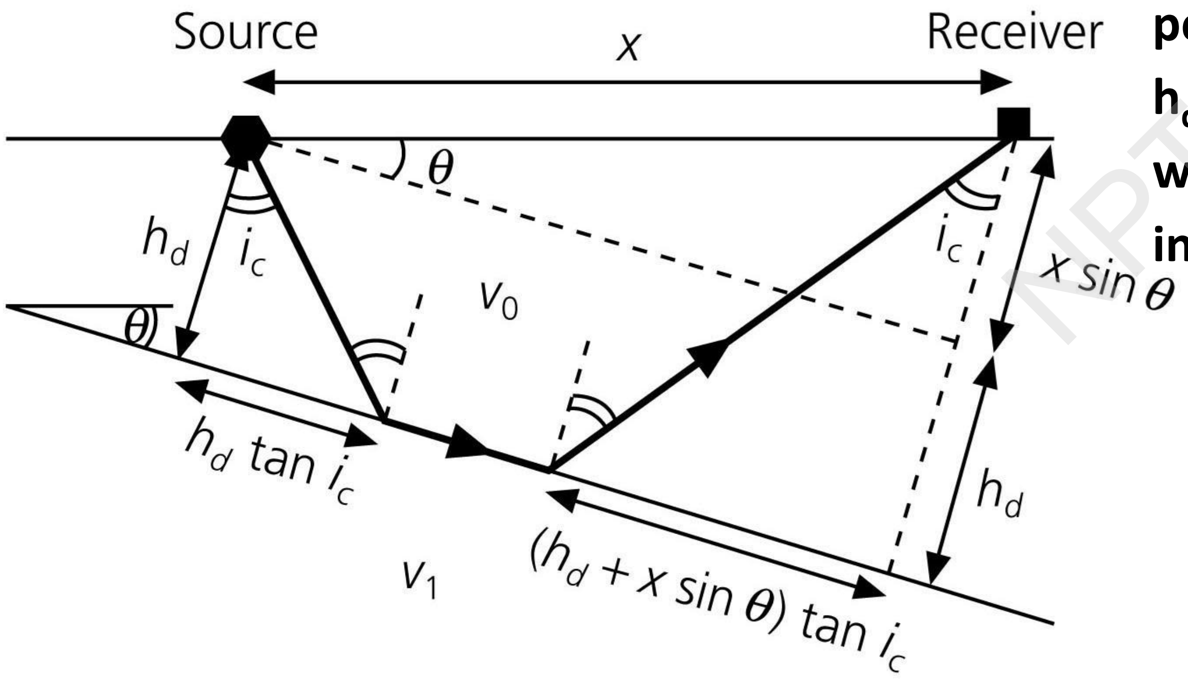
Right: Different up-dip and down-dip travel times occur because, for a given source position, waves going the same distance along the surface in opposite directions sample the dipping interface differently.

Refraction Seismology

For dipping layers

In this geometry, the depths to the interface below the source and the receiver differ due to the dip angle, θ

Figure 3.2-10: Ray paths for a dipping layer over a halfspace.



Consider, the down-dip ray path (Fig. 3.2-10) from a source, below which the perpendicular distance to the interface is h_d, to a receiver at a distance x, below which the perpendicular distance to the interface is ($h_d + x \sin \theta$).

$$\underbrace{T_d(x)}_{\text{Travel time of head wave in downdip direction}} = \underbrace{\frac{x\cos\theta - (2h_d + x\sin\theta)\tan i_c}{v_1}}_{\text{Travel time along the interface}} + \underbrace{\frac{2h_d + x\sin\theta}{v_o\cos i_c}}_{\text{Travel time of upgoing and downgoing wave}}$$

$$\begin{aligned} T_d(x) &= \frac{x\cos\theta\sin i_c}{v_o} + \frac{(2h_d + x\sin\theta)\left(1-\sin^2 i_c\right)}{v_o\cos i_c} \\ &= \frac{x\sin(i_c+\theta)}{v_o} + \frac{2h_d\cos i_c}{v_o} = \frac{x}{v_d} + \tau_d \end{aligned}$$

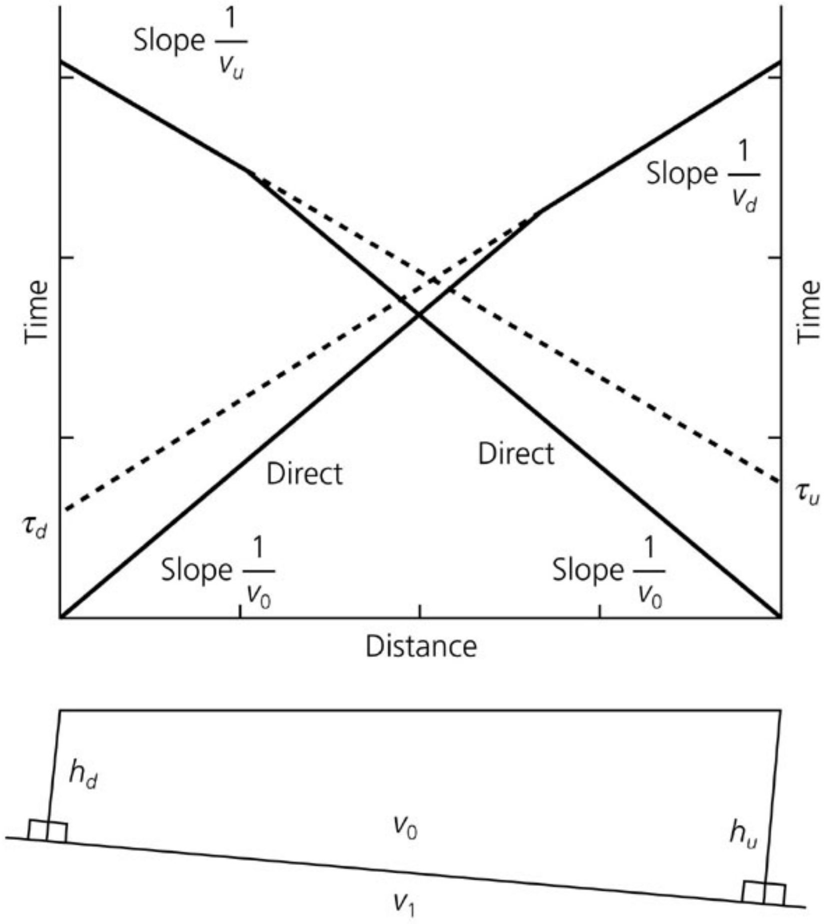
which is a equation of straight line with slope $1/v_d$ and intercept τ_d

Similarly, travel time of head wave in updip direction

$$T_u(x) = \frac{x\sin(i_c-\theta)}{v_o} + \frac{2h_u\cos i_c}{v_o} = \frac{x}{v_u} + \tau_u$$

The apparent velocity in updip and downdip direction are given as $v_u = \dfrac{v_o}{\sin(i_c - \theta)}$ $v_d = \dfrac{v_o}{\sin(i_c + \theta)}$

Figure 3.2-11: Travel time plot for up-dip and down-dip profiles.



The apparent velocity in the up-dip direction is greater than the half space velocity, and that in the down-dip direction is smaller. Time axis intercept $\tau_u = \dfrac{2h_u \cos i_c}{v_o}$, $\tau_d = \dfrac{2h_d \cos i_c}{v_o}$

also differ. The direct wave travel time is the same in both directions, so the crossover distances differ.

Travel time plot for a reversed profile and its interpretation. The up-dip and down-dip slopes and intercepts differ.

The slopes of the direct and head wave travel times yield the dip angle

$$\theta = \frac{1}{2}\left(\sin^{-1}\frac{v_o}{v_d} - \sin^{-1}\frac{v_o}{v_u}\right)$$

$$i_c = \frac{1}{2}\left(\sin^{-1}\frac{v_o}{v_d} + \sin^{-1}\frac{v_o}{v_u}\right)$$

The halfspace velocity v_1 is found from the critical angle and v_0,and the intercept times then yield the layer thickness.

By reciprocity, the two experiments give the same travel time. Thus, for a ray path connecting
two points, it does not matter whether the wave travels up-dip or down-dip.

- By contrast, for two receivers at the same distance from a source, one up-dip and one downdip, the travel times differ because the ray paths encounter the dipping interface at different depths.

- The travel times for two sources at the same distance from a receiver, one up-dip and one down-dip.

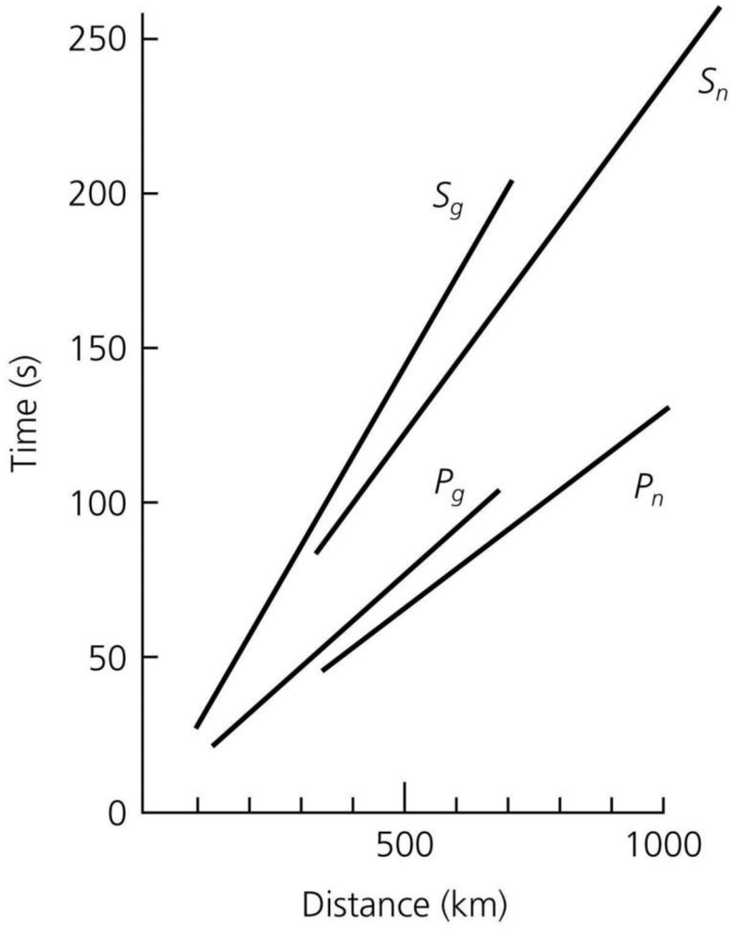
- If the dip were zero, then the travel times would be the same for all these cases because all ray paths encounter the interface at the same depth.

- For a flat geometry the travel time depends only on the distance between the source and the receiver.

- The dip found from a reversed profile is not a true dip if the profile is not perpendicular to the strike of the layer. Instead, the measured dip is an apparent dip along the profile.

Crustal Structure from Refraction Seismology

igure 3.2-5: Example of seismograms from a refraction profile.



- Seismic refraction data led A. Mohorovicic in 1909, observing two P arrivals, he identified the first as having traveled in a deep high-velocity (7.7 km/s) layer, and the second as a direct wave in a slower (5.6 km/s) shallow layer about 50 km thick.
- These layers, are known as the crust and the mantle. The boundary between them is known as the Mohorovicic discontinuity, or Moho.
- The arrival times are head wave as P_n and the direct wave as P_g ("g" for "granitic"). Similarly, for S-waves it is S_n and S_g.

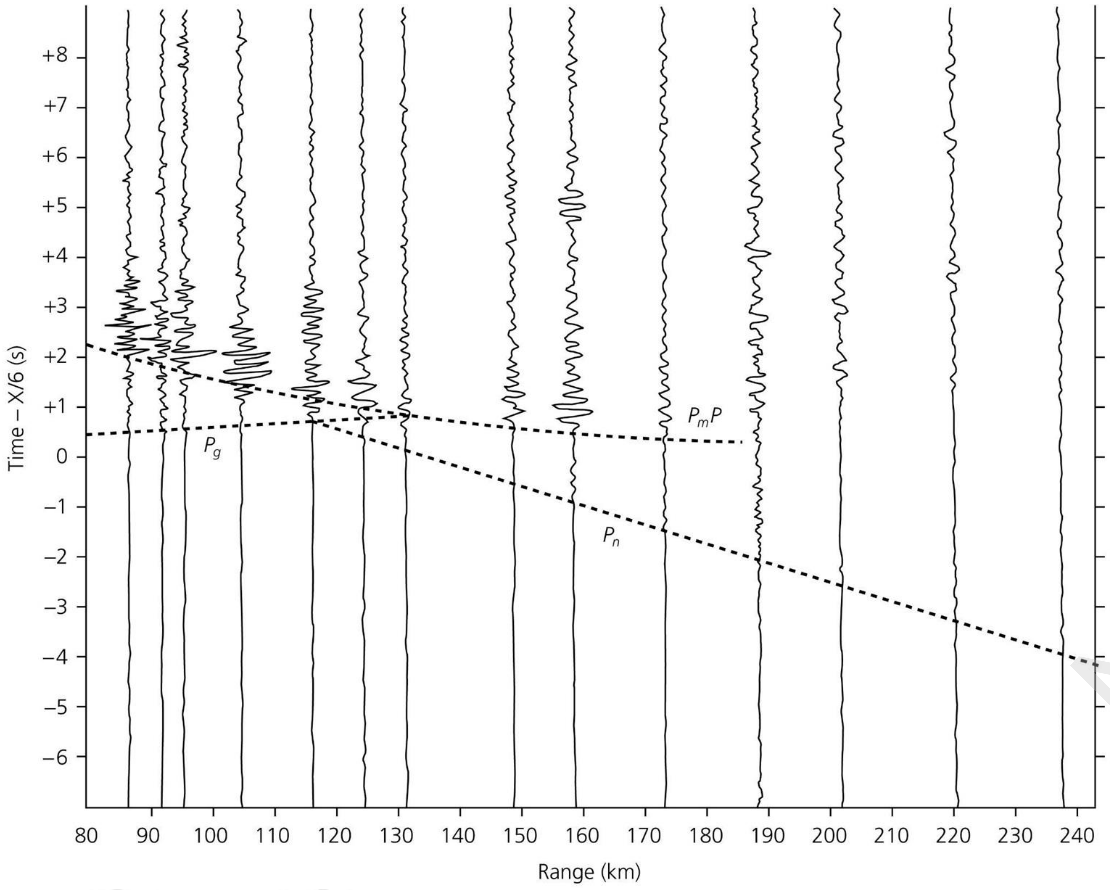
What are P_n velocities?

- P_n is a seismic phase that travels at the crust-mantle interface. The apparent velocity of P_n gives an average of the mantle velocity along the path that it travelled.

- P_n is the first arrival at the distances b/w 2^0 and 16^0.

- slow P_n – warm upper mantle.
- fast P_n – cool upper mantle.

Figure 3.2-5: Example of seismograms from a refraction profile.



This is a reduced travel time plot.

Record section: Seismograms plotted by location of seismometer from the source.

Moveout: change in phase location as a function of distance.

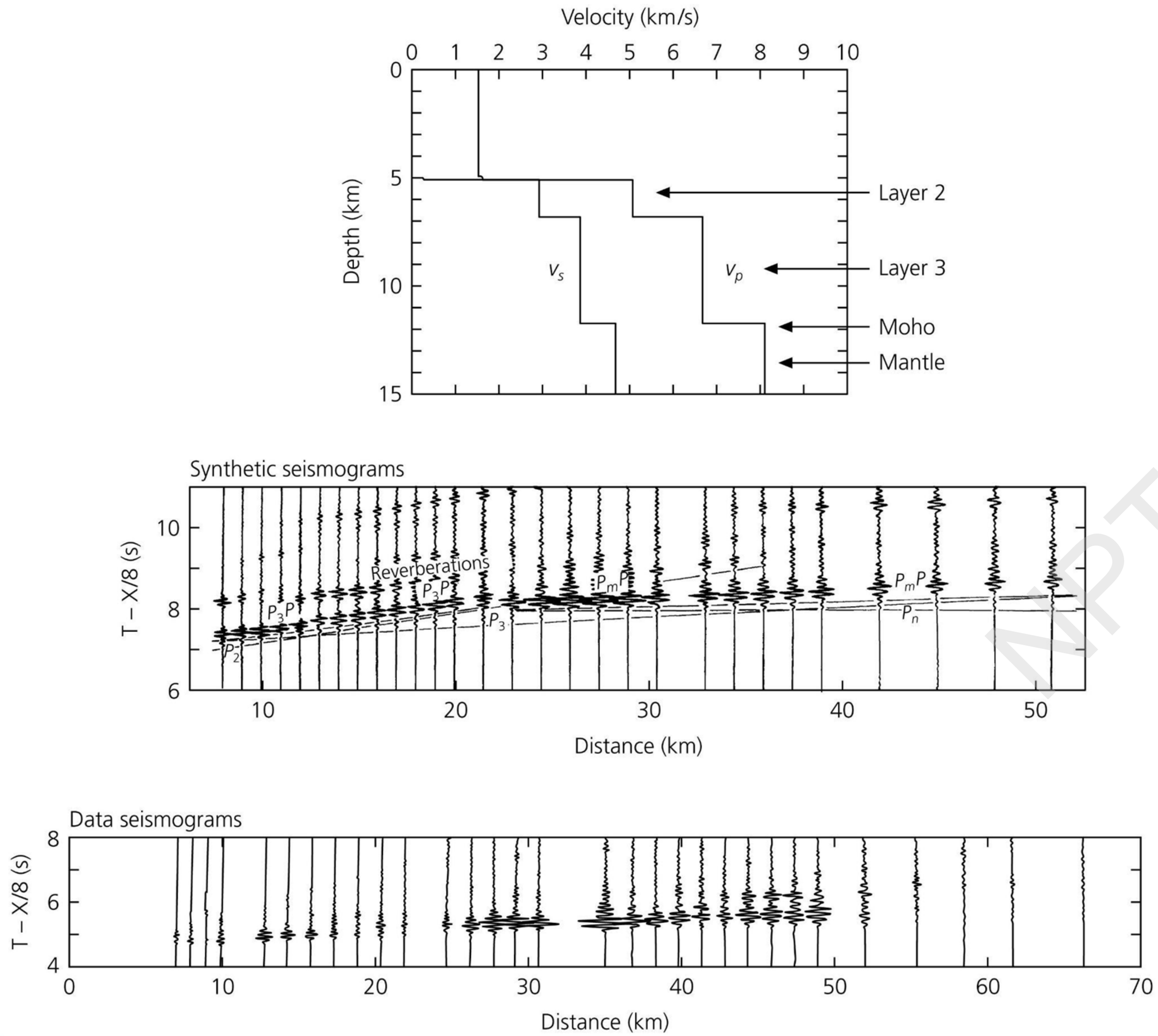
Reduced velocity: Time axis is changed to a horizontal line corresponds to a specific velocity (in this case 6 km/s).

Pg - Phases with moveout that have positive slopes, are seeing Vp<6 km/s because they are travelling in crust.

Pn - Phases with moveout having negative slopes are seeing Vp> 6 km/s are travelling in mantle.

So, reduced travel time plot provides an easy way to distinguish between velocities.

Figure 3.2-15: Modeling ocean crustal layers with synthetic seismograms.



Top: Oceanic crust model with sharp transitions between layer 1 (water), layer 2 (unconsolidated sediment), layer 3 (crustal rock), and the mantle. Center

Synthetic seismograms for this model. P_2 , P_3 , and P_n are head waves from layers 2, 3, and the mantle. P_3P and P_mP are reflections off the tops of layer 3 and the mantle.

Data showing an absence of the large P_3P arrivals predicted by the layered model.

- Velocity structures are often interpreted in terms of composition.
- To do this, seismological results are combined with other geophysical data (e.g., gravity), geological fieldwork, and laboratory studies of the seismic velocities of rocks.
- Interpreting seismological results for the crust and mantle in terms of composition requires knowing something about rocks and the minerals that compose them.
- Physical properties of rocks, such as density and seismic velocity, depend on their mineral composition.

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THANK YOU !